Magnetic fabric, palaeomagnetic and structural investigation of the accretion of lower oceanic crust using ophiolitic analogues

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This copy of the thesis has been supplied on condition that anyone who consults it is understood to recognise that its copyright rests with its author and that no quotation from the thesis and no information derived from it may be published without the author's prior consent.
Pope Julius II: “When will you make an end?”
Michelangelo: “When I am finished!”
— The Agony and the Ecstasy (1965)

“It’s snowing still,” said Eeyore gloomily.”
“So it is.”
“And freezing.”
“Is it?”
“Yes,” said Eeyore. “However,” he said, brightening up a little, “we haven’t had an earthquake lately.”
— Winnie-the-Pooh: A.A. Milne (1928)
Abstract

This thesis presents the results of a combined magnetic fabric and palaeomagnetic analysis of lower crustal rocks exposed in the Oman (Semai) ophiolite. This has long been an important natural laboratory for understanding the construction of oceanic crust at fast spreading axes and its subsequent tectonic evolution, but magnetic investigations in the ophiolite have been limited. Analyses presented here involve using: (i) magnetic anisotropies as a proxy for magmatic petrofabrics in lower crustal rocks in order to contribute to outstanding questions regarding the mode of accretion of fast-spread oceanic crust; and (ii) classical palaeomagnetic analyses to determine the nature of magnetization in these rocks and gain further insights into the regional-scale pattern of tectonic rotations that have affected the ophiolite.

The extensive layered gabbro sequences exposed in the Semai ophiolite have been sampled at a number of key localities. These are shown to have AMS fabrics that are layer-parallel but also have a regional-scale consistency of the orientation of maximum anisotropy axes. This consistency across sites separated by up to 100 km indicates large-scale controls on fabric development and may be due to consistent magmatic flow associated with the spreading system or the influence of plate-scale motions on deformation of crystal mushes emplaced in the lower crust. Detailed analysis of fabrics in a single layer and across the sampled sections are consistent with either magmatic flow during emplacement of a melt layer into a lower crustal sill complex, or traction/drag of such layers in response to regional-scale stresses (e.g. mantle drag). Together, results support formation of the layered gabbros by injection of melt into sill complexes in the lower crust. New anisotropy data from the overlying foliated gabbros sampled at two key localities also provide insights into the style of melt migration at this crustal level. Fabrics are consistent with either focused or anastomosing magmatic upwards flow through this layer, reflecting melt migration beneath a fossil axial melt lens.

Previous palaeomagnetic research in lavas of the northern ophiolitic blocks has demonstrated substantial clockwise intraoceanic tectonic rotations. Palaeomagnetic data from lower crustal sequences in the southern blocks, however, have been more equivocal due to complications arising from remagnetization. Systematic sampling resolves for the first time a pattern of remagnetized lowermost gabbros and retention of earlier magnetizations by uppermost gabbros and the overlying dyke-rooting zone. Results are supported by a positive fold test that shows that remagnetization of lower gabbros occurred prior to Campanian structural disruption of the Moho. NW-directed remagnetized remanences in the lower units are consistent with those used previously to infer lack of significant rotation of the southern blocks. In contrast, E/ENE-directed remanences in the uppermost gabbros imply a large, clockwise rotation of the southern blocks, of a sense and magnitude consistent with that inferred from extrusive sections in the northern blocks. Hence, without the control provided by systematic crustal sampling, the potential for different remanence directions being acquired at different times may lead to erroneous tectonic interpretation.
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Figure 3.1: Distribution of the Tethyan ophiolites spanning the length of the Alpine-Himalayan mountain belt and tracing the suture of the closed Neotethys Ocean. The Troodos (#3) and Semail (#5) ophiolites lie at the western and eastern end, respectively, of the “Pan-Arabian Ophiolite Crescent”. A progression from MOR-type to SSZ-type ophiolites in also observed from the Jurassic ophiolites in the west to the Cretaceous ophiolites in the east (from Dilek & Furnes, 2009).

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Figure 3.31: Schematic geological map of the Wadi Nassif lower layered gabbro crustal section with location of sampling sites (blue stars). Inset map shows location of Wadi Nassif within the Samad-Wadi Tayin block in the southern part of the Semail ophiolite. Inset stereonet presents representative structural data from Wadi Nassif.

Figure 4.1: AF demagnetization curves for a range of magnetite grain sizes of 0.1 mT TRM. The changing shape of the demagnetization curves reflects an increase in the grain size of magnetite from SD (sigmoidal curve) to MD (exponential curve) (from Dunlop and Özdemir, 1997, after Argyle et al., 1994).

Figure 4.2: Normalised magnetization intensity curves against field strength during AF demagnetization with calculated median destructive field (MDF) values for layered gabbros from: (a) Wadi Abyad; (b) Wadi Khafifah; (c) Somrah; (d) Wadi Nassif.

Figure 4.3: Normalised magnetization intensity curves against field strength during AF demagnetization with calculated median destructive field (MDF) values for foliated gabbros and dykes sampled in: (a) Wadi Abyad; (b) Wadi Khafifah.

Figure 4.4: Normalised magnetization intensity curves against field strength during AF demagnetization with calculated median destructive field (MDF) values for varitextured gabbros and dykes sampled in: (a) Wadi Abyad; (b) Wadi Khafifah; (c) Somrah.
Figure 4.5: Normalised magnetization intensity curves against temperature showing characteristic responses to thermal demagnetization for layered gabbros from: (a) Wadi Abyad; (b) Wadi Khafifah; (c) Somrah; (d) Wadi Nassif.

Figure 4.6: Normalised magnetization intensity curves against temperature showing characteristic responses to thermal demagnetization for foliated gabbros from: (a) Wadi Abyad; (b) Wadi Khafifah.

Figure 4.7: Normalised magnetization intensity curves against temperature showing characteristic responses to thermal demagnetization for varitextured gabbros and dykes sampled in: (a) Wadi Abyad; (b) Wadi Khafifah; (c) Somrah.

Figure 4.8: Representative normalised IRM acquisition and back-field IRM acquisition curves for layered gabbros from: (a) Wadi Abyad; (b) Wadi Khafifah; (c) Somrah; (d) Wadi Nassif.

Figure 4.9: Representative normalised IRM acquisition and back-field IRM acquisition curves for foliated gabbros from: (a) Wadi Abyad; (b) Wadi Khafifah.

Figure 4.10: Representative normalised IRM acquisition and back-field IRM acquisition curves for varitextured gabbros and dykes sampled in: (a) Wadi Abyad; (b) Wadi Khafifah; (c) Somrah.

Figure 4.11: Characteristic normalised thermomagnetic curves for layered gabbros from: (a) Wadi Abyad; (b) Wadi Khafifah; (c) Somrah; (d) Wadi Nassif. Solid lines represents heating curves up to 700°C, whilst dashed lines represents the cooling curves from 700°C back to ambient temperature.

Figure 4.12: Characteristic normalised thermomagnetic curves for foliated gabbros from: (a) Wadi Abyad; (b) Wadi Khafifah.

Figure 4.13: Characteristic normalised thermomagnetic curves for varitextured gabbros and dykes sampled in: (a) Wadi Abyad; (b) Wadi Khafifah; (c) Somrah.

Figure 4.14: Temperature dependence of inverse magnetic susceptibility for a selection of thermomagnetic curves from sites sampled within: (a) the layered gabbros (LG); (b) the foliated gabbros (FG); (c) the varitextured gabbros (VG); (d) the discrete dykes (D). The Curie temperature (Tc) has been calculated using the inverse susceptibility method of Petrovsky & Kapička (2006).

Figure 5.1: Variation of low field magnetic susceptibility with concentration of various common rock forming minerals (from Tarling & Hrouda, 1993).

Figure 5.2: Histograms of low field magnetic susceptibility for each pseudostratigraphic level sampled in lower crustal rocks of the Semail ophiolite.

Figure 5.3: Summary of AMS parameters for layered and foliated gabbros sampled in this study. Left: variation of corrected anisotropy degree with mean susceptibility; Right: Jelinek plot of corrected anisotropy degree against shape parameter.

Figure 5.4: AMS data from layered gabbros sampled in the Semail ophiolite, showing orientations of specimen AMS principal axes, with associated site mean axes (larger symbols) and Jelinek (1978) confidence ellipses. WA = Wadi Abyad, KF = Wadi Khafifah area, SR = Somrah and WN = Wadi Nassif. Black great circles = orientation of macroscopic layering observed in the field; red diamonds = orientation of lineations defined by preferred alignments of crystal long axes observed in the field; red great circles = orientation of fault planes observed at site WA04A+B.
Figure 5.5: AMS (left) and ApARM (right) data from layered gabbros at site KF08 at Khaffifah. Note the excellent correlation between these two forms of anisotropy, demonstrating that AMS data are not affected by single-domain-related inverse fabrics.

Figure 5.6: Photomicrographs of thin sections of layered gabbros (lefthand side = plane polarised light; righthand side = under crossed polars), showing: (a, b) magmatic textures defined by interlocking, undeformed crystals of plagioclase, clinopyroxene and olivine; (c, d) the typical nature of serpentinization of olivine crystals and presence of alteration halos; and (e, f) equant opaque minerals, assumed to be magnetite, distributed along cleavage planes in clinopyroxene crystals. Ol = olivine, Cpx = clinopyroxene, Pl = plagioclase feldspar, Mag = magnetite. Mineral symbols in accordance with Kretz (1983).

Figure 5.7: Results of point-counting analysis of thin sections from Somrah Layer A (400 counts), showing the variation of the three main igneous phases (clinopyroxene, plagioclase and olivine) as well as the percentages of magnetite and alteration through a typical gabbro layer.

Figure 5.8: Photomicrographs of thin sections of layered gabbros from Somrah (lefthand side = plane polarised light; righthand side = under crossed polars), illustrating changes in mineralogy through a single layer. See text for details. Ol = olivine, Cpx = clinopyroxene, Pl = plagioclase feldspar.

Figure 5.9: Graphs showing the variation of mean susceptibility and AMS parameters through Layer A at Somrah, with subdivisions (Zones Aii-iii) based on analysis of thin section point count data.

Figure 5.10: Graph of mean susceptibility against corrected anisotropy degree for samples from Layer A at Somrah. A strong correlation in Zone Aiii (best-fitting power law relationship indicated by red dashed line and regression equation) suggests a mineralogical control on anisotropy in the top of the layer, controlled by co-variations in clinopyroxene and plagioclase modal proportions (see text).

Figure 5.11: AMS data from foliated gabbros sampled in the Semail ophiolite, showing orientations of specimen AMS principal axes, with associated site mean axes (larger symbols) and Jelinek (1978) confidence ellipses. Note that stereonets for sites WA18-37 and KF12-27 show the combined results of detailed sampling across transects running perpendicular to the strike of the macroscopic foliation. WA = Wadi Abyad, KF = Wadi Khaffifah area, SR = Somrah and WN = Wadi Nassif. Black great circles = orientation of macroscopic foliation observed in the field; red diamonds = orientation of lineations defined by preferred alignments of crystal long axes observed in the field.

Figure 5.12: Simplified geological maps showing the location of the transects sampled within foliated gabbros (immediately beneath the varitextured gabbros/fossil axial magma chamber horizon) at Wadi Abyad and Khaffifah.

Figure 5.13: Map showing the location of sampling sites along a transect across the foliated gabbros in Wadi Abyad. Foliation orientation = 72/254 (measured at site WA21, but consistent across the sampled area). Green sites = coarse-grained facies; red sites = fine-grained facies. Grid lines every 10 m.

Figure 5.14: Map showing the location of sampling sites along a transect across the foliated gabbros in Wadi Khaffifah. Foliation orientation = 35/277 (measured at site KF25, but consistent across the sampled area). Green sites = coarser-grained olivine-rich gabbro; red sites = equigranular fine-grained gabbro. Grid lines every 10 m.

Figure 5.15: Field photographs of foliated gabbro facies sampled along: (a) the Wadi Abyad and; (b) the Khaffifah transects. KF20 = coarser-grained, olivine-rich gabbro; KF21 = equigranular, fine-grained gabbro.
Figure 5.16: AMS data from foliated gabbros across the transect in Wadi Abyad, showing orientations of specimen AMS principal axes. Dashed great circles = orientation of macroscopic foliation observed in the field. See Figure 5.13 for base map showing relative locations of sites.

Figure 5.17: AMS data from foliated gabbros across the transect in Wadi Khafifah, showing orientations of specimen AMS principal axes. Dashed great circles = orientation of macroscopic foliation observed in the field. See Figure 5.14 for base map showing relative locations of sites.

Figure 5.18: Equal area stereographic projections showing principal axes of ApARM ellipsoids of samples collected across transects in the foliated gabbros at (a) Wadi Abyad and (b) Khafifah.

Figure 5.19: Photomicrograph of swathes across four thin-sections from the Wadi Abyad transect, with $K_{max}$ aligned along the length of the images.

Figure 5.20: Photomicrograph of swathes across four thin-sections from the Khafifah transect, with $K_{max}$ aligned along the length of the images.

Figure 5.21: Photomicrograph of thin section from sample KF2303B at Wadi Khafifah, showing details of magmatic, anastomosing flow fabrics in foliated gabbros.

Figure 5.22: Obliquity of grain long axes relative to the orientation of $K_{max}$ axes from: (left) the Wadi Abyad transect; and (right) the Wadi Khafifah transect.

Figure 5.23: Summary of AMS results from the varitextured gabbros, associated dykes and the dyke-rooting zone in Wadi Abyad.

Figure 5.24: Equal area stereographic projections showing (a) AMS maximum and (b) minimum principal axes from all layered gabbros sites sampled in this study (Wadis Abyad, Khafifah, Somrah and Nassif). Tilt corrected data produced by restoring observed magmatic layering at each site to the horizontal using a standard tilt correction. Primitives on right hand projections therefore correspond to the plane of magmatic layering. Note the consistent ENE-WSW orientation of $K_{max}$ axes across all sampling localities, suggesting regional-scale controls on AMS fabrics in the layered gabbros.

Figure 5.25: Variation in the inclination (dip) of $K_{max}$ axes through Layer A of the layered gabbros at Somrah. Note that mean inclination is shown where more than one sample was collected at the same height.

Figure 5.26: Equal area stereographic projections showing the orientation of $K_{max}$ axes in samples from Layer A at Somrah, relative to magmatic layering and magmatic lineation. (left) In situ coordinates, showing distribution of $K_{max}$ axes centred on the magmatic lineation and straddling the plane of magmatic (modal) layering; (right) same data (with samples from the middle of the layer omitted for clarity) after rotating the plane of layering and the magmatic lineation to vertical, i.e. looking down the lineation in the direction of the $K_{max}$ axes. Note systematic difference between $K_{max}$ axes from the lower and upper parts of the layer, suggesting obliquity of crystal orientations relative to the layer margins (especially in near the base of the layer).

Figure 5.27: Schematic illustration of the theoretical arrangement of magnetic fabric axes that may result from emplacement of a melt layer. The obliquity of fabrics relative to layer margins as well as the trend of the $K_{max}$ axes may potentially be used to infer emplacement direction (Morris, unpublished diagram).

Figure 5.28: Alternative models for producing oblique fabrics at the margins of a layered gabbro unit. See text for details. Green/yellow lines schematically illustrate olivine/clinopyroxene & plagioclase crystal orientations, respectively.
Figure 5.29: The variation with depth through the lower crust of several key quantifiable parameters as predicted by the gabbro glacier (orange line) and multiple-sills (green dashed line) models (modified from a figure designed and drafted by Roz Coggan).

Figure 5.30: The relationship between strain and AMS during experimental deformation of a magnetite–bearing synthetic sandstone (Borradaile and Alford, 1987).

Figure 5.31: Vertical variations in corrected anisotropy degree at a range of scales.

Figure 5.32: Combined AMS data from all samples along the (a) Wadi Abyad and (b) Wadi Khafifah foliated gabbro transects in geographic coordinates.

Figure 5.33: Combined AMS data from all samples along the (a) Wadi Abyad and (b) Wadi Khafifah foliated gabbro transects following correction of the local Moho orientation to the horizontal (tilt corrections: Wadi Abyad = 32/032, Wadi Khafifah = 29/173).

Figure 5.34: Variation in inferred melt transportation styles through the foliated gabbros sections beneath the melt lenses in (a) Wadi Abyad and (b) Wadi Khafifah.

Figure 5.35: Schematic detailing the processes inferred to be taking place beneath the melt lens at a fast-spreading ridge and key predictions of the melt migration model of MacLeod & Yaouancq (2000): (a1) As partially crystallized mush moves ‘off-axis’ further cooling results in crystals becoming locked into the sub-vertical flow direction as highlighted by vertical l-s crystal shape fabrics; (a2) After crystallization of the lower and upper gabbros, residual (interstitial) melt rises through the crystalline mush towards the high-level melt at the base of the sheeted dyke complex by intergranular (i.e. grain-supported) porous flow; (a3) No progressive steepening of the gabbro fabric at the varitextured-foliated gabbro transition is observed, implying that the little or no crystal subsidence from the melt lens down into the underlying gabbros is taking place; (a4) Residual melt after in situ crystallization of the layered and foliated gabbros accumulates within the high-level melt lens giving the varitextured gabbros their evolved geochemistry similar in composition to the overlying upper crust; (b) Anticipated magmatic architecture of the lower oceanic crust as suggested by the melt migration model (from MacLeod & Yaouancq, 2000).

Figure 6.1: Histograms of NRM intensities for each pseudostratigraphic level sampled in the Semail ophiolite.

Figure 6.2: Characteristic Zijderveld plots of thermal (Th) and alternating field (AF) demagnetization data for specimens from the Wadi Abyad crustal section. LG = layered gabbros; FG = foliated gabbros; VG = varitextured gabbros; D = dykes; DRZ = dyke-rooting zone.

Figure 6.3: Characteristic Zijderveld plots of thermal (Th) and alternating field (AF) demagnetization data for specimens from the Wadi Khafifah crustal section. LG = layered gabbros; FG = foliated gabbros; VG = varitextured gabbros; D = dykes.

Figure 6.4: Characteristic Zijderveld plots of thermal (Th) and alternating field (AF) demagnetization data for specimens from the Somrah layered gabbros and discrete dykes. LG = layered gabbros; D = dykes.

Figure 6.5: Characteristic Zijderveld plots of thermal (Th) and alternating field (AF) demagnetization data for specimens from Wadi Nassif. LG = layered gabbros.

Figure 6.6: Equal area stereographic projections in geographic (left) and tilt corrected (right) coordinates showing the distribution of site mean magnetization directions from Wadi Abyad. Note the large declination anomaly between the layered (blue) and foliated (red) gabbros compared to the overlying varitextured gabbros (orange) (including discrete dykes) and dyke-rooting zone (dark purple). Dykes in the layered
gabbros shown in green. The large light purple symbol on each plot denotes the locality mean magnetization direction. Solid/open symbols = lower/upper hemisphere projections; ellipses = projection of $\alpha_{95}$ cones of confidence around site means.

Figure 6.7: Equal area stereographic projections in geographic (left) and tilt corrected (right) coordinates showing the distribution of site mean magnetization directions from Wadi Khafifah. Layered gabbros = blue symbols; foliated gabbros = red; dykes = green. The large light purple symbol on each plot denotes the locality mean magnetization direction. Solid/open symbols = lower/upper hemisphere projections; ellipses = projection of $\alpha_{95}$ cones of confidence around site means.

Figure 6.8: Equal area stereographic projections in geographic (left) and tilt corrected (right) coordinates showing the distribution of site mean magnetization directions from Somrah. Layered gabbros = blue symbols; dykes = green. The large light purple symbol on each plot denotes the locality mean magnetization direction. Solid/open symbols = lower/upper hemisphere projections; ellipses = projection of $\alpha_{95}$ cones of confidence around site means.

Figure 6.9: Equal area stereographic projections in geographic (left) and tilt corrected (right) coordinates showing the distribution of site mean magnetization directions from the Wadi Nassif lower layered gabbro crustal section. Layered gabbros = blue symbols. The large light purple symbol on each plot denotes the locality mean magnetization direction. Solid/open symbols = lower/upper hemisphere projections; ellipses = projection of $\alpha_{95}$ cones of confidence around site means.

Figure 6.10: Summary of published palaeomagnetic data (black outlined clock diagrams) from the southern massifs of the Semail ophiolite, combined with results from this study (red outlined clock diagrams) (modified from Weiler, 2000; base map modified from e.g. Lippard et al. 1986; Nicolas & Boudier, 2011). Arrows = mean tilt corrected remanence directions. Sources: L = Luyendyk & Day (1982); S = Shelton (1984); T = Thomas (1988); F = Feinberg et al. (1999); W = Weiler (2000); M = Meyer (this study). Blue stars: WA = Wadi Abyad; SR = Somrah; WN = Wadi Nassif; KF = Wadi Khafifah). Note the consistency between the N/NW-directed magnetizations reported in this study to the published data. Note also the anomalous NE-directed magnetization recorded in the highest-levels of the Wadi Abyad crustal section.

Figure 6.11: Equal area stereographic projections in geographic (left) and tilt corrected (right) coordinates showing the distribution of secondary site mean magnetizations from sites with N/NW-directed ChRM. Note how secondary magnetization components disperse after tilt correction. Wadi Abyad = dark purple symbols; Wadi Khafifah = blue; Somrah = green; Wadi Nassif = yellow. The large light purple symbol on each plot denotes the mean direction of the secondary components combined. Solid/open symbols = lower/upper hemisphere projections; ellipses = projection of $\alpha_{95}$ cones of confidence around site means; red star = present day field direction for the southern part of the Al Hajar Mountains (approximately 360/41).

Figure 6.12: Equal area stereographic projections in geographic (left) and tilt corrected (right) coordinates showing the distribution of secondary and tertiary site mean magnetizations from sites with NE-directed ChRM (i.e. from the upper units of the Wadi Abyad crustal section). Varitextured gabbros = orange symbols; dyke-rooting zone = dark purple; dykes = green. Large light purple symbols denote the mean direction of the secondary and tertiary components in geographic and corrected coordinates. Solid/open symbols = lower/upper hemisphere projections; ellipses = projection of $\alpha_{95}$ cones of confidence around site means.

Figure 6.13: Equal area stereographic projections in geographic (left) and tilt corrected (right) coordinates showing the distribution of secondary (green ellipses) and tertiary (blue ellipses) site mean magnetizations from the upper part of the Wadi Abyad section relative to the site mean ChRM from the whole of Wadi Abyad. ChRM from the upper
Wadi Abyad units have red ellipses. Black symbols = site mean magnetizations from the layered and foliated gabbros. Note how the secondary and tertiary magnetization components of the varitextured gabbros (WA15), intruding dykes (WA16) and the DRZ (WA17) form an arc (along a great circle) between the NE-directed ChRMs and the N/NW-directed remanences in the underling layered and foliated gabbros. Solid/open symbols = lower/upper hemisphere projections; ellipses = projection of $\alpha_{95}$ cones of confidence around site means.

Figure 6.14: Equal area stereographic projection in tilt corrected coordinates showing thermal demagnetization data from representative individual samples. Note how layered (and foliated) gabbros typically have single component magnetizations, whereas varitextured gabbro and dyke-rooting zone samples migrate from N to NE/ENE directions as low temperature overprints are removed. NRM = natural remanent magnetization.

Figure 6.15: Stereographic projections in geographic (left) and tilt corrected (right) coordinates showing the distribution of site mean magnetizations for all sites (plotted by lithology) from all localities. Layered gabbros = blue symbols; foliated gabbros = red; varitextured gabbros = orange; dyke-rooting zone = dark purple; dykes = green. The large light purple symbol on each plot denotes the combined mean magnetization direction for the southern massifs of the Semail ophiolite. Solid/open symbols = lower/upper hemisphere projections; ellipses = projection of $\alpha_{95}$ cones of confidence around site means.

Figure 6.16: Fold test (Tauxe and Watson, 1994) results for (a) the layered and foliated gabbros from Wadi Abyad, Wadi Khafifah, Somrah and Wadi Nassif and (b) the layered and foliated gabbros from Wadi Abyad, Wadi Khafifah and Somrah (excluding Wadi Nassif).

Figure 6.17: Declination, inclination and palaeolatitude calculated from the African apparent polar wander path of Torsvik et al. (2012) for a site at 23.44°N, 57.62°E (Wadi Abyad). Red symbol = palaeolatitude calculated from the tilt corrected site mean direction at site WA17 (dyke-rooting zone in Wadi Abyad).

Figure 6.18: Summary of published palaeomagnetic data (black outlined clock diagrams) from the Semail ophiolite, combined with results from this study (red outlined clock diagrams) (modified from Weiler, 2000; base map modified from e.g. Lippard et al. 1986; Nicolas & Boudier, 2011). Arrows = mean tilt corrected remanence directions. Sources: L = Luyendyk & Day (1982); P = Perrin et al. (1994, 2000); S = Shelton (1984); T = Thomas (1988); F = Feinberg et al. (1999); W = Weiler (2000); M = Meyer (this study). Blue stars: WA = Wadi Abyad; SR = Somrah; WN = Wadi Nassif; KF = Wadi Khafifah.

Figure 6.19: Schematic model for rotation of the Semail ophiolite: (a) formation by spreading along a NE-SW oriented axis, and acquisition of original remanences; (b) impingement of the Arabian margin on a previously intraoceanic subduction zone, leading to roll-back and clockwise rotation of the future ophiolite; and (c) eventual emplacement of the ophiolite on the Arabian margin, with remagnetization and anticlockwise rotation of the southern blocks related to the influence of the Jebel Akhdar structural high (culmination).

Figure A.1: Discrete dykes sampled within the Somrah layered gabbros; (a) Dyke site SR06 cutting gabbro layering at a high angle (photo credit: M. Anderson); (b) Detail of interfingering dyke material intruding into gabbros (photo credit: M. Anderson); (c) Dyke site SR07 cutting gabbro layering at a high angle (photo credit: M. Anderson).

Figure A.2: Equal area stereographic projections in geographic coordinates showing the results of magnetic fabric analysis of discrete dykes sampled within the layered and
foliated gabbros. WA38 from Wadi Abad; KF07 and KF09 from Wadi Khaffah; SR06 and SR07 from Somrah.

Figure B.1: Geological map of the Tuf area (north of Maqsad) with location of sampling sites (TU01-03). Inset map shows location of Tuf within the Samail block in the southern part of the Semail ophiolite (modified from Korenaga & Kelemen, 1997).

Figure B.2: Sampled gabbroic sill near Tuf: (a) View of sill in relation to surrounding mantle rocks (photo credit: M. Maffione); (b) Basal contact of sill, site TU01 (source: M. Maffione); (c) Middle of sill, site TU02 (photo credit: M. Maffione); (d) Top of sill, site TU03. Note the small-scale compositional layering (photo credit: M. Maffione).

Figure B.3: Equal area stereographic projections in geographic coordinates showing the results of magnetic fabric analysis of samples collected from a gabbroic sill within the MTZ near the village of Tuf.

Figure B.4: Characteristic Z-plots from the thermal demagnetization of specimens from the gabbroic sill sampled near the village of Tuf. LG = layered gabbros.

Figure B.5: Equal area stereographic projections in geographic (left) and tilt corrected (right) coordinates showing the distribution of site mean magnetization directions from Tuf. The large light purple symbol on each plot denotes the locality mean magnetization direction. Solid/open symbols = lower/upper hemisphere projections; ellipses = projection of α₉₅ cones of confidence around site means.

Figure C.1: Geological map of the central part of the Troodos ophiolite with accompanying lithostratigraphic log. Insert: detailed geological map of the area to the north of Agros with sampling localities marked with a solid black dot and labelled appropriately. Location of the study area within the ophiolite on Cyprus is given in the outline map of Cyprus (redrawn from the Geological Survey Department (Cyprus), Geological Map of Cyprus, Scale 1:250,000, 1995).

Figure C.2: Examples of demagnetization data from sheeted dykes at sites AG01 and AG02 (in geographic coordinates).

Figure C.3: Examples of demagnetization data from gabbros at sites AG03 and AG04 (in geographic coordinates).

Figure C.4: Examples of demagnetization data from sites AG05 and AG07 (in geographic coordinates).

Figure C.5: Examples of demagnetization data from site AG08 (in geographic coordinates).

Figure C.6: Site mean directions of magnetization for the gabbro sites and sheeted dyke sites sampled in Cyprus in geographic and tilt corrected coordinates for comparison to the Troodos mean direction of magnetization. A. Mean directions of magnetization taken from Table C.2 in geographic coordinates for sheeted dyke and gabbro sites with Troodos mean direction shown. The late dyke sampled at AG08 is shown in black to show is magnetization affinity with the gabbro suite compared to the sheeted dyke complex. B. Mean directions of magnetization in tilt corrected coordinates for the sheeted dyke and gabbro sites, once again the Troodos mean direction is shown. C. Formation mean directions of magnetization for the gabbro suite and sheeted dyke complex with the Troodos mean direction of magnetization shown.

Table 2.1: Breakdown of the number of samples collected and specimens prepared from each sampling locality.

Table 5.1: Summary of site-level magnetic susceptibility and AMS parameters.
Table 5.2: Comparison of the orientations of $K_{\text{max}}$ and $p\text{ARM}_{\text{max}}$ axes at site KF08, with calculated angular differences. Dec. = declination; Inc. = inclination.

Table 5.3: Summary of statistics relating to angular deviation of silicate crystal long-axes from $K_{\text{max}}$ axes in oriented thin sections of foliated gabbros

Table 6.1: Site mean magnetization directions from the Semail ophiolite in both geographic and tilt corrected coordinates. N = number of specimens; Dec. = declination; Inc. = Inclination; k = Fisher precision parameter; $\alpha_{95}$ = semi-angle of 95% cone of confidence; LG = layered gabbros; FG = foliated gabbros; VG = varitextured gabbros; D = dykes; DRZ = dyke-rooting zone.

Table 6.2: Locality mean magnetization directions

Table 6.3: Secondary magnetization component data from the Semail ophiolite, excluding results from the top of the Wadi Abyad crustal section. Secondary (i.e. low coercivity/unblocking temperature) component site mean magnetization directions from sites with N/NW-directed ChRM in both geographic and tilt corrected coordinates. N = number of specimens; Dec. = declination; Inc. = Inclination; k = Fisher precision parameter; $\alpha_{95}$ = semi-angle of 95% cone of confidence; LG = layered gabbros; FG = foliated gabbros; VG = varitextured gabbros; D = dykes.

Table 6.4: Overprint magnetization component data from the upper part of the Wadi Abyad crustal section in the Semail ophiolite. Secondary and tertiary component site mean magnetization directions from the varitextured gabbros (including intruding discrete dykes) and the dyke-rooting zone (i.e. sites with NE-directed ChRM). N = number of specimens; Dec. = declination; Inc. = Inclination; k = Fisher precision parameter; $\alpha_{95}$ = semi-angle of 95% cone of confidence; VG = varitextured gabbros; D = discrete dykes; DRZ = dyke-rooting zone.

Table A.1: AMS results from discrete dykes sampled within the layered and foliated gabbros of the Semail ophiolite. N = number of specimens; Dec. = declination; Inc. = inclination; $P_J$ = corrected anisotropy degree; T = shape parameter.

Table B.1: AMS results from a gabbroic sill within the MTZ sampled near the village of Tuf in the Semail ophiolite. N = number of specimens; Dec. = declination; Inc. = inclination; $P_J$ = corrected anisotropy degree; T = shape parameter.

Table C.1: Description and location details of sampling sites around the town of Agros (AG) in the Troodos ophiolite.

Table C.2: Palaeomagnetic data from the Troodos ophiolite. Site mean magnetization directions in both geographic and tilt corrected coordinates. N = number of specimens; Dec. = declination; Inc. = Inclination; k = Fisher precision parameter; $\alpha_{95}$ = semi-angle of 95% cone of confidence; SD = sheeted dykes; G = gabbros; D = dykes.
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AUTHOR'S DECLARATION

At no time during the registration for the degree of PhD Geology has the author been registered for any other University award without prior agreement of the Graduate Committee. This study was financed with the aid of a studentship form the Plymouth University.

Work submitted for this research degree at the Plymouth University has not formed part of any other degree either at Plymouth University or at another establishment.

Relevant scientific seminars and conferences have been regularly attended during the course of this research and findings have been presented at conferences in the form of oral and poster presentations. Numerous relevant workshops, fieldtrips and training courses were also attended, including several Plymouth University postgraduate courses for generic research and academic skills. It is anticipated that several manuscripts will be prepared and submitted for publication in peer-reviewed journals upon completion of this PhD. During this research extensive fieldwork in the Semail ophiolite, Sultanate of Oman, was undertaken in January/February of 2011 and 2012, which allowed for the collection of samples and the understanding of the geological context of the lower oceanic crust.

Conferences attended:

- November 2010 – CRES (Centre for Research in Earth Sciences) Postgraduate Conference 2010, Plymouth
- January 2011 – Magnetic Interactions 2011, Oxford
- November 2011 – CRES Postgraduate Conference 2011, Plymouth
- January 2012 – Internal Conference on the Geology of the Arabian Plate and the Oman Mountains, Muscat, Sultanate of Oman
- November 2012 – CRES Postgraduate Conference 2012, Plymouth
- January 2013 – Magnetic Interactions 2013, Lancaster
- April 2013 – EGU General Assembly 2013, Vienna, Austria
- November 2013 – CRES Postgraduate Conference 2013, Plymouth
- November 2014 – CRES Postgraduate Conference 2014, Plymouth
Presentations given:

- November 2010 – CRES Postgraduate Conference 2010, Plymouth
  - “Palaeomagnetic investigations of lower oceanic crustal accretion and tectonics (Oman and Troodos ophiolites)” – oral presentation
  - “Testing models of accretion of the lower oceanic crust using magnetic fabric analyses” – poster presentation
- January 2011 – Magnetic Interactions 2011, Oxford
  - “Testing models of accretion of the lower oceanic crust using magnetic fabric analyses” – poster presentation
- January 2013 – Magnetic Interactions 2013, Lancaster
  - “Magnetic fabric of gabbros beneath a fossil melt lens in the Oman ophiolite” – oral presentation
- April 2013 – EGU General Assembly 2013, Vienna, Austria
  - “Magnetic fabric analysis of gabbros beneath a fossil melt lens in the Oman ophiolite” – poster presentation
- November 2013 – CRES Postgraduate Conference 2013, Plymouth
  - “Magnetic fabric analysis of gabbros beneath a fossil melt lens in the Oman ophiolite” – oral presentation
- December 2014 – AGU Fall Meeting, San Francisco, USA
  - “New paleomagnetic data from the Wadi Abyad crustal section and their implications for the rotation history of the Oman ophiolite” – poster presentation
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Date: ............................................................
Chapter 1: Introduction

1.1 Rationale

Oceanic crust covers c. 60-70% of the Earth’s surface and yet the magmatic processes by which the lower crust accretes are still both poorly constrained and remain outstanding problems in geodynamics. Over the last 25 years the processes of oceanic crustal accretion have been the subject of increased attention and much debate, principally as a result of geophysical experiments over spreading ridges that demonstrate the presence of only thin melt lenses beneath spreading axes (Sinton & Detrick, 1992). For example, Sinton and Detrick (1992) presented seismic data from the fast-spreading East Pacific Rise that showed evidence for laterally extensive but very thin, magma chambers at the sheeted dyke-gabbro transition that overlie a large area of hot crystal mush. However, the processes that take place beneath the melt lens and the melt transportation system that feeds it were and still are poorly understood. This contrasts with earlier models in which the lower crust was inferred to form by crystallization in large, kilometric-scale, long-lived magma chambers (e.g. Cann, 1974; Pallister & Hopson, 1981), leading to an intensive debate about how the 3-4 km thick lower oceanic crust could be generated from thin melt lenses at a high crustal level.

Current models for formation of lower crust at fast-spreading axes focus on two end-member models: the gabbro glacier model and the multiple-sills model. The gabbro glacier model involves crystallization in a high-level melt lens located at the base of the sheeted dyke section, followed by downwards ductile flow towards the base of the crust and away from the ridge generating the gabbroic lower crust (e.g. Quick & Denlinger, 1993; Figure 1.1a). The
multiple-sills models involves the \textit{in situ} emplacement and crystallization of melt in a series of “sheeted sills”, generating the lower crust through multiple intrusive episodes without significant vertical transport of the products of crystallization (e.g. Kelemen \textit{et al.}, 1997; Figure 1.1b). These alternate models predict different variations in a number of quantifiable parameters through the lower crust (Figure 1.1c). For example, the gabbro glacier model predicts an exponential increase in strain with depth resulting from downward ductile flow. Hybrid models that amalgamate these two end members have also been put forward by some authors and involve a combination of downward ductile flow and sill intrusion (e.g. Boudier \textit{et al.}, 1996). More recently, MacLeod \& Yaouancq (2000) have proposed a new model that focuses on the transportation of melt though the crust itself.

These models have fundamentally different implications for the nature of heat and mass transfer between the Earth’s interior and exterior at constructive plate margins. They may be tested using core samples recovered by scientific ocean drilling (e.g. the Ocean Drilling Program and Integrated Ocean Drilling Program), but these are restricted to a few key locations and lack three-dimensional context. Alternatively, ophiolites provide extensive, easily accessible exposures of ancient oceanic lithosphere. The Semail ophiolite in Oman is the best analogue for fast-spreading crust, with large areas of well-exposed lower crustal rocks, and has been a key location over the past 30 years (e.g. Garrido \textit{et al.}, 2001; Coogan \textit{et al.}, 2002a). Supporters of both end-member models have used sections (sometimes the same section) of the ophiolite to validate their preferred models using a range of petrographic and geochemical techniques.
However, little attempt has been made to use magnetic fabric (anisotropy) analyses to quantify the variation in crystalline fabrics in the well-exposed gabbro sections in the Semail ophiolite. The models for accretion of the lower oceanic crust introduced above predict the development of very different
petrofabrics in various parts of the gabbro section. The principal focus of this research is, therefore, to perform systematic sampling and magnetic anisotropy analyses of these sequences in an attempt to distinguish between the potential models. Palaeomagnetic analyses of these same samples are also used to determine the nature of the remanent magnetization carried by the lower crustal gabbros of the Semail ophiolite. These provide insights into the timing of remanence acquisition and the pattern of remagnetization in these sections, leading to a reconciliation of contradictory directions of magnetization previously reported in the literature.

1.2 Ophiolites

Ophiolitic rock assemblages were first identified as fragments of oceanic crust exposed at the Earth’s surface by the work of Gass (1968) on the Troodos ophiolite (Cyprus). Gass (1968) envisaged that ophiolitic terranes were most likely formed at mid-ocean ridges (MORs) by sea-floor spreading with subsequent emplacement onto continental margins (Vine, 1966; Moores & Vine, 1971). Ophiolites were officially defined at the Penrose Conference in 1972 as a distinctive assemblage of mafic and ultramafic rocks with a classic layer-cake sequence of mantle peridotites overlain by gabbros, sheeted dykes, extrusive volcanic rocks, and a sedimentary cover (Anonymous, 1972; Robertson, 2004) (Figure 1.2). The definition also stated that this idealised model could be incomplete, tectonically dismembered, or metamorphosed (Anonymous, 1972). Dilek & Furnes (2011) updated this definition in light of the nearly 40 years of research into ophiolites since the 1972 conference, and defined ophiolites as allochthonous terranes of upper-mantle and oceanic crust that have been tectonically transported from their place of formation due to plate convergence.
(see Dilek & Furnes, 2011 for a full description). Ophiolites, therefore, provide the opportunity to study the architecture and evolution of the oceanic crust and have consequently become important natural laboratories for understanding the processes associated with oceanic crustal accretion.

Figure 1.2: Simplified section through an idealised Penrose-type ophiolite sequence.
1.2.1 Ophiolite types

It is now accepted that ophiolites represent analogies of oceanic crust, however, crust formed at mid-ocean ridges (i.e. MOR-type crust) is generally subducted during ocean closure and complicated tectonic processes are required to preserve MOR-type ophiolitic assemblages. Geochemical analyses by Miyashiro (1973), however, revealed that many ophiolites have chemical characteristics similar to that of island arcs and thus indicative of subduction zones (Shervais, 2001; Dilek & Flower, 2003; Robertson, 2004; Dilek & Furnes, 2011). Petrological investigations agreed with this subduction-related interpretation, as it was found that hydrous mineral phases could be observed within ophiolitic rocks along with wet-melt crystallization patterns (e.g. clinopyroxene before plagioclase) and orthopyroxene in the lower crustal rocks (Miyashiro, 1973; Shervais, 2001; Dilek & Furnes, 2011). The hydrous characteristics seen in ophiolitic rocks were attributed to the melting of the subducting slab with this wet-melt then being the source, along with material from the underlying mantle, for new oceanic crust (Shervais, 2001). The majority of ophiolites were thus interpreted to have formed in overriding plates above subduction zones (i.e. in a suprasubduction zone (SSZ) setting) and were subsequently referred to as SSZ-type ophiolites (e.g. Shervais, 2001; Dilek & Furnes, 2011). By contrast, MOR-type ophiolites are characterised by normal MOR geochemistry and, when well preserved, generally have a Penrose-type crustal architecture. However, the majority of MOR-type ophiolites are only preserved as extensively metamorphosed dismembered thrust sheets or blocks within ophiolitic mélanges (Robertson, 2002; Metcalf & Shervais, 2008; Dilek & Furnes, 2011). The identification that many of the most well studied and structurally intact ophiolites (such as the Troodos and Semail ophiolites) formed in a SSZ setting explained their unique geochemical, structural and petrological
characteristics, as well as how large fragments of oceanic crust were preserved during ocean closure and not lost due to subduction (Miyashiro, 1973; Shervais, 2001; Beccaluva et al., 2004; Dilek & Furnes, 2011).

1.2.2 Formation and emplacement of SSZ-type ophiolites

SSZ-type ophiolites are interpreted to form in the upper plate above or near to a subduction zone after the initiation of intraoceanic subduction (Miyashiro 1973; Shervais, 2001; Stern, 2004; Beccaluva et al., 2004; Dilek & Furnes, 2011). Intraoceanic subduction, like subduction beneath continental crust, occurs as a result of either a forcing factor, such as subduction zone locking leading to plate motion changes and/or subduction zone transference, or naturally, due to gravitational instability of the oceanic crust, resulting in passive margin or transform collapse (Figure 1.3a) (Stern, 2004). Whether intraoceanic subduction initiates at the ridge (the MOR) or off-axis either at a transform or fracture zone is, however, still a subject of debate (Searle & Cox, 1999, Shervais, 2001; Stern 2004). It is now widely accepted that the majority of ophiolites formed within a SSZ-type setting, shortly after the initiation of subduction as a direct result of slab rollback of the down-going plate, causing a brief but rapid period of extension, rifting and subsequent sea-floor spreading in the protoarc-forearc region before island arc magmatism/volcanism can take hold (Shervais, 2001; Stern, 2004; Dilek & Thy, 2009; MacLeod et al., 2013) (Figure 1.3b). Slab rollback can also lead to back-arc spreading and the development of geochemically MOR-type crust as spreading removes the nascent ridge away from the subducting slab (Figure 1.3c) (Beccaluva, et al., 2004). The preservation of back-arc crust is, however, harder to achieve (Robertson, 2002; Stern, 2004; Dilek & Flower, 2003).
Figure 1.3: Schematic model of SSZ-type ophiolite formation: (a) Induced or spontaneous intra-oceanic thrusting leads to the formation of an subduction zone within an MORB-type ocean basin; (b) Slab rollback of the down-going plate results in rapid formation of SSZ-type crust (green) in the protoarc-forearc region of the upper plate leading to a geochemical progression from MORB-type to IAT-type to SSZ-type crust; (c) If the subduction zone is allowed to mature, continued slab rollback can subsequently lead to back-arc spreading (light green) with asthenospheric diapirism and increased fluid influx causing partial melting of the shallow sub-arc mantle leading to the production of boninitic melts (modified from Shervais, 2001; Beccaluva et al., 2004; Stern, 2004; Dilek & Furnes, 2011).
The evolution of the magmatic suite of SSZ-type crust displays a geochemical progression from MOR-type crust to island arc tholeiites (IAT) and finally boninitic lavas, which are indicative of a SSZ setting (Dilek & Furnes, 2011) (Figure 1.4). The initial stages of subduction generate a magma with a MOR-like composition produced by decompression melting of fertile lherzolite-type mantle and this magma intrudes into and onto the original MOR-type crust of the upper plate (Shervais, 2001; Beccaluva et al., 2004; Dilek and Furnes, 2011; MacLeod et al., 2013). As subduction progresses, slab rollback initiates rapid sea-floor spreading in the protoarc-forearc region with the generation of melt that is strongly influenced by slab dehydration, producing IAT-type oceanic crust (Shervais, 2001; Beccaluva et al., 2004; Stern 2004; Dilek et al., 2007; Dilek & Furnes, 2011; MacLeod et al., 2013). As IAT-type oceanic crust continues to be produced, mantle melting and metasomatism (due to fluid influx) converts the lherzolite-type mantle to a depleted harzburgite-type mantle source (Shervais, 2001; Beccaluva et al., 2004; Dilek & Furnes, 2011). Continuing slab rollback then causes asthenospheric diapirism and further fluid influx resulting in partial melting of shallow sub-arc mantle material, and production of boninitic melts from a highly depleted mantle source (refractory harzburgite) in the protoarc-forearc region (Shervais, 2001; Beccaluva et al., 2004; Dilek & Furnes, 2011). An inter/back-arc-spreading basin may also develop, geochemically showing a gradual return from IAT to MOR-type melt as spreading removes the ridge away from the subducting slab (Beccaluva, et al., 2004).
When subduction is long-lived the chances of ophiolite preservation decreases as island-arc magmatism can take hold if the subduction zone is allowed to mature and become stable (Shervais, 2001). The production of ophiolitic assemblages is, therefore, likely to be short-lived (c. <10 m.y.) followed by a sudden termination as a result of trench-margin collision (Shervais, 2001; Robertson, 2002; Dilek et al., 2007). Collision of the active trench with a passive margin leaves ophiolitic crust trapped above the subduction zone and unable to be subducted (Shervais, 2001). The advancing margin then attempts to subduct, resulting in the emplacement (obduction) of the ophiolite onto the margin; the buoyancy of the partially subducted margin, typically of continental affinity, then locks up the subduction zone (Shervais, 2001; Beccaluva, 2004; Stern, 2004) (Figure 1.5). Subsequent rebound of the partially subducted margin results in exhumation of the ophiolite above the surface (Shervais, 2001).

Figure 1.4: Architectural and geochemical evolution of the magmatic suite of SSZ-type ophiolites with time (modified from Dilek & Furnes, 2011).
Figure 1.5: Schematic model of SSZ-type ophiolite emplacement: (a) Continued subduction eventual leads to trench-margin collision and results in the newly formed ophiolitic crust becoming trapped about the subduction zone and, therefore, unable to subduct; (b) Emplacement of the ophiolitic crust occurs as the advancing passive margin attempts to subduct, however, the buoyancy of continental crust prevents it and, therefore, causes the subduction zone to lock-up; (c) Rebound of the partially subducted continental passive margin leads to the exhumation of the ophiolite above the surface (modified from Shervais, 2001; Beccaluva et al., 2004; Stern, 2004; Dilek & Furnes, 2011).
1.2.3 Analogues or anomalous?

As a result of their formation in a protoarc-forearc setting above a subduction zone, SSZ ophiolites can be seen as anomalous, and their appropriateness for testing models of oceanic crustal accretionary processes taking place a MORs must, therefore, be considered carefully. MOR-type crust typically subducts and is, therefore, rarely preserved (Robertson, 2002). However, when preserved, it usually forms complex dismembered assemblages of oceanic crustal material (Lippard et al., 1986; Robertson, 2002). Subduction-related ophiolites on the other hand generally preserve large, structurally intact sheets of oceanic crust (e.g. the Semail and Troodos ophiolites) and, therefore, offer excellent field laboratories to test models of oceanic crustal accretion (e.g. Kelemen et al., 1997; Coogan et al., 2002a), as they provide extensive, three-dimensional exposures of crustal levels that are difficult to access in the modern oceans by scientific drilling alone (Lippard et al., 1986; Robertson, 2002).

A comparison of ophiolite types by Dilek & Furnes (2011) suggested that SSZ ophiolites are to be considered the most anomalous given their mode of formation and geochemistry. Nevertheless, Kelemen et al. (1997) state that compositional similarities between the Semail ophiolite lavas and mid-ocean ridge basalts (MORBs) suggest that petrogenetic processes involved in their formation must have been similar to the processes that take place at normal MORs. MacLeod et al., (2013) however, warn that although SSZ-type crust is clearly formed by seafloor spreading, the interaction of water with the melt might have hitherto unknown chemical and physical consequences on melt transportation, bulk fractionations and, therefore, the crustal architecture. However, SSZ ophiolites, such as the Semail and Troodos ophiolites, remain invaluable assets when studying oceanic crustal processes (e.g. Shervais, 2001; Robertson, 2004 and Dilek et al., 2007).
1.3 **Aims and objectives**

The key aim of this research is to apply magnetic fabric and palaeomagnetic techniques to systematically analyse the gabbroic sequence of the Semail ophiolite of Oman in order to investigate the processes by which the lower oceanic crust forms and deforms.

Specific objectives are:

1) To collect oriented samples from a suite of key lower crustal localities within the Semail ophiolite.

2) To investigate the relationships between principal axes of anisotropy of low field magnetic susceptibility (AMS) ellipsoids and macroscopic layering and microscopic preferred crystal alignments in layered gabbros with a variety of modal compositions.

3) To perform anisotropy of remanence experiments on selected samples, and integrate these with petrofabric observations on oriented thin sections in order to establish the source and significance of the AMS signal in these rocks.

4) To use these data to determine preferred mineral alignments and degrees of anisotropy at key levels in the gabbro section in order to test alternative models of magmatic accretion. Specific questions include whether magnetic fabric data can quantify variations in strain through the lower crust (as predicted by the gabbro glacier model), and whether fabrics immediately below the fossil melt lens (in the foliated gabbros) provide evidence of melt migration pathways.

5) To undertake palaeomagnetic experiments on samples from various localities in Oman to isolate characteristic remanence directions and to use these data to determine whether relative rotations and/or remagnetization of crustal blocks has occurred, either during seafloor
spreading or during later emplacement of the ophiolite. Specifically, these data will be used to determine whether systematic sampling through crustal sections allows the extent of remagnetization to be constrained, thereby allowing previously reported data to be interpreted with more confidence.
Chapter 2: Magnetic theory and methodologies

2.1 Introduction

The Earth’s magnetic field, generated by the motion of the fluid outer core, can impart a magnetization in rocks that can be preserved for millions of years (Butler, 1998; Tauxe, 2009). The study of the magnetic properties and magnetization of rocks is known as palaeomagnetism. The range of techniques incorporated into palaeomagnetic studies not only enables research into the variation of the Earth’s magnetic field over time but can also be used to study the formation and history of the rocks themselves. The Earth’s essentially dipolar magnetic field generates magnetic field lines that come out of the southern hemisphere (pointing upwards) and go into the northern hemisphere (pointing downwards) when in a normal polarity state (Butler, 1998; Tauxe, 2009). The field at any given point is a vector that can be described in polar coordinates by its declination, inclination and intensity (measured in nT) or in Cartesian coordinates by components along orthogonal axes (x, y, z or N, E, down) (Butler, 1998; Tauxe, 2009). The declination (D) is the angle clockwise from the geographic North Pole while the inclination (I) is the angle from horizontal, being positive downward and negative upward (Butler, 1998; Tauxe, 2009). As well as studying the changes of the Earth’s magnetic field over geological time (either on the timescales of secular variation or full reversals) and tracking the movement of the tectonic plates, magnetism can also be used to study the nature of the rocks themselves. Using the magnetic properties of rock-forming minerals, the processes involved in their formation through the development of fabrics can be analysed. These magnetic fabrics, produced either during their formation (e.g. related to primary layering) or during
subsequent tectonic events (e.g. structural foliations and/or lineations), are taken to reflect the shape of grains and/or the alignment of minerals, which can then be related to observed petrofabrics and subsequently interpreted in terms of geological processes and modes of formation (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009).

At the atomic level, magnetization occurs because of the orbit and spin of electrons around a nucleus that can generate an atomic magnetic moment; everything, therefore, has magnetic potential but different substances react differently when exposed to a magnetic field dependent upon their composition (Tarling & Hrouda, 1993; Butler, 1998; Nesse, 1999; Tauxe, 2009). The ability of rocks to record the Earth’s magnetic field at the time of their formation is, therefore, dependent on the minerals they contain. Rock-forming minerals are either diamagnetic, paramagnetic or ferromagnetic, with each behaving differently when exposed to a magnetic field (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009). When an external magnetic field (\(H\)) is applied to a diamagnetic material, a magnetization in the opposite direction to the applied field will be produced, while a paramagnetic substance will acquire a magnetization in the same direction as the applied field. However, when the applied field is removed both substances will lose their magnetization (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009). Ferromagnetic minerals on the other hand are able to retain a magnetization (a remanent magnetization) after an applied field is removed, recording the direction and intensity of the magnetizing field (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009). This property of ferromagnetic minerals forms the basis of palaeomagnetism and allows rocks to retain records of the Earth’s magnetic field over geological time. Nevertheless, diamagnetic and paramagnetic minerals may make significant contributions to
the magnetic properties of rocks in terms of magnetic fabric studies in the absence of significant concentrations of ferromagnetic minerals.

The intensity of magnetization \( M \) in any given substance is proportional to the applied magnetic field \( H \). In this relationship, the constant of proportionality is the magnetic susceptibility \( k \), and can be expressed as:

\[
k = \frac{M}{H}.
\]

Susceptibility is dimensionless in SI units, can have anisotropy (see section 2.2.3.6) and also varies with mineralogy and grain size (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009). Diamagnetic minerals have a negative susceptibility with \( k < -10^{-5} \) [SI], paramagnetic minerals have positive susceptibility \( k \) between \( 10^{-5} \) and \( 10^{-3} \) [SI], while ferromagnetic minerals have strong positive magnetic susceptibility with \( k \) in the order \( > 10^{-2} \) [SI] (cf. Figure 3.1) (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009).

### 2.2 Magnetic minerals and their properties

Rocks are composed of predominantly diamagnetic and paramagnetic minerals with a small percentage of ferromagnetic particles dispersed throughout. The intensity of magnetization will be determined by the concentration and mineralogy of the ferromagnetic fraction while the total (bulk) susceptibility \( k \) will be the sum of the diamagnetic, paramagnetic and ferromagnetic fractions combined. Ferromagnetic particles at \( >0.1\% \) concentration will, however, dominate the susceptibility of a specimen due to their high \( k \). If no ferromagnetic particles are present, the paramagnetic minerals (if \( >1\% \)) will dominate the susceptibility (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009). The magnetic behaviour (either dia-, para-, or ferromagnetic) and the properties of a substance are dependent upon the
2.2.1 Diamagnetism

Minerals containing atoms with no unpaired electrons (i.e. the spins of electrons are in equilibrium) will have a net atomic magnetic moment of zero. These minerals exhibit diamagnetic behaviour and some typical examples include quartz [SiO$_2$] and calcite [CaCO$_3$]; sulphur [S] is also a diamagnetic substance (Butler, 1998; Nesse, 1999; Tauxe, 2009). Diamagnetic substances have weak, negative susceptibilities and acquire a small magnetization in the opposite direction to that of an applied field, which reduces to zero when the applied field is removed (Tarling & Hrouda, 1993; Butler, 1998; Nesse, 1999; Tauxe, 2009). The diamagnetic response to a magnetizing field is illustrated in Figure 2.1a.

![Figure 2.1: The magnetization (M) versus applied magnetizing field (H) for: (a) diamagnetic materials where k (magnetic susceptibility) is negative; (b) paramagnetic materials where k is positive; and (c) ferromagnetic (s.l.) material where the path of magnetization exhibits hysteresis and is, therefore, not a simple constant (n.b. a paramagnetic response to an applied field is 10-100 times stronger than a diamagnetic response, whereas a ferromagnetic (s.l.) response is exponentially stronger than both diamagnetic and paramagnetic responses) (modified from Butler, 1998).](image-url)
2.2.2 Paramagnetism

Paramagnetic minerals have atoms that have unpaired electrons (i.e. the spins of the electrons are not in equilibrium creating a bias) and, therefore, generate an atomic magnetic moment (Butler, 1998; Nesse, 1999; Tauxe, 2009). This atomic magnetic moment is randomly orientated but when exposed to an external magnetic field will align parallel with the applied field producing a relatively weak magnetization with a positive magnetic susceptibility (Tarling & Hrouda, 1993; Butler, 1998; Nesse, 1999; Tauxe, 2009). Due to the lack of interaction between neighbouring atomic magnetic moments, once the applied field is removed the magnetization reduces to zero (Tarling & Hrouda, 1993; Butler, 1998; Nesse, 1999; Tauxe, 2009). In nature, most rock-forming minerals are paramagnetic; examples include olivine [(Mg,Fe)\(_2\)SiO\(_4\)], biotite [K(Fe,Mg)\(_3\)AlSi\(_3\)O\(_{10}\)(OH)\(_2\)] , plagioclase [NaAlSi\(_3\)O\(_8\)-CaAl\(_2\)Si\(_2\)O\(_8\)] and other iron-bearing silicate minerals (Tarling & Hrouda, 1993; Nesse, 1999). The paramagnetic response to a magnetizing field is illustrated in Figure 2.1b.

2.2.3 Ferromagnetism (sensu lato)

Certain materials are capable of retaining a magnetization in the absence of an external field and these magnetic substances are in a broad sense termed ferromagnetic (sensu lato). Similar to paramagnetic substances, the presence of unpaired electrons within their atomic structure generates an atomic magnetic moment that if exposed to a magnetizing field will produce a magnetization parallel (with positive susceptibility) to the applied field (Tarling & Hrouda, 1993; Butler, 1998; Nesse, 1999; Tauxe, 2009). However, unlike paramagnetic substances, strong interactions (coupling) occur between adjacent magnetic moments, the effect of which is to produce a magnetization of greater intensity than that generated in a paramagnetic substance in the
same strength field and one that will be retained after the magnetizing field is removed (Tarling & Hrouda, 1993; Butler, 1998; Nesse, 1999; Tauxe, 2009). The path of magnetization of a ferromagnetic substance exhibits hysteresis (see section 2.2.3.2) and is therefore irreversible with magnetic susceptibility not a simple constant, as illustrated in Figure 2.1c (Butler, 1998; Tauxe, 2009).

The ability of a ferromagnetic (s.l.) material to retain a magnetization (i.e. a remanent magnetization) is inversely proportional to temperature, as with increasing temperature the coupling of the atomic magnetic moments becomes weaker and above a threshold temperature, known as the Curie temperature, the coupling is destroyed and the material will behave paramagnetically (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009). This relationship is known as the Curie-Weiss law and is given by the equation:

$$\frac{M}{H} = \frac{\mu_0 N m^2}{3k(T - T_c)} \equiv X_f$$

However, if the material is allowed to cool back to below its Curie temperature, it will once again become ferromagnetic (s.l.), and if this takes place in the presence of a magnetizing field, such as the Earth’s magnetic field, the material can acquire a magnetization parallel to the applied field that can become fixed, preserving the direction and intensity of the Earth’s magnetic field for millions or even billions of years (Tarling & Hrouda, 1993; Butler, 1998; Nesse, 1999; Tauxe, 2009).

Primarily of magmatic origin, ferromagnetic (s.l.) minerals are often found in rocks as accessory minerals. The most common magnetic minerals are the iron oxides, magnetite [Fe₃O₄] and hematite [Fe₂O₃]. If either is present they will dominate the magnetic properties of the rock (Tarling & Hrouda, 1993; Butler, 1998; Nesse, 1999; Tauxe, 2009). Other significant ferromagnetic (s.l.) minerals
include the iron oxyhydroxide, goethite \([\text{FeO(OH)}]\) and the iron sulphides greigite \([\text{FeFe}_2\text{S}_4]\) and pyrrhotite \([\text{Fe}_{1-x}\text{S}]\) (Tarling & Hrouda, 1993; Nesse, 1999).

2.2.3.1 **Types of ferromagnetism (s.l.)**

Ferromagnetic (s.l.) minerals retain a remanent magnetization due to the strong coupling of adjacent atomic magnetic moments and depending on the crystallographic arrangement of the mineral, the electron spins coupling the magnetic moments can either be parallel or anti-parallel (Tarling & Hrouda, 1993; Butler, 1998; Nesse, 1999; Tauxe, 2009). Ferromagnetism (s.l.) can consequently be divided into groups related to the coupling of the atomic magnetic moments as illustrated in Figure 2.2.

When the spins of adjacent magnetic moments are parallel, a large focused magnetization is produced (Figure 2.2a); this type of alignment is termed ferromagnetism (senso stricto) and can be observed in pure iron \([\text{Fe}]\) (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009). If, however, the spins are perfectly antiparallel (equal both ways), the magnetic moments cancel each other and there is, therefore, no net magnetization (Figure 2.2b); this is antiferromagnetism (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009). Sometimes, however, the spins of antiferromagnetic minerals are not perfectly antiparallel but are in fact inclined or tilted, which can produce a weak but significant net magnetization (Figure 2.2c); this canted antiferromagnetism occurs in hematite (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009). Figure 2.2d illustrates ferrimagnetism. Here, adjacent spins are aligned antiparallel but the magnitudes of each magnetic moment are unequal. This imbalance results in a net magnetization aligned in the direction of the greater magnitude moment;
this is the situation for magnetite (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009).

![Diagram](image)

Figure 2.2: Types of ferromagnetism (s.l.) in response to a magnetizing field: (a) ferromagnetism (s.s.) produces a large focused magnetization parallel to the applied field; (b) antiferromagnetism results in a net magnetization of zero; (c) canted antiferromagnetism produces a weak but potentially significant magnetization; and (d) ferrimagnetism results in a net magnetization aligned in the direction of the greater magnitude magnetic moments (modified from Tauxe, 2009).

### 2.2.3.2 *The hysteresis loop*

As previously mentioned, the path of magnetization (and susceptibility) of a ferromagnetic (s.l.) material varies as a function of the applied field and is described by a hysteresis loop (Butler, 1998; Tauxe, 2009). Unlike diamagnetic and paramagnetic substances, where magnetization (M) increases linearly within an increasingly magnetizing field (H) (Figure 2.1a & 2.1b), ferromagnetic (s.l.) materials can reach a maximum magnetization value called the saturation magnetization (Ms) (Butler, 1998; Tauxe, 2009) (Figure 2.1c) The value of this can change with temperature and the ferromagnetic (s.l.) material in question (Butler, 1998; Tauxe, 2009).

The hysteresis loop of SD magnetite (Figure 2.3) indicates that when a ferromagnetic (s.l.) material with an initial magnetization, M = 0, is magnetized, its magnetization (M) will increase with increasing applied field (H) to point 1 on Figure 2.3. As the applied field is increased the magnetization of the material eventually plateaus and cannot increase further; here at point 2 the saturation magnetization (Ms) has been reached. At point 2 the magnetization of all
ferromagnetic (s.l.) grains within the material are aligned parallel with the applied field. Upon removal of the magnetizing field, the magnetization of the material follows the path from point 2 to point 3, not returning to the origin ($M = 0$) but retaining a remanent magnetization ($M_r$). The magnetization can only reach $M = 0$ again by applying a magnetic field in the opposite direction (a back-field) to force it back to zero. The magnetization will, therefore, follow the path from point 3 to point 4, where $M = 0$. The magnetic field needed to achieve this is called the bulk coercive force (or the coercivity of remanence) ($H_c$). At $H_c$, the magnetizations of ferromagnetic (s.l.) particles cancel each other out, resulting in a zero net magnetization. The hysteresis loop is completed by forcing the ferromagnetic (s.l) particles into a back-field saturation magnetization ($-M_s$), then back to positive saturation magnetization ($M_s$). As can be seen, once magnetized, ferromagnetic (s.l.) particles are very resistant to demagnetization and easily acquire a remanent magnetization.

Figure 2.3: The hysteresis loop showing the magnetization behaviour of ferromagnetic (s.l.) materials. $M$ is the magnetization, $H$ is the applied magnetic field, $M_s$ is the saturation magnetization, $M_r$ is the remanent magnetization, $H_c$ is the coercive force or coercivity of remanence. See text for details (after Butler, 1998; Tauxe, 2009).
The Curie/Néel temperature, blocking temperature and relaxation time

In section 2.2.3 the relationship between temperature and ferromagnetic (s.l.) properties was introduced. When above a certain temperature (the Curie temperature), ferromagnetic (s.l.) materials will behave paramagnetically (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009). The Curie temperature (also known as the Néel temperature for antiferromagnetic substances) varies depending on the material. For example, the Curie temperature of pure magnetite is 578°C, while for pure hematite it is 680°C (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009). The other common ferromagnetic (s.l.) minerals have considerably lower Curie temperatures, with greigite at 350°C, pyrrhotite at 320°C and goethite at 127°C (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009). As multiple ferromagnetic (s.l.) minerals can be present within a rock, the wide range of Curie temperatures observed can significantly help in their identification.

Above the Curie temperature, ferromagnetic (s.l.) minerals behave like paramagnetic materials and are unable to retain a magnetization. Below the Curie temperature, however, certain ferromagnetic (s.l.) minerals (e.g. magnetite) can hold onto a remanent magnetization for millions of years (Butler, 1998; Tauxe, 2009). As the Curie temperature for any ferromagnetic (s.l) material is approached, the coupling of the atomic magnetic moments gradually decreases. At this point, the material is said to be superparamagnetic and can only hold onto a magnetization for a very short amount of time (less than 100 seconds); any further increase in temperature will destroy the coupling completely and all ferromagnetic (s.l.) properties will be lost (Butler, 1998; Tauxe, 2009). However, a lowering of the temperature will cause an exponential rise in the time the magnetization can be retained. If the temperature is allowed
to cool further, the relaxation time (the time it takes for the magnetization of a substance to decay to zero) will continue to increase, locking in the remanence for increasingly longer periods of time (Butler, 1998; Tauxe, 2009). Therefore, the relaxation time depends upon the inverse relationship between magnetic anisotropy and thermal energy and is given by:

\[
\tau = \frac{1}{C} \exp \left( \frac{\text{anisotropy energy}}{\text{thermal energy}} \right) = \frac{1}{C} \exp \left( \frac{a_0 V H_c(T) M_s(T)}{2kT} \right),
\]

where \( \tau \) is the relaxation time, \( \frac{1}{C} \) is a frequency factor \( (10^9 - 10^{10}/s) \), \( V \) is the volume of the grain, \( H_c(T) \) is the grains coercivity (see section 2.2.3.5) at temperature \( T \), \( M_s(T) \) is the saturation magnetization at temperature \( T \), \( k \) is the Boltzmann’s constant and \( T \) is the absolute temperature (Néel, 1955; Tauxe, 2009). The transition from superparamagnetic to ferromagnetic (s.l.) behaviour or vice versa is known as the blocking/unblocking temperature and can vary due to mineralogy and grain size (Butler, 1998; Tauxe, 2009).

### 2.2.3.4 Grain size, domains and their magnetic properties

The intensity of magnetization and the ability of a ferromagnetic (s.l.) particle to retain a magnetization for a long period of time while also being resistant to demagnetization are dependent upon the size of the magnetic particle itself (Butler, 1998; Tauxe, 2009). As the size of a ferromagnetic (s.l.) particle increases, its internal magnetic energy increases due to the greater number of atomic magnetic moments it contains. In order to reduce this internal energy, the spins of magnetic moments will organise themselves into regions of opposing magnetizations (magnetic domains) each with its own magnetic dipole moment (Butler, 1998; Tauxe, 2009). Magnetite grains, for example, of >10 µm may contain dozens of magnetic domains and are referred to as multi-domain (MD) grains (Butler, 1998; Tauxe, 2009). MD magnetite is easily magnetized
and becomes magnetically saturated in fields of only a few 10’s mT. This low saturation magnetization leads to a low coercive force and, therefore, a low resistance to demagnetization (Butler, 1998; Tauxe, 2009). As grain size decreases the number of magnetic domains required to lower the internal energy also decreases. Magnetite grains between 1-10 µm may contain only a few magnetic domains or vortex structures and are called pseudo-single-domain (PSD) grains (Butler, 1998; Tauxe, 2009). PSD grains can display large magnetic moments and are able to retain remanent magnetizations for a significant time whilst also being resistant to demagnetization (Butler, 1998; Tauxe, 2009). As the grain size decreases further still, it no longer becomes necessary to form multiple magnetic domains resulting in just a single magnetic moment (Butler, 1998; Tauxe, 2009). These single-domain (SD) grains are <1 µm in size and are magnetically strong, being able to hold on to a remanent magnetization for a considerable amount of time and in very high demagnetizing fields and only become magnetically saturated in fields of approximately 300 mT (Butler, 1998; Tauxe, 2009). SD and PSD grains, therefore, make the most effective palaeomagnetic recorders due to their strong resilience to demagnetization while MD grains although easily magnetized lose their magnetization more easily and quickly (Butler, 1998; Tauxe, 2009).

2.2.3.5 Coercivity

In section 2.2.3.2 the idea of coercive force was introduced as the strength of the applied magnetic field needed to reduce the net magnetization of a ferromagnetic (s.l.) substance to zero (Butler, 1998; Tauxe, 2009). Often referred to simply as coercivity, the coercive force can be (loosely) considered as the room/ambient temperature equivalent of the unblocking temperature.
Each individual ferromagnetic (s.l.) particle will have its own coercivity, above which it will be unable to hold onto a remanent magnetization (Butler, 1998; Tauxe, 2009). The coercivity of a ferromagnetic (s.l.) particle is, therefore, its resistance to demagnetization, which varies according to the type of magnetic grain and the domain structure of the mineral (either SD, PSD or MD grains). Typically, MD grains will have low coercivities while SD and PSD are stronger magnetically speaking and have higher coercivities and, therefore, greater resistance to demagnetization (Butler, 1998; Tauxe, 2009).

### 2.2.3.6 Anisotropy

Magnetic susceptibility is the relationship between an applied magnetic field and the intensity of the resulting induced magnetization (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009). The intensity of induced magnetization can, however, vary with the direction of the applied field (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009). Some substances show no directional dependence of magnetization and are termed magnetically isotropic. Most substances, however, do show this relationship and are called magnetically anisotropic. The anisotropy of individual grains depends upon their internal crystalline structure, which controls the magnetic moment of the particle, and the shape of the grain itself (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009). The crystalline structure produces easy axes along which magnetization will tend to align, whereas an elongated shape to a grain can force its magnetization to align along the long axis (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009). When only ferromagnetic (s.l.) grains are considered, magnetite will show strong shape anisotropy, whereas hematite is controlled by its crystalline structure. However, grain size also plays a vital role in determining anisotropy (Tarling &
Hrouda, 1993; Butler, 1998; Tauxe, 2009). Small SD magnetite grains show maximum susceptibility perpendicular to their long axes, whereas the maximum susceptibility in MD grains falls along their long axes. This SD effect can produce inverse fabrics when anisotropy is measured and will be described in detail later (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009).

2.3 Types of remanent magnetization

Given the nature of this investigation into the formation of the lower oceanic crust only magnetizations that are commonly found in igneous rocks will be discussed.

2.3.1 Thermoremanent magnetization (TRM)

Igneous rocks, either intrusive or extrusive, typically acquire a thermoremanent magnetization (TRM) as they cool from above the Curie temperature to below the blocking temperatures of its constituent ferromagnetic minerals in the presence of the Earth’s magnetic field (Butler, 1998; Tauxe, 2009). As cooling takes place the magnetic domains of the ferromagnetic (s.l.) minerals align themselves parallel to the external magnetic field and eventually with further cooling become fixed (Butler, 1998; Tauxe, 2009). Igneous rocks, however, contain a number of varying grain shapes and sizes, as well as magnetic domain types (e.g. SD, PSD or MD), resulting in a possible distribution of blocking temperatures from the Curie temperature down to approx. 20°C (Butler, 1998; Tauxe, 2009). Once cooled through its blocking temperature, each individual grain will no longer be affected by thermal perturbations. The overall magnetization of the rock is the vector sum of the individual remanences of constituent ferromagnetic grains. Each individual grain
may have a magnetization that does not parallel the Earth’s field, but the statistical sum of the signal from all grains will provide an accurate record of the ambient magnetic field (Butler, 1998; Tauxe, 2009). Cooling through the spectrum of blocking temperatures of ferromagnetic grains in a rock can be geologically instantaneous, as with some extrusive rocks, or can take place over thousands or even millions of years as seen with intrusive rocks (Butler, 1998; Tauxe, 2009). It is possible, then, that TRMs record multiple vector directions if reversals of the Earth’s magnetic field take place while cooling is still on-going, or just one direction if the magnetic field was very stable for long periods of time or due to rapid cooling. The resulting TRM will, therefore, be the sum of all the magnetizations acquired over the range of blocking temperatures during cooling (Butler, 1998; Tauxe, 2009). TRMs are normally interpreted as primary magnetizations of igneous rocks, acquired at (or very soon after) the time of rock formation, but may also result from reheating of host rocks (or all types) by intrusions (e.g. in the baked contact of a dyke) (Butler, 1998; Tauxe, 2009).

2.3.2 Chemical remanent magnetization (CRM)

Over time the magnetic mineralogy of a rock can alter as a response to changing geological/chemical environments, as a result pre-existing minerals (possibly also ferromagnetic) can be altered to form (new) ferromagnetic (s.l.) minerals or entirely new ferromagnetic (s.l.) minerals can precipitate from a circulating solution (Butler, 1998; Tauxe, 2009). Formation of new ferromagnetic (s.l.) minerals by chemical alteration/change below the Curie temperature (of the specific mineral growing) and in the presence of an external field (e.g. the Earth’s magnetic field) can result in the acquisition of a chemical remanent magnetization (CRM) that records information of the external field (i.e. Earth’s
magnetic field) at the time of mineral formation and can potentially be geological significant (Butler, 1998; Tauxe, 2009). During acquisition of a CRM, temperature can be considered constant, however, individual grains formed during chemical alteration/precipitation grow from an initial volume of zero and are, therefore, initially superparamagnetic (Butler, 1998; Tauxe, 2009). As grain growth continues, however, grain volume increases and individual grains will, therefore, experience a dramatic increase in relaxation time from superparamagnetic to stable single domain at a critical threshold known as the blocking volume at which point they are able to hold a stable remanence (Butler, 1998; Tauxe, 2009). In principle, therefore, CRMs are similar to TRMs, but the acquisition of a stable CRM is related to progressive grain growth of the magnetic mineral below the Curie temperature. CRMs are most commonly encountered in sedimentary rocks (such as redbeds), however, processes such as sea-floor metamorphism (e.g. the serpentinization of olivine/hydrothermal alteration of clinopyroxene) may potentially produce CRMs in oceanic crustal rocks due to the formation of new ferromagnetic (s.l.) minerals (e.g. Borradaile & Gauthier, 2006).

2.3.3 Secondary magnetizations

Once cooled, ferromagnetic (s.l.) minerals within an igneous rock are able to preserve a record of the magnetic field for millions of years. However, geological processes, time and the Earth’s magnetic field itself can cause the acquisition of additional magnetizations. Most igneous rocks are, at some point after they have formed, affected by geological events that can result in alteration of the ferromagnetic (s.l.) minerals present (Butler, 1998; Tauxe, 2009). These events can result from numerous geological processes (e.g. deep
burial, obduction, serpentinization and contact metamorphism) and can result in the acquisition of new, secondary, magnetizations (Butler, 1998; Tauxe, 2009).

2.3.3.1  *Thermo-viscous remanent magnetization (TVRM)*

If an igneous rock is heated to temperatures below the Curie temperature and then allowed to cool in the presence of a magnetizing field (e.g. the Earth’s magnetic field), the ferromagnetic (s.l.) minerals with blocking temperatures at or below the peak temperature will become remagnetized (Butler, 1998; Tauxe, 2009). Ferromagnetic (s.l.) minerals with blocking temperatures above the peak temperature will, however, be left unaffected and, therefore, still record the original magnetization (Butler, 1998; Tauxe, 2009). Furthermore, if sufficient heating is achieved new magnetic minerals can form, either from the alteration of the existing mineralogy or by the addition of new material from an external source. These metamorphic minerals if ferromagnetic (s.l.) will acquire a magnetization much like a TRM as the temperature decreases (Butler, 1998; Tauxe, 2009).

Chemical alteration of iron-bearing minerals to ferromagnetic (s.l.) minerals can also take place at elevated temperatures due to the movement of fluids through the rock (Butler, 1998; Tauxe, 2009). The penetration of seawater, for example, related to hydrothermal circulation or obduction processes, can result in the formation of new ferromagnetic (s.l.) minerals such a MD magnetite, for example, during serpentinization. Interaction of seawater with olivine-bearing rocks of the lower crust and upper mantle can produce secondary magnetite, which will subsequently become magnetized. These secondary magnetizations related to elevated temperatures are loosely termed thermo-viscous remanent magnetization (TVRM) and can be stable over millions of years (Butler, 1998;
Tauxe, 2009). Although not related to the formation of the rock, these TVRMs reveal important information about the history of the rock after its formation.

### 2.3.3.2 Viscous remanent magnetization (VRM)

Magnetizations held by minerals with low relaxation times do not last over geological time and are, therefore, inherently unstable. These low relaxation time components tend to align themselves to any applied external magnetic field and change with the field as it changes (Butler, 1998; Tauxe, 2009). Rocks found at the Earth’s surface are constantly exposed to the Earth’s magnetic field, which although weak can impart a magnetization at ambient temperatures (Butler, 1998). Magnetic domains within the rock with low relaxation times will, over time, align themselves to the Earth’s magnetic field producing a viscous remanent magnetization (VRM) (Butler, 1998; Tauxe, 2009). The acquisition of VRMs depends upon the length of time the rock is exposed to the external field and the temperature at which exposure took place (Bardot & McClelland, 2000). VRMs are superimposed over more stable magnetizations and are perhaps the most common secondary magnetization and because of their short relaxation times they are usually easily removed by demagnetization. In the case of thermal demagnetization, a simple relationship between the blocking temperature and VRM acquisition time was derived by Bardot & McClelland (2000) for relatively short geological time intervals \(10^2 – 10^6\) years, as follows:

\[
T_b = 75 + 15 \log \text{(acquisition time in years)},
\]

where \(T_b\) is the blocking temperature. This, therefore, implies that any VRM acquired by rocks in the Semail ophiolite should be removed in temperatures below 170ºC (Bardot & McClelland, 2000). Additionally, most VRMs are
interpreted to record the present day geomagnetic field direction and are palaeomagnetically speaking regarded as noise (Butler, 1998; Tauxe, 2009).

2.3.4 Natural remanent magnetization (NRM)

The natural remanent magnetization (NRM) is the vector sum of the remanent magnetizations acquired by a rock sample during its geological history and can be composed of multiple magnetization components (Butler, 1998; Tauxe, 2009). The magnetization acquired at the time of formation, such as a TRM in igneous rocks, is known as the primary magnetization. However, this can be overprinted by subsequent secondary magnetizations such as TVRMs or VRMs that can be acquired over time (Butler, 1998; Tauxe, 2009). In some cases, the primary magnetization can be completely lost by high temperature metamorphism or other tectonic processes long after formation. Establishing the timing of acquisition of a remanence component can be difficult (see below) and it is rare that a magnetization can be identified as a primary remanence with total confidence. Hence, it has become good practice to refer to the most stable, highest unblocking temperature/coercivity component as the characteristic remanent magnetization (ChRM), and then to attempt to establish whether this is a primary or secondary magnetization using a variety of field tests of stability.

2.4 Palaeomagnetic field tests

The timing of remanence acquisition is usually determined using essentially relative dating techniques, these include:

1. Fold and tilt tests, where the distribution of remanence directions from folded/tilted sequences are compared before and after synthetically
removing the effects of tilting. If ChRM directions become more closely grouped after untilting (compare Figure 2.4c and 2.4f), this indicates that the ChRM was acquired before folding (positive fold test). Increased dispersion of ChRM directions after untilting indicates that the magnetization is post-folding in age (negative fold test). A plethora of statistical tests have been devised to determine the significance of fold tests, the most employed being those of McElhinny (1964) and McFadden & Jones (1981). In the case of complexly deformed sequences where relative vertical axis rotations may have occurred or where fold geometries cannot be constrained accurately, it is sometimes more appropriate to apply an inclination-only fold test formulation (e.g. Beck, 1980; Demarest, 1983; Arason & Levi, 2010) based on the distribution of inclinations before and after untilting.

2. Baked contact tests, where samples are collected from an igneous intrusion, and from its baked contact and host rock. If the host rock is much older than the igneous rock, then ChRM directions from the igneous (Figure 2.4a) and host (Figure 2.4c) rocks may be expected to be significantly different. The baked contact, however, should have the same direction as the igneous rock since its magnetization should be thermally reset during intrusion (Figure 2.4b). A positive test is where the igneous rock and baked contact have similar magnetization directions which are different from that of the host rock far from the intrusive contact, suggesting that the primary TRM has been isolated. A negative test (uniform directions throughout) suggests that all units have experienced a remagnetization event.
3. Conglomerate tests, where samples are collected from clasts within a conglomerate, and ChRM directions are determined for each clast by demagnetization experiments. Uniform directions of magnetization within individual clasts but a random distribution between clasts indicates that the magnetization of the source rock has been stable since at least the time of formation of the conglomerate (positive test; Figure 2.4d). Uniform directions between clasts indicate that the ChRM of the conglomerate was formed after deposition (negative test;
Figure 2.4e). A positive test carried out on an intraformational conglomerate strongly suggests that the ChRM of the source formation is a primary remanence, additionally, the demagnetization characteristics of the clasts should ideally mirror that of the original source rock as well.

4. Reversal tests (Figure 2.4g), which can be applied to sequences deposited or emplaced during different geomagnetic polarity chrons. A sequence that preserves an interpretable magnetostratigraphy is assumed to have retained its primary magnetizations.

2.5 Sample collection and specimen preparation

Sampling primarily focused on the collection of orientated samples for magnetic fabric and palaeomagnetic analysis from the Semail ophiolite, Oman. The Semail ophiolite is an approximately 600 km long, 150 km wide and 18 km thick thrust sheet of Neo-Tethyan oceanic crust exposed on the Arabian Peninsula (e.g. Pearce et al., 1981; Lippard et al., 1986; Searle & Cox 1999; Rioux et al., 2012, 2013). The ophiolite suffered relatively little tectonic deformation during its emplacement and thus features related to original oceanic accretion are generally very well preserved (e.g. Hopson et al., 1981; Boudier et al., 1996; Yaouancq & MacLeod 2000). Over two field seasons in Oman during February 2011 (18 days) and January 2012 (13 days) a total of 86 sites (including sub-sites) were sampled from five localities either by in situ coring using a petrol-powered rock drill or by hand sampling (Table 2.1). The most intensive sampling took place within Wadi Abyad and in the Wadi Khafifah area in the Rustaq and Samad-Wadi Tayin blocks respectively (both described in detail later). Here, a total of 67 sites (including sub-sites) were sampled (40
sites at Wadi Abyad and 27 sites in the Wadi Khafifah area). Site selection and sampling frequency varied depending on specific objectives and the questions pertinent to each locality (discussed later). The results from the February 2011 field season were also used to focus the sampling strategy further during January 2012. Sampling attempted to retrieve at least 8 cores or 2-3 hand samples from each site so that statistically relevant averages could be calculated to provide confidence in the results. Also, it was preferable to collect as fresh samples as possible as weathering can oxidize magnetite to hematite, potentially resulting in magnetic overprinting (Tarling, 1983; Butler, 1992; Tauxe, 2009).

<table>
<thead>
<tr>
<th>Localities</th>
<th>Sites Feb. 2011</th>
<th>Sites Jan. 2012</th>
<th>Total</th>
<th>Field drilled core samples</th>
<th>Hand samples (cores drilled)</th>
<th>Specimens</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wadi Abyad</td>
<td>12</td>
<td>28</td>
<td>40</td>
<td>97</td>
<td>43 (121)</td>
<td>321</td>
</tr>
<tr>
<td>Khafifah</td>
<td>3</td>
<td>24</td>
<td>27</td>
<td>66</td>
<td>27 (61)</td>
<td>220</td>
</tr>
<tr>
<td>Khafifah South</td>
<td>-</td>
<td>4</td>
<td>4</td>
<td>-</td>
<td>5 (21)</td>
<td>46</td>
</tr>
<tr>
<td>Somrah</td>
<td>5</td>
<td>2</td>
<td>7</td>
<td>47</td>
<td>8 (20)</td>
<td>101</td>
</tr>
<tr>
<td>Wadi Nassif</td>
<td>5</td>
<td>-</td>
<td>5</td>
<td>-</td>
<td>8 (42)</td>
<td>46</td>
</tr>
<tr>
<td>Tuf</td>
<td>3</td>
<td>-</td>
<td>3</td>
<td>24</td>
<td>-</td>
<td>43</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td><strong>28</strong></td>
<td><strong>58</strong></td>
<td><strong>86</strong></td>
<td><strong>234</strong></td>
<td><strong>91 (265)</strong></td>
<td><strong>777</strong></td>
</tr>
</tbody>
</table>

Table 2.1: Breakdown of the number of samples collected and specimens prepared from each sampling locality.

Field-drilled cores were collected using a portable petrol-powered rock drill with a water-cooled diamond drill bit (internal diameter of 2.5 cm) using standard palaeomagnetic techniques as given by Tarling (1983), Butler (1992) and Tauxe (2009). Each sampling site can be regarded as representing roughly 1 m$^3$ of rock. The cores were drilled to a minimum depth of 2.5 cm so that at least one standard-size palaeomagnetic specimen (2.2 cm in height) could be recovered. It was preferable, however, to drill to roughly 5.0 cm so that two specimens could be obtained. Core orientations were recorded using both sun and magnetic compasses with a typically variation of approximately 1-2° between each methods. Where available, the calculated sun-compass direction
has been used for the core azimuth, unless the variation between the directions of the magnetic compass and sun-compass were greater than 5°, in which case the magnetic compass direction was favoured.

When drilling could not take place, orientated hand samples (roughly 15 cm x 15 cm x 5 cm) were collected that were then drilled in the laboratory. Hand samples were selected carefully as they are inherently collected from the more fractured portions of an outcrop increasing the likelihood of weathering and issues concerning alteration. The dip and dip direction of a flat planar surface was measured *in situ* and the data recorded then also marked directly onto the rock surface to ensure accurate orientation of the sample back in the laboratory. The hand samples were then removed from the outcrop and labelled for identification.

Magnetic fabric and palaeomagnetic investigations not only require accurately oriented samples but also measurement of the orientation of related structural features (e.g. layering/foliations and lineations). Where possible, structural readings were taken using a compass-clinometer as close to the sampling point as possible and used subsequently in the interpretation of results.

Final specimen preparation took place at Plymouth University. In total, 234 field-drilled cores and 91 hand samples (providing a further 265 cores) were collected during fieldwork. Hand samples were drilled perpendicular to the oriented surface with a pillar drill converted for palaeomagnetic coring using a water swivel and diamond drill bit. The orientation marks made directly onto the hand samples in the field were then used to orientate the cores to their *in situ* coordinates. All core samples were sliced into standard 2.2 cm high palaeomagnetic specimens using a water-cooled, dual-blade rock saw with
non-magnetic bronze diamond-edged blades. At least one specimen was prepared from each core but most provided two and occasionally three individual specimens.

2.6 Laboratory techniques and equipment

A number of different techniques have been employed during the course of this research, both palaeomagnetic and general geological methods (see Figure 2.5 for an outline with details described in the following sections). Palaeomagnetic work has included remanence and susceptibility experiments, thermal and alternating field demagnetization experiments and magnetic fabric determinations, while non-palaeomagnetic work mainly focused on petrological studies concentrating on petrography, mineralogy and micro-structural analysis.

The majority of laboratory analyses were performed at the Plymouth University Palaeomagnetism Laboratory, UK; some analyses were also carried out at the Ludwig-Maximilians-Universität München, Germany and will be discussed later. In the Plymouth laboratory, Helmholtz cages shield the working areas from external magnetic fields such as the Earth’s magnetic field. Outside these cages the Earth’s magnetic field is approximately 46,000 nT while inside the cages this is reduced by roughly 98% to <1,000 nT. An AGICO JR-6A spinner magnetometer has been used for the measurement of remanent (natural, isothermal and anhysteretic) magnetizations. Samples were demagnetized using either thermal or alternating field demagnetization techniques, the former using MMTD1 and MMTD80 thermal demagnetizers manufactured by Magnetic Measurements Ltd (UK) and the latter using an AGICO LDA-3A AF demagnetizer. The LDA-3A also has an AMU-1A anhysteretic magnetizer attachment, which was used to impart anhysteretic
remanences. A Molspin pulse magnetizer was used for isothermal remanent magnetization experiments. Lastly, an AGICO KLY-3 Kappabridge with a CS-3 furnace apparatus was used for the measurement of anisotropy of low-field magnetic susceptibility of samples and the variation of susceptibility with temperature.

Figure 2.5: Flowchart detailing all techniques used during this research from sample collection to full palaeomagnetic analysis.

2.7 Measurement of remanent magnetizations and demagnetization

Rocks containing ferromagnetic minerals can be regarded as palaeocompasses, recording and retaining a memory of the Earth’s ancient geomagnetic field. Palaeomagnetic studies aim to identify the characteristic remanent magnetization (ChRM), which is generally interpreted as the most ancient magnetization of a specimen formed at the time of their formation or during some major tectonic event in the rocks geological past (Butler, 1998;
Magnetic grains can, however, over time, be magnetically realigned, or new grains can grow during metamorphic events. These grains will acquire a new and younger magnetization (magnetization component) that may add to, but not destroy the original ancient magnetization, depending on metamorphic temperatures (Butler, 1998; Tauxe, 2009). The vector sum of these magnetization components is termed the natural remanent magnetization (NRM) and can consist of multiple components obtained over geological time (Butler, 1998; Morris, 2003; Tauxe, 2009).

In order to isolate the ChRM of a rock sample, it is necessary to remove all additional components that combine to produce the NRM. This is achieved by stepwise demagnetization of the sample in order to gradually eliminate any secondary magnetization components that may be present. The ChRM is typically the most stable magnetization component retained by a sample and is, therefore, the most resistant to demagnetization: subsequent magnetizations are generally less stable with low unblocking temperature/coercivity values and are, therefore, easier to demagnetize and remove, while leaving the desired ChRM intact (Butler, 1998; Tauxe, 2009). Demagnetization of a sample can be performed using two different techniques: the first subjects the sample to stepwise-increasing alternating magnetic fields to realign the magnetic domains, effectively randomising the magnetization (Butler, 1998; Tauxe, 2009); the second utilizes the inverse relationship between temperature and relaxation time, using thermal energy to destroy the magnetization (Butler, 1998; Tauxe, 2009). Both demagnetization techniques have been used during this research in order to allow for comparison of the results from the two techniques.

Demagnetization experiments were conducted in the absence of external magnetic fields within shielded areas that neutralise the Earth’s magnetic field.
and prevent the acquisition of new magnetizations in the laboratory (Butler, 1998; Tauxe, 2009). This was facilitated by Helmholtz cages within the laboratory and also by shielding internal to many of the instruments used.

The stepwise demagnetization procedure used during this research subjected the specimens to increasingly higher demagnetization treatments with the magnetic remanence measured after each step. As the alternating field/temperature is increased with each step, the lower stability components are gradually destroyed, thus reducing their contribution to the NRM until only the ChRM remains (Butler, 1998; Morris, 2003; Tauxe, 2009). In total 563 specimens were demagnetized (318 AF and 245 thermal). Magnetizations were measured using an AGICO JR-6A spinner magnetometer (which is capable of accurately measuring magnetizations down to $2 \times 10^{-6}$ A/m), operated using the AGICO Rema6W software. This system measures the magnetization of a specimen in three positions, so as to generate two determinations of each of the x, y and z components of the magnetization relative to the fiducial line on the sample, in order to define the remanent magnetization vector. The in situ geographic coordinates of the specimen were also inputted into the software plus (where appropriate) information regarding the tectonic situation of the specimen (i.e. layering/foliation) so that a full tectonic (tilt) correction for the resulting magnetization component(s) could be constructed.

2.7.1 Alternating field (AF) demagnetization

Alternating field (AF) demagnetization uses the principle that magnetic moments can be affected by an external magnetic field (Butler, 1998; Tauxe, 2009). The procedure involves progressively increasing alternating fields being applied to a specimen in the absence of an external magnetic field.
During the demagnetization procedure the alternating field increases from zero up to a pre-set peak value while the specimen is rotated around two axes in a tumbling device. As the field increases, the magnetizations of grains with coercivities equal to or less than the field strength align themselves to the changing field (Butler, 1998; Tauxe, 2009). After holding at the peak strength for 30-60 seconds, the field is then gradually allowed to decay to zero, effectively randomizing the magnetizations of grains with coercivities less than the peak applied field (Butler, 1998; Tauxe, 2009). Figure 2.6a illustrates the sinusoidal linear decay of the alternating field with time from the peak field $H_{AF}$, whereas Figure 2.6b shows in detail the AF decaying with successive maxima (either peaks or troughs) 1 mT less than the previous (Butler, 1998; Tauxe, 2009). As the specific coercivities of individual grains are passed (as the field decays), the magnetic domains of these grains will become locked in the direction of the field at that moment (either up or down with respect to Figure 2.6b) (Butler, 1998; Tauxe, 2009).

Figure 2.6: Schematic detailing the AF demagnetization procedure: (a) Sinusoidal linear decay of the AF with time from a peak field of $H_{AF}$ to zero, the grey shaded area is enlarged in part (b); (b) Detail of AF decay with successive maxima (either peaks or troughs) 1 mT less than the previous (modified from Butler, 1998).
As the decay of the field continues to zero, half the magnetic moments will point in one direction with the other half pointing in the antiparallel direction, effectively cancelling each other out (Butler, 1998; Tauxe, 2009). The AF demagnetization method randomises the remanence of grains within a specimen but does not physically alter the rock itself (unlike thermal demagnetization); after full AF demagnetization, specimens can still be used for several further analyses such as ApARM (see section 2.8) and isothermal remanent magnetization (IRM) acquisition (see section 2.9.1).

Magnetic particles with coercivities higher than the peak field are left unaffected while the net contribution to the remanence from the grains at or below the peak field is zero (Butler, 1998; Tauxe, 2009). Low stability magnetizations such as VRMs tend to be easily removed because of their low coercive force (Butler, 1998; Tauxe, 2009). The procedure is repeated with increasingly larger peak AF strengths and the magnetization measured after each step, gradually eliminating any low coercivity components that may be present until only a single high coercivity and, therefore, very stable component (i.e. the ChRM) remains (Butler, 1998; Tauxe, 2009).

An AGICO LDA-3A AF demagnetizer (with triple µ-metal shielding) was used during this procedure. The LDA-3A can produce an AF of up to 100 mT and uses a tumbler apparatus to rotate the specimen during demagnetization so that all specimen axes are demagnetized at the same time. After the NRM was measured, specimens (at least three or four from each site after preliminary tests to see if AF demagnetization was an effective method) were demagnetized in a stepwise manner with small increases of maximum AF at first followed by increasingly larger steps (e.g. a typical run included steps of: 5, 10, 15, 20, 25, 30, 40, 50, 60, 70, 80, 90, 100 mT).
2.7.2 Thermal demagnetization

Thermal demagnetization uses increasing thermal energy to progressively destroy the magnetization of a specimen in the absence of an external magnetizing field, such as the Earth’s magnetic field. The procedure is based upon the inverse relationship between relaxation time and temperature: as temperature increases the relaxation time of magnetic grains decreases (Butler, 1998; Tauxe, 2009).

Rock samples are subjected to multiple cycles of heating/cooling in a field-free space, with step-wise increasing peak temperatures used and the magnetization measured after each step (Butler, 1998; Tauxe, 2009). As the unblocking temperature of each ferromagnetic (s.l.) grain in the sample is approached, its relaxation time decreases, eventually becoming superparamagnetic when the unblocking temperature is reached. When the Curie temperatures of specific ferromagnetic (s.l.) grains are passed, they will behave paramagnetically and lose their remanent magnetizations completely (Butler, 1998; Tauxe, 2009). The complete heating/cooling cycle is done in a zero-field environment to prevent the acquisition of new TRMs (Butler, 1998; Tauxe, 2009). As the samples are cooled back below the blocking temperatures of the individual grains, they are once again able to hold a remanent magnetization, but in the absence of a field they acquire a random magnetization and, therefore, no longer contribute towards the net magnetization (Butler, 1998; Tauxe, 2009). The low stability secondary magnetizations with low relaxation times will be the first to be destroyed leaving only the high stability ChRM intact (Butler, 1998; Tauxe, 2009).

Two thermal demagnetizers were used, the MMTD1 and MMTD80 both of which have four layers of µ-metal shielding. Specimens selected for thermal demagnetization (about three or four per site after preliminary demagnetization
runs showed the thermal demagnetization treatment was appropriate) were exposed to increasing temperature steps, which decreased in step-size as the Curie temperature of magnetite (578°C; Tarling & Hrouda, 1993; Morris, 2003) was approached (e.g. a typical thermal run included steps of: 100, 150, 200, 250, 300, 350, 400, 450, 500, 520, 540, 550, 560, 570, 580°C). During each heating/cooling cycle the specimens were heated to the pre-set temperature and held at that temperature for 40 minutes, then cooled back down to room temperature with the aid of a fan. Thermal demagnetization unlike AF demagnetization can cause mineralogical changes within the specimens at high temperatures, resulting in the growth of new, possibly magnetic minerals. Therefore, after each thermal demagnetization step, the bulk susceptibility was also measured using either the AGICO KLY-3 Kappabridge or a Bartington susceptibility meter and the value recorded. Once full demagnetization had taken place, the data were added to the remanence data for analysis.

2.7.3 Data visualisation, principal component analysis (PCA) and getting a direction

During demagnetization the magnetization of the sample is gradually destroyed to reveal the ChRM, if multiple components are recorded, significant changes in the direction of magnetization can be observed (Butler, 1998; Tauxe, 2009). These changes can be visualised using stereographic projections and orthogonal vector (or Zijderveld) plots.

Stereographic projections graphically show the declination and inclination of the remanence components as demagnetization progresses (Figure 2.7a). By convention, solid symbols denote downward directions (i.e. positive inclinations) while open symbols are used for upward directions (i.e. negative inclinations). A drawback in using stereographic projections, however,
is the loss of information on the intensity of magnetization (i.e. the magnitude of the vector) that also changes during the demagnetization process (Figure 2.7b) (Butler, 1998; Tauxe, 2009).

Orthogonal vector plots or Zijderveld diagrams (Zijderveld, 1967), on the other hand, show the evolution of remanence components during demagnetization, graphically showing changes in both direction and intensity simultaneously (Figure 2.8). The data from each demagnetization step are plotted as points on two sets of superimposed axes. The N and E Cartesian components of magnetization are plotted on N–S/E–W axes. The projection of the vertical Cartesian component of magnetization Z is plotted on to either N–S/Up–Down or E–W/Up–Down axes. Successive points are usually joined by straight lines. The angle subtended by each point with the N axis is the
declination. The angle between each point and the horizontal in the vertical plane gives the apparent inclination $I_{app}$, which is related to the true inclination by:

\[
\tan I = \tan I_{app} |\cos D| \quad \text{for the N-S vertical plane}
\]

\[
\tan I = \tan I_{app} |\sin D| \quad \text{for the N-S vertical plane}
\]

The distance of each point from the origin is proportional to the intensity of the component of magnetization plotted on to that plane. A linear segment in the demagnetization path defined by a number of successive points on these plots indicates demagnetization of a single component of magnetization with a constant direction (or conceivably two components of magnetization with identical unblocking temperature or coercivity spectra). The declination and inclination of successively removed components can be easily calculated. When only a single component of magnetization is present the demagnetization path will plot as a straight line towards the origin (Figure 2.8a), however if multiple components are present the demagnetization path is composed of multiple linear segments separated by kinks/bends in the path, each representing a unique magnetization component, with the ChRM typically being the highest coercivity/unblocking temperature component with a trajectory towards the origin (Figure 2.8b). Curved paths between linear demagnetization segments represent partially overlapping coercivity/unblocking temperature spectra of the components while sharp kinks are indicative of two completely separate coercivity/blocking temperature spectra.
Least squares best-fit lines may be calculated for each remanence component using principal component analysis (PCA; Kirschvink, 1980) in order to define their declination, inclination and intensity range (Butler, 1998, Tauxe, 2009) (Figure 2.9). The accuracy of PCA is estimated by the maximum angle of deviation (MAD). Normally a MAD of less than 10° is considered to be statistically valid; any directions picked during PCA with a MAD angle greater than 10° were disregarded from subsequent analysis (Butler, 1998, Tauxe, 2009).

Figure 2.8: Orthogonal vector plots or Zijderveld diagrams (Z-plots) showing the evolution of the magnetization direction and intensity during demagnetization. The distance between each point on the Z-plot is proportional to the intensity loss between each successive demagnetization step. The declination of the magnetization component/s is plotted on the horizontal plane with solid symbols, while the apparent inclination is plotted on the vertical plane with open symbols: (a) Z-plot displaying the demagnetization of a sample with only one magnetization component. Note the linear path towards the origin; (b) Z-plot displaying an example of a sample with two unique magnetization components. Note the two linear paths separated by a sharp bend. Once the low coercivity/unblocking temperature component is removed the ChRM is revealed.
Once PCA had been carried out on all specimens, site means (i.e. of the ChRM) for each sampling site were calculated using Fisher statistics (Fisher, 1953), calculated using AGICO Remasoft software. Site mean magnetization directions were then combined, where appropriate, to produce locality mean directions (i.e. for Wadi Abyad and Wadi Khaffafah). The statistical significance of the calculated mean direction is assessed by defining the dispersion of the data around the mean direction. This is given by the parameters k (a measure of the
dispersion of the vector directions around the calculated mean) and $\alpha_{95}$ (the semi-angle of the cone of 95% confidence about the mean direction).

$$k = \frac{(N-1)}{N-R},$$

and $\alpha_{95}$ is given by:

$$\cos \alpha_{95} = 1 - \frac{N-R}{R} \left[ \left( \frac{1}{0.05} \right)^{\frac{1}{N-1}} - 1 \right],$$

where R is the length of the resultant vectors. $k$ and $\alpha_{95}$ are two key measures in assessing the dispersion (and hence the quality) of palaeomagnetic data. High $k$ values represent minimal dispersion of data around the mean with the rate of increase dramatically reduced once $N$ is about 6 or greater, whereas in large sample-sets $k$ generally reaches a plateau (Tauxe, 2009). The measure of scatter, $\alpha_{95}$, is more dependent on $N$ and it is possible to get artificially low estimates of scatter by increasing the number of samples (Tauxe, 2009). Values of $\alpha_{95} = <15^\circ$ can be regarded as being statistically relevant, however, all $\alpha_{95}$ values obtained in this investigation are well below $10^\circ$ and most below $5^\circ$. Additionally, in order to assess how directions are grouped the randomness of a group of directional data must first be examined. Watson (1956) developed a test for randomness by calculating $R$ (i.e. the length of the resultant vectors). Randomly oriented vectors will give small values of $R$, whereas with increased clustering of directions $R$ will approach $N$ (Tauxe, 2009). Watson (1956) defined the parameter $R_0$ as an estimate of randomness:

$$R_0 = \sqrt{7.815 \frac{N}{3}}.$$

If $R$ is greater than $R_0$ then the hypothesis of randomness can be rejected at the appropriate confidence level.
2.8 Magnetic fabric analysis techniques

Anisotropy of magnetic susceptibility (AMS) is a technique that can be used to study the development of petrofabrics by quantifying the magnetic fabric of rocks that results from the crystallographic alignment and the grain shapes of the assemblage of diamagnetic, paramagnetic or ferromagnetic (s.l.) minerals (Tarling & Hrouda, 1993). Another technique used is the determination of the anisotropy of anhysteretic remanent magnetization (AARM), which results from the preferred alignment and shape of remanence carrying ferromagnetic (s.l.) grains only. The relative speed (especially for AMS) and precision of these techniques make them powerful tools in fabric analysis.

Anisotropy of magnetic susceptibility defines an ellipsoid with three orthogonal axes that correspond to the ellipsoid principal axes (the maximum, intermediate and minimum axes), here referred to as $K_{\text{max}}$, $K_{\text{int}}$ and $K_{\text{min}}$ (Tarling & Hrouda, 1993; Butler, 1998; Tauxe, 2009) (Figure 2.10). Results relate to the shape and alignment of minerals within a specimen and can then be interpreted in terms of the observed macroscopic fabric measured in the field (Tarling & Hrouda, 1993). The degree of anisotropy is controlled by the orientation of the grains within the rock, if $K_{\text{max}} = K_{\text{int}} = K_{\text{min}}$ then the sample is isotropic (i.e. there is no preferred direction of the grains). If $K_{\text{max}} > K_{\text{int}} > K_{\text{min}}$ a triaxial ellipsoid is defined (Tarling & Hrouda, 1993). If however $K_{\text{max}} = K_{\text{int}} > K_{\text{min}}$ then the ellipsoid is oblate (pancake-shaped). Oblate fabrics are generally seen in rocks with planar structures, with $K_{\text{min}}$ perpendicular to the foliation (Tarling & Hrouda, 1993). When $K_{\text{max}} > K_{\text{int}} = K_{\text{min}}$ the ellipsoid is prolate (cigar-shaped), which can reflect the alignment of crystal long axes possibly due to the action of a flow (Tarling & Hrouda, 1993), for example during the emplacement of an intrusion or during its subsequent deformation.
The shape of the ellipsoid, whether prolate, oblate or triaxial defines the parameter $T$, known as the shape factor, which ranges between -1 and 1. Values between $T = 0$ to -1 indicate a prolate fabric with increasing lineation dominance, while values between $T = 0$ to 1 indicate an oblate fabric with an increasing dominance of a foliation towards $T = 1$ (Tarling & Hrouda, 1993). When $T$ is close to zero, a triaxial fabric can be identified where neither a foliation nor lineation dominates (Tarling & Hrouda, 1993).

AMS of all 789 specimens collected in this study was measured using an AGICO KLY-3 Kappabridge run using the AGICO SUSAR software. This susceptibility meter measures anisotropy by spinning the specimen around three axes (achieved by inserting the specimen into the holder in three successive positions). Before each set of spins, the instrument sets to zero after the specimen is lowered into the measurement region in order to zero out the bulk susceptibility and enhance directional differences. The deviatoric
susceptibility is then measured (to a sensitivity of $2 \times 10^{-8}$ [SI]) in 64 positions per revolution. These data are then modelled by a best-fit 2D sinusoidal ($\sin 2\theta$) curve. The 2D curves from each spin axis are then combined with a bulk susceptibility measurement to yield the final best-fit (3D) susceptibility tensor, that may be represented by an ellipsoid specified by the orientation and magnitude of its principal axes ($K_{\text{max}}$, $K_{\text{int}}$ and $K_{\text{min}}$, being the maximum, intermediate and minimum susceptibility axes, respectively).

The AMS ellipsoid usually acts as a proxy for the statistical alignment of all the minerals within the specimen. This includes both diamagnetic and paramagnetic grains, but (if present) the signal will be dominated by the ferromagnetic fraction (Tarling & Hrouda, 1993; Tauxe, 2009). The AGICO Anisoft software and Allmendinger’s Stereonet programme (Allmendinger et al., 2013; Cardozo & Allmendinger, 2013) were then used to visualize the data, with the $K_{\text{max}}$, $K_{\text{int}}$ and $K_{\text{min}}$ axes being represented by blue squares, green triangles and purple circles, respectively (Figure 2.11).

![Figure 2.11: Example of a stereographic projection showing principal anisotropy axes. $K_{\text{max}}$ (blue squares), $K_{\text{int}}$ (green triangles) and $K_{\text{min}}$ (purple circles).](image)
Bulk susceptibility values from AMS determinations may also provide information about mineralogical content and concentrations (Figure 2.12; Tarling & Hrouda, 1993). However, ferromagnetic (s.l.) minerals, such as magnetite and hematite, have considerably higher magnetic susceptibility values than silicates and, therefore, if present, will dominate the bulk susceptibility of a rock (Tarling & Hrouda, 1993).

![Figure 2.12: The contribution of some common ferromagnetic (s.l.) and paramagnetic minerals to the bulk susceptibility of a rock. Both ferromagnetic (s.l.) and paramagnetic minerals will contribute to the overall susceptibility depending upon their concentration within the rock. However, if present even in small amounts, ferromagnetic (s.l.) minerals (i.e. magnetite) will dominate the signal and swamp out the contribution of the paramagnetic fraction (modified from Tarling & Hrouda, 1993).](image)

A caveat that must be addressed when interpreting AMS ellipsoids is that, when particular magnetic minerals and grains sizes are present within a rock,
inversion of the susceptibility axes can take place (Jackson, 1991; Tarling & Hrouda, 1993; Ferré, 2002). Seen predominantly in SD magnetite, AMS ellipsoids may be inverted so the minimum rather than maximum axes are parallel to the long axes of the grains (Jackson, 1991; Tarling & Hrouda, 1993; Ferré, 2002). AARM analyses can however be used to detect the presence of inverse fabrics as the flipping of axes only takes place within an induced magnetizing field (as used during AMS), while AARM techniques rely on remanent magnetizations (Jackson, 1991; Tarling & Hrouda, 1993; Ferré, 2002). AARM techniques can therefore be used as reliability checkers for AMS results, highlighting inverse fabrics, if present, so that sound interpretations of magnetic fabrics can be made. Whereas AMS measures the bulk anisotropy of the whole specimen (the dia-, para- and ferromagnetic fractions), anisotropy of anhysteretic remanent magnetization (AARM) provides information on the preferred alignment of only the remanence-carrying (ferromagnetic) minerals (Tarling and Hrouda, 1993; Butler, 1998; Potter, 2004; Tauxe, 2009; Wack & Gilder, 2012).

Before anisotropy of magnetic remanence (of which AARM is an example) techniques can be conducted the selected specimens must first be fully demagnetized using AF demagnetization. During an AARM experiment, the specimen is simultaneously exposed to an AF and a direct current (biasing) field (DC), producing an anhysteretic remanent magnetization; the AF mobilises the remanence of grains with coercivities less than the peak of the applied field while the DC acts like the Earth’s magnetic field and provides a biasing field that promotes acquisition of the anhysteretic remanence (Potter, 2004). This is repeated multiple times in different directions and the intensity of the acquired remanence measured in three orthogonal directions after each acquisition step.
AARM experiments also give the possibility of focusing on the anisotropy of specific grain-size fractions of ferromagnetic (s.l) particles (such as just fine-grained, single-domain particles) based on their magnetic coercivity (Wack & Gilder, 2012). This is known as the anisotropy of partial anhysteretic remanent magnetization (ApARM). Anhysteretic remanent magnetization experiments are grain-size dependent with fine-grained, single-domain, (for example) magnetite particles susceptible only to higher AF fields while coarse-grained, multi-domain magnetite is sensitive to lower fields. This is related to the coercivity of single-domain (high) and multi-domain (low) grains (Tarling and Hrouda, 1993; Butler, 1998; Potter, 2004; Tauxe, 2009).

ApARM was determined for 133 specimens: 40 were measured in the Plymouth University Palaeomagnetism Laboratory and 99 were measured at the Ludwig-Maximilians-Universität München (LMU), Department of Earth and Environmental Sciences in Munich, Germany on the SushiBar system (see Wack & Gilder (2012) for full details of the SushiBar system). At both laboratories an anhysteretic remanence was imparted using a DC 0.05 mT and an AF window of 60-90 mT and the intensity of magnetization measured in three orthogonal directions. A DC biasing field of 0.05 mT (50 µT) was used as it is close to the Earth’s magnetic field value (46 µT) and the use of a higher field would swamp the magnetometers during measurement. At Plymouth this was achieved by exposing each specimen to an AF of 90 mT and simultaneously a DC of 0.05 mT. The specimen was then tumble demagnetized at 60 mT so that only a 60-90 mT window retained the acquired remanence. An AGICO LDA-3A demagnetizer with AMU-1A anhysteretic magnetizer was used to impart the anhysteretic remanence, using a 12 position magnetizing procedure, which consists of six magnetizing directions and six
antiparallel magnetizing directions. This methodology is considered by AGICO as the most accurate but is also the most time-consuming. The subsequent demagnetization was also carried out using the LDA-3A to tumble the specimen in a demagnetizing AF of 60 mT. The acquired remanence was then measured on an AGICO JR-6A spinner magnetometer using the AGICO AREM software. After each measurement of an even-numbered position (after a set of both parallel and antiparallel magnetizing directions), the specimen was fully demagnetized in an AF of 100 mT to return it back to a baseline remanence intensity. Before visualisation of the results, the raw data were processed using the AGICO AREF software where the in situ geographic coordinates (azimuth and hade) of the specimen were used to calculate the remanence tensors. A slightly different methodology was used whilst conducting the ApARM experiment on the SushiBar system in Munich: the same parameters were used but the additional demagnetization step after anhysteretic remanence acquisition was not required. Instead of imparting an anhysteretic remanence over the full 90 mT coercivity window, the SushiBar can control at which point, within the decaying alternating field, to switch on and off the DC bias so that only the specific range targeted (here 60-90 mT) acquires the remanence (Wack & Gilder, 2012), leaving the rest of the ferromagnetic minerals (with coercivities less than 60 mT) magnetically randomized (Tarling & Hrouda, 1993). A 12 position procedure was also used during ApARM acquisition using the SushiBar, the remanence measured and the anisotropy tensor for each specimen calculated. See Wack & Gilder (2012) for a full description of the setup, features and possible applications of the SushiBar system.

AMS provides a measure of the preferred orientation of all rock-forming minerals in a specimen and the targeted ApARM experiment used here isolates
the anisotropy of just the high coercivity remanence-carrying phases (principally pseudo-single domain and single-domain magnetite) (Tarling and Hrouda, 1993; Butler, 1998; Potter, 2004; Tauxe, 2009). Also and importantly, anisotropy of magnetic remanence experiments are not affected by the single-domain effect of magnetite that can result in inverse AMS fabrics (Butler, 1998; Potter, 2004).

2.9 Further remanence and susceptibility experiments

The techniques employed to reveal the constitutive components of NRM are in a magnetic sense destructive, for once demagnetized the information the rock contained is lost. However, further information about the magnetic properties of rocks, such as their magnetic mineralogy, ferromagnetic (s.l.) mineral concentration and the magnetic domain state (SD, PSD or MD) of grains can be obtained using various laboratory-imparted magnetizations and further rock magnetic experiments (Butler, 1998; Tauxe, 2009).

2.9.1 Isothermal remanent magnetization (IRM) acquisition

An isothermal remanent magnetization (IRM) is acquired from short-term exposure to a strong magnetizing field applied at a constant temperature (room/ambient temperature). All ferromagnetic (s.l.) particles with coercivities equal or less than the applied field will acquire a magnetization. Applying a strong magnetic field to a rock sample will force the ferromagnetic (s.l.) particles with coercivities less than the peak applied field to have magnetic moments aligned along the easy axes that are closest to the field direction, resulting in a net increase of magnetization in that direction (Butler, 1998; Tauxe, 2009). Once a sufficient magnetizing field is reached all grains will have become magnetized along such easy axes and the sample will be magnetically
saturated, reaching a state known as the saturation isothermal remanent magnetization (SIRM) with this value depending upon the concentration of magnetic particles present. For SD magnetite the maximum SIRM is acquired in fields of approximately 300 mT while MD magnetite will saturate in fields of only a few 10’s mT (Morris, 2003). Hematite by contrast requires very strong magnetizing fields between 1.5-5.0 T to reach saturation (a field strength that is not possible with the equipment present at Plymouth) (Morris, 2003). Hence, the strength of IRM and the field required to achieve a SIRM in addition to the shape of the IRM curve can provide useful information related to the ferromagnetic (s.l.) mineralogy of a sample and the concentration and domain state of these minerals.

Using a pulse magnetizer, a direct field of increasing magnitude was applied (at room temperature) along the +Z axis of the specimen imparting the specimen with an IRM. After each step, the intensity of the acquired remanence (along the Z axis) was measured on the AGICO JR-6A. Applied direct fields of 10, 20, 30, 40, 50, 80, 100, 200, 300, 500 and 800 mT were used. Once saturated, along the +Z axis, the specimen was flipped over and the procedure repeated, this time magnetizing along the -Z axis with increasingly larger magnitude (back-field) steps. The same steps as with IRM were used to remagnetize half the magnetic particles along the -Z axis until the net remanence was zero. The field required to achieve this is termed the coercivity of remanence and is controlled by the grain-size/mineral phase of the ferromagnetic (s.l.) particles (Dunlop & Özdemeir, 1997). The data were added to the IRM results, plotted together and the coercivity of remanence calculated. A total of 67 specimens from all six sampling localities underwent IRM acquisition and back-field IRM analysis.
2.9.2 Temperature variation of magnetic susceptibility and Curie temperature (Tc) determinations

The Curie temperature of a specimen is directly controlled by the specific magnetic mineralogy present within the sample. The determination of this value in thermomagnetic susceptibility experiments, therefore, provides useful information for identifying the ferromagnetic (s.l.) mineral or minerals that are carrying the magnetization. These measurements were achieved using the AGICO CS-3 furnace apparatus in combination with the AGICO KLY-3 Kappabridge. Core off cuts from a total of 52 specimens from a range of sites and localities were crushed using a ceramic pestle and mortar (cleaned thoroughly after each use to avoid contamination) to a fine powder. The powdered specimen was placed into a test tube to a depth of approx. 1-1.5 cm and a thermocouple inserted. The furnace was then heated from room temperature to 700°C and then cooled back to room temperature and the susceptibility measured continuously throughout the heating/cooling cycle. The procedure is automated and controlled using the AGICO SUSTE software. Throughout the experiments, argon gas was pumped into the test tube to provide an inert atmosphere to prevent oxidation (chemical changes). The AGICO CUREVAL software was then used to process and graphically display the data for analysis and the Curie temperature was determined using the inverse susceptibility method of Petrovský & Kapička (2006).

2.10 Petrography

Thin sections were prepared from a number of specimens to determine the relationship between the AMS axes and the preferred mineral alignment (i.e. petrofabric) and where field observations suggested that more in-depth thin section analyses would be beneficial. Each thin section was cut in the plane of
the maximum-minimum \((K_{\text{max}}-K_{\text{min}})\) magnetic fabric ellipsoid principal axes, perpendicular to the intermediate axis \(K_{\text{int}}\) (Figure 2.13). In addition to standard petrographic analyses (optical mineralogy), a number of other techniques were used to assess the thin sections, these included point counting and statistical orientation analysis. Point counting (of between 300-500 points per thin section) was conducted in order to calculate the percentage of the minerals present, whilst the angle between mineral long axes and the \(K_{\text{max}}\) axis (as determined by AMS analysis) was also examined in an effort to distinguish between focused and diffuse/variable fabrics.

![Figure 2.13: Orientated thin section preparation method. The trend of the three principal axes \((K_{\text{max}}, K_{\text{int}}, \text{and } K_{\text{min}})\) in specimen coordinates was marked directly onto the core sample. The core was then cut along the plane that includes the maximum-minimum \((K_{\text{max}}-K_{\text{min}})\) axes, perpendicular to the intermediate axis (i.e. with \(K_{\text{int}}\) as the pole to the cut surface). The orientation of the \(K_{\text{max}}\) and \(K_{\text{min}}\) axes were then marked directly on to the thin section so comparison between the AMS axes and the preferred mineral alignment (i.e. petrofabric) could be made.](image-url)
Chapter 3: The Semail ophiolite

3.1 Introduction

The Middle to Late Cretaceous SSZ ophiolites of the eastern Mediterranean and Middle East, such as the Troodos (Cyprus) and Semail (Oman) ophiolites, are by far the best-preserved and most complete ophiolite sequences in the world (e.g. Lippard et al., 1986; Robertson, 2004). The Semail ophiolite is of particular importance for this study as it provides the most extensive exposures of lower crustal rocks of any ophiolite (Lippard et al., 1983; Rioux, 2012; MacLeod et al., 2013). The Troodos and Semail ophiolites lie at the western and eastern end, respectively, of a 3500 km “crescent” of ophiolites that extends in an arc from Cyprus and Turkey, through Syria and Iran to the Semail ophiolite in Oman (the so-called “Croissant Ophiolitique Peri-Arabe” of Ricou, 1971) (Lippard et al., 1983). These ophiolites together form part of a dissected chain of close to 50 ophiolitic rock assemblages collectively known as the Tethyan ophiolites that extend nearly 8,000 km from Italy to India spanning the length of the Alpine-Himalayan mountain belt and trace the suture of the Neotethys Ocean (Lippard et al., 1983) (Figure 3.1). The Tethyan ophiolites can be divided into two groups according to their age: the Jurassic ophiolites of the Alps-Apennines-Carpathians-Dinarides-Hellenides and the Cretaceous ophiolites of Turkey-Syria-Iran-Oman-Himalaya (Lippard et al., 1986) (Figure 3.1). There is also a notable progression from a mid-ocean ridge-type (MOR) geochemical signature in the Italian and western Balkan ophiolites to one showing subduction-related petrological and geochemical features in the eastern Balkan and Cretaceous ophiolites of the Eastern Mediterranean and Middle East (Lippard et al., 1986) (Figure 3.1).
The Tethyan ophiolites formed as a result of the closure of the broadly east-west orientated Neotethys (Lippard et al., 1986; Robertson, 2004; Dilek et al., 2007). The Neotethys formed during diachronous rifting, east to west, of micro-continental blocks off the northern edge of Gondwana in the earliest Triassic, replacing the ancient Paleotethys and separated from it by a chain of rifted micro-continents (e.g. Anatolia, Iran and Tibet) (Lippard et al., 1986; Stampfli, 2000; Stampfli & Borel, 2002; Dilek et al., 2007). Closure of the Neotethys is believed to have started in the Late Jurassic with subduction (beneath Laurasia) early on at the western end of the active spreading ridge, resulting in a zipper-like closure to the east which in turn generated the diachronous nature (Late Jurassic to Late Cretaceous) of the ophiolite chain as closure progressed (Stampfli, 2000; Stampfli & Borel, 2002; Dilek & Flower, 2003). Initiation of intraoceanic subduction within the eastern part of Neotethys, as a result of the north-eastward migration of the Afro-Arabian plate (i.e.
Gondwana) during the late-Early Cretaceous (probably Albian), led to the formation of SSZ-type crust in the overriding plate within the backarc/protoarc-forearc region (Lippard, 1986; Searle & Cox, 1999; Shervais, 2001; Stampfli & Borel, 2002; Dilek & Polat 2008).

3.2 Formation, emplacement and characteristics of the Semail ophiolite

The initiation of a NE-dipping intraoceanic subduction zone off the Arabian passive-margin within the Neotethys during the Albian (between 110-100 Ma), followed by short-lived spreading and crustal accretion above the subduction zone (i.e. suprasubduction zone (SSZ) magmatism) during the Cenomanian (between 100-90 Ma), resulted in the rapid formation of the future Semail ophiolite in the protoarc-forearc region of the overriding plate (Pearce et al., 1981; Lippard et al., 1986; Searle & Cox, 1999; Shervais, 2001; Stampfli & Borel, 2002) (Figure 3.2a). U/Pb zircon geochronology suggests that the oceanic lithosphere represented by the Semail ophiolite formed during the Cenomanian between 96-95 Ma (Tilton et al., 1981; Warren et al., 2005; Rioux et al., 2012, 2013). The continuous nature of the ophiolite stratigraphy, along with its along-strike continuity, are also believed to indicate that the Semail ophiolite formed at a fast-spreading ridge with steady-state magmatism, similar to the present-day East Pacific Rise (EPR) (Pallister & Hopson, 1981; Tilton et al., 1981; MacLeod & Rothery, 1992; Kelemen, 1997; Searle & Cox, 1999; Nicolas et al., 2000). Dating of the associated metamorphic sole rocks beneath the ophiolite (using both U/Pb zircon and Ar-Ar hornblende ages) reveals that detachment of the ophiolite from the spreading centre occurred within several
million years of crust formation (Hacker et al., 1996; Warren et al., 2005; Rioux et al., 2013; Cowan et al., 2014).

Figure 3.2: Formation and emplacement of the Semail ophiolite: (a) Rapid accretion of SSZ-type crust above a NE-dipping intraoceanic subduction zone between 110-90 Ma led to the formation of the (future) Semail ophiolite in the overriding plate around 96-95 Ma; (b) Entrainment of the Arabian continental passive-margin within the subduction zone at approximately 93 Ma halted further spreading and emplacement began; (c) Emplacement of the Semail ophiolite nappe-stack was completed by around 74 Ma after approximately 350-400 km of southwestward transportation; (d) Post-emplacement exhumation and extensional collapse, due to isostatic rebound of the underlying continental margin, broke-up the ophiolite into about a dozen tectonic blocks (from Searle & Cox, 1999).
Then at approximately 93 Ma ago the Arabian continental passive-margin became entrained into and locked the subduction zone, halting further spreading (Lippard et al., 1986; Searle & Cox, 1999; Shervais, 2001) (Figure 3.2b). Following the subduction of the leading edge of the Arabian plate, emplacement of the ophiolite onto the continental margin began, and was completed (after 350-400 km of southwestward transportation onto the Arabian platform) by the end of the Campanian at around 74 Ma (within approximately 22 m.y. of its formation) (Lippard et al., 1986; Searle & Cox, 1999; Shervais, 2001) (Figure 3.2c). The ophiolite was initially obducted as a large coherent thrust sheet, but late-stage (post-emplacement) extensional collapse of the nappe-stack, due to isostatic rebound on the underlying partially subducted continental margin, subsequently broke-up the ophiolite into about a dozen tectonic blocks (Figure 3.2d). (Lippard et al., 1986; Searle & Cox, 1999; Shervais, 2001). Finally, save for some weak Cenozoic compressional tectonics, no further major tectonic events have affected the Semail ophiolite, and it has not yet been incorporated into a large orogenic mountain belt (e.g. Lippard et al., 1986; Nicolas & Boudier 1995).

The geometry and along-strike continuity of the Semail ophiolite (most notably of the sheeted-dyke complex) led Smewing et al. (1984) to suggest that the ophiolite possibly represented a near continuous, approximately NW-SE orientated, outcrop of a palaeo-spreading axis. Further detailed work by MacLeod & Rothery (1992) on the inferred palaeo-ridge preserved in the ophiolite, revealed that the extent of the ophiolite was comparable to the longest between-transform segment of the EPR (Orozco to Clipperton). They also showed that segmentation of the palaeo-ridge axis was again similar to that seen in the EPR between the Orozco and Clipperton transform faults and
revealed that ridge offsets were taken up by second-order axial discontinuities (e.g. overlapping spreading centres), not oceanic transform faults (i.e. first-order axial discontinuities), and that they were centred on zones of high magma supply (MacLeod & Rothery, 1992). These zones (of high magma supply) had previously been identified as pipe-like diapiric structures with vertical asthenospheric flow patterns (i.e. mantle diapirs) that were inferred to be feeding overlying axial magma chambers (Ceuleneer et al., 1988) (Figure 3.3). MacLeod & Rothery (1992) also introduced work by Nicolas et al. (1990) that is discussed in more detail in subsequent literature by A. Nicolas and co-workers (e.g. Nicolas & Boudier, 1995; Nicolas et al., 2000) regarding the origin of the southern massifs (i.e. Nakhl-Rustaq, Sumail and Samad-Wadi Tayin). Structural mapping, specifically the orientation of sheeted dykes, revealed that these blocks preserved a slightly older (approximately 1-3 m.y.) lithosphere with NE-SW trending dykes through which a younger NW-SE ridge system had opened (Nicolas & Boudier, 1995; Nicolas et al., 2000) (Figure 3.3). These results added to work by Reuber (1988) and MacLeod & Rothery (1992) (and references therein, e.g. MacLeod & Reuber, 1990) that had previously suggested that the northern massifs (i.e. Aswad and Fizh) of the ophiolite were part of a large-offset overlapping spreading centre or (and the more favoured interpretation by the authors) a southward-directed propagating ridge structure. The existence of the NW-trending propagator in the central and southern massifs (along with the identification of a southward-directed propagator in the northern massifs) led Boudier et al. (1997) to compare the Semail ophiolite palaeo-ridge system with that the Juan Fernandez microplate system (between the EPR and Pacific-Antarctic Ridge), with possible implications for the cause of
intraoceanic detachment due to potential thrusting within a rotating microplate system (MacLeod & Rothery, 1992; Nicolas et al., 2000).

Figure 3.3: Reassembled geological map of the Semail ophiolite showing preserved ridge-related structures along the inferred palaeo-ridge (from Nicolas et al., 2000). The along-strike continuity of the northern massifs is taken to suggest they preserve a semi-coherent palaeo-ridge formed within a (now) southward-directed propagating ridge, while the central and southern massifs preserve an older NE-SW trending ridge that is cut at a high angle by a younger NW-SE trending propagating ridge system. Note the mapped mantle diapirs located along the palaeo-ridge systems are thought to be located at ridge-segment centres where magma supply was highest and suggest that ridge offsets were taken up by second-ridge axial discontinuities (i.e. overlapping spreading centres). Inset map shows present day configuration of the Semail ophiolite for comparison (modified from e.g. Lippard et al., 1986; Nicolas & Boudier, 2011).
The allochthonous nappe of the ophiolite was thrust over additional structural units during its emplacement onto the Arabian passive margin (e.g. Hopson et al., 1981; Lippard et al., 1986; Robertson & Searle 1990; Warren et al., 2005; Cooper et al., 2014) (Figure 3.4). At the base of this nappe-stack, autochthonous to paraautochthonous pre-Permian basement rocks are overthrust by Permian-Santonian shallow continental shelf carbonates (the Hajar and Sumeini Groups along with the later Aruma group) (Hopson et al., 1981; Lippard et al., 1986; Robertson & Searle 1990; Gray & Gregory; 2003). Moving up through the succession, Permian-Turonian continental slope, rise and basinal sedimentary rocks of the Hawasina Complex and the volcano-sedimentary rocks of Haybi Complex are thrust over the basement and platform sedimentary facies (Hopson et al., 1981; Lippard et al., 1986; Robertson & Searle 1990; Gray & Gregory; 2003). The Semail nappe, bounded at its base by the metamorphic sole that is believed to relate to initial intraoceanic detachment, overrides both these thrust sheets (Hopson et al., 1981; Lippard et al., 1986; Robertson & Searle 1990; Gray & Gregory; 2003; Warren et al., 2005; Cowan et al., 2014). Unconformably overlying all these units is the neo-autochthonous, post-nappe, sedimentary cover sequence of late Campanian-Maastrichtian (the Batinah Complex and Qahlah Fm.) to late Paleocene-Eocene age (the Simsima Fm.) (Hopson et al., 1981; Lippard et al., 1986; Robertson & Searle 1990).
3.3 Stratigraphy of the Semail ophiolite nappe

Lippard et al. (1986) provide the most complete and detailed examination of the stratigraphy of the ophiolite with a review of all the major subdivisions (see also Hopson et al., 1981) (Figure 3.5), which are now briefly described below with specific emphasis given to the lower oceanic crust. A key aim of this research is to systematically analyze several key gabbroic sections within the Semail ophiolite, such as the complete lower crustal (and accompanying upper mantle) sequences exposed along Wadi Abyad (in the Rustaq block) and Wadi Khaffifah (Samad-Wadi Tayin block), as well as the layered gabbro exposures near the village of Somrah (Samail block) and along Wadi Nassif (Samad-Wadi Tayin block). Additionally, limited sampling of a 20 m thick gabbroic sill within the MTZ near the village of Tuf (in the western part of the Samail block) also
took place. These sections provided ideal natural laboratories to investigate the processes by which the lower oceanic crust forms (Figure 3.6). Sampling efforts were exclusively focused within the southern ophiolitic blocks where the best exposures of lower oceanic crustal rocks are found, with each section sampled providing the opportunity to study fundamentally different aspects of the lower crustal sequence.

Figure 3.5: Lithostratigraphic column of the Semail ophiolite (from Lippard et al., 1986). See text for discussion.
The mantle

A thick mantle sequence composed of variably serpentinitised peridotites (depleted harzburgites and dunites) forms the bulk (60-70%) of the obducted nappe and is interpreted to represent the upper part of the sub-oceanic mantle.
(Hopson *et al.*, 1981; Pallister & Hopson, 1981; Lippard *et al.*, 1986; Nicolas *et al.*, 2000). The maximum thickness of the mantle sequence is, however, variable and ranges from approximately 7 km in the Rustaq ophiolitic block to 11 km in the Samad-Wadi Tayin block (Lippard *et al.*, 1986). The lowermost peridotites throughout the ophiolite display a distinct banded appearance and are characterized by intensely mylonitized zones as a result of their proximity to the basal thrust/metamorphic sole that was formed during detachment and subsequent emplacement of the ophiolite (Lippard *et al.*, 1986). Above this intensely mylonitized zone (approximately 300-500 m stratigraphically above the basal thrust) a fabric (generally defined by a mineral alignment) can sometimes be observed within the peridotites, because they have not been fully overprinted by later detachment/emplacement related processes, but generally the mantle peridotites are unfoliated and massive in nature (Figure 3.7a) (Lippard *et al.*, 1986). Where observable, these petrofabrics reveal that high temperature (1000-1200°C) plastic flow dominated beneath the palaeo-ridge with recorded lineations and foliations generally analogous (with the exception of zones of vertical asthenospheric flow) to those measured in the overlying layered gabbros (Figure 3.7b+c), whereas petrofabrics within areas of inferred vertical asthenospheric flow (i.e. mantle diapirs) are seen to be parallel with those recorded in the foliated gabbros, suggesting that a coherent set of accretionary processes were operating on the entire crustal/mantle sequence (Ceuleneer *et al.*, 1988; Nicolas *et al.*, 1996; MacLeod & Yaouancq, 2000).
3.3.2 The crust/mantle transition

The transition between the crust and mantle sequences, often referred to as the Mantle Transition Zone (MTZ), is generally marked by an irregular heterogeneous zone (that varies in thickness from <10 metres to a few hundred metres) of harzburgite and dunite that are mutually intruded by/intruding cumulate ultramafic and gabbroic rocks (Figure 3.8a+b) (Lippard et al., 1986; Nicolas et al., 1996; Korenaga & Kelemen, 1997; MacLeod & Yaouancq, 2000). Thick crust/mantle transition zones are predominately seen overlying mantle diapirs (i.e. zone of high magma supply), whilst thin MTZs are believed to form away from zones of high magma supply in magma poor regions of the palaeo-ridge or where once thick MTZs have subsequently been stretched/transposed.
so as to generate a sharp planar contact between mantle rocks below and layered gabbros above, such as is seen in the Wadi Abyad crustal section (Figure 3.8c) (Lippard et al., 1986; Ceuleneer et al., 1988; MacLeod & Rothery, 1992; Nicolas & Boudier, 1995; Nicolas et al., 1996; MacLeod & Yaouancq, 2000). The petrological Moho is, therefore, taken to be located between the top of the MTZ and the base of the continuous layered gabbro sequence (i.e. >95% layered gabbros) (Figure 3.8d) (Lippard et al., 1986; Nicolas et al., 1996; Korenaga & Kelemen, 1997).

Figure 3.8: The crust/mantle transition zone and Moho: (a) The thin MTZ in Wadi Abyad is characterized by a heterogeneous zone of harzburgite and dunite that are mutually intruded by/intruding cumulate ultramafic and gabbroic rocks (photo credit: M. Meyer); (b) Site WN01 (see Figure 3.9d and section 3.7.4) sampled an intrusion of gabbroic material approximately 20 m below the Moho within the Wadi Nassif MTZ. This zone is dominated by intrusions of Moho-parallel foliated gabbroic rocks into massive mantle peridotites (photo credit: M. Maffione); (c) The Moho (yellow dashed line) in Wadi Abyad is a sharp planar contact that separates almost exclusively gabbroic material above from mantle rocks (here mostly dunite) below (dipping 32° NE) with only a thin (approximately 20 m thick) MTZ between them (photo credit: L. Bibby); (d) The approximately 50 m thick MTZ in Wadi Nassif separates the crustal rocks from the mantle with the Moho (yellow dashed line) placed between the last occurrence of gabbro intruding peridotite and the first occurrence of gabbro being intruded by peridotite. The Moho in Wadi Nassif dips 32/180 (photo credit: M. Maffione).
3.3.3 The oceanic crustal sequence

The mantle and MTZ are overlain by a 4-7 km thick oceanic crustal sequence with associated pelagic sedimentary cover (Hopson et al., 1981; Lippard et al., 1986). This crustal sequence can generally be divided into a lower plutonic series and an upper sheeted dyke complex that passes up into an extrusive sequence of lavas and interbedded sediments, all capped by a <500 m thick sedimentary cover sequence (Hopson et al., 1981; Lippard et al., 1986). The crustal sequence of the Semail ophiolite is accordingly interpreted as representing a complete cross-section through oceanic crust that formed at a sub-marine spreading centre as evidenced by the presence of a sheeted dyke complex: a structural feature indicative of sea-floor spreading (Korenaga & Kelemen, 1997).

3.3.4 The gabbros

Accounting for the majority (by exposure) of the oceanic crust preserved within the ophiolite, the plutonic rocks are some of the best-exposed lower oceanic crustal rocks anywhere in the world. They are predominantly composed of gabbros and olivine-gabbros with intervals of ultramafics (e.g. dunite and wehrlite) that are more common near the base of the section and +/- dioritic rocks outcropping near the upper contact with the sheeted dyke complex (e.g. Pallister & Hopson, 1981; Hopson et al., 1981; Lippard et al., 1986; Boudier et al., 1996; MacLeod & Yaouancq, 2000). Three distinctive igneous facies can be identified within the 1-5 km thick lower crust that can be used to subdivide the plutonic rocks (Nicolas et al., 1996; MacLeod & Yaouancq, 2000). The lower two-thirds of the plutonic series are characterized by Moho-parallel layered gabbros, these pass upwards into a steeply inclined foliated gabbro facies (approximately the upper-third) with the entire sequence capped by a relatively
thin (~200 m) varitextured gabbro facies that passes upwards gradationally through a dyke-rooting zone into the sheeted dyke complex (e.g. Pallister & Hopson, 1981; Lippard et al., 1986; Nicolas et al., 1996; MacLeod & Yaouancq, 2000).

3.3.4.1 The layered gabbros

The lower two-thirds of the plutonic suite is dominated by an approximately 2 km thick (average value from a range of 0.5-3 km), interbedded, layered succession of gabbros and olivine-rich gabbros that are characterised by ubiquitous cumulus layering on a cm to m scale and often show distinctive variations in the modal proportions of the three main constituent minerals (i.e. olivine, clinopyroxene and plagioclase feldspar) within individual layers (Figure 3.9a+b) (e.g. Pallister & Hopson, 1981; Lippard et al., 1986; Nicolas et al., 1996; MacLeod & Yaouancq, 2000). The strong compositional control on layering is defined by clinopyroxene-rich (with ± olivine) bases that grade upwards into plagioclase-rich tops (Figure 3.9c+d) (e.g. Pallister & Hopson, 1981; Smewing, 1981; MacLeod & Yaouancq, 2000). Individual layers are generally laterally extensive and can be traced (when exposure allows) for several hundreds of metres along strike in certain localities (Pallister & Hopson, 1981; Lippard et al., 1986). Late ultramafic cumulate layers of dunite and wehrlite can also be identified and are more commonly found near to the base of the layered gabbros close to the Moho (Pallister & Hopson, 1981; Boudier et al., 1996). The layering and subsequently produced mineral foliation are commonly sub-parallel to the palaeo-Moho; this trait is also mimicked by the fabric of individual layers with the orientation of mineral long-axes aligning parallel with layer contacts (Figure 3.9e) (Boudier et al., 1996; MacLeod & Yaouancq, 2000; Coogan et al.,
Mineral lineations within the layered gabbros are also common and are defined by the long axes of elongate plagioclase feldspar and clinopyroxene (MacLeod & Yaouancq, 2000; Yaouancq & MacLeod, 2000). In a few localities plagioclase crystallization clearly precedes that of clinopyroxene (i.e. typical MORB-like crystallization), however, commonly the crystallization of clinopyroxene and plagioclase tends to be simultaneous (typical of SSZ-type crystallization), indicating the addition of water into mantle source of the accreting Semail crust (Hopson et al., 1981; Pearce et al., 1984; MacLeod & Yaouancq, 2000). The addition of water lowers the solidus temperature in the source, triggering wholesale melting, which results in higher CaO/Al₂O₃ ratios in primary magmas generated above the subduction zone, the increase in Ca and Al consequently shifts the eutectic in favour of plagioclase crystallization (Pearce et al., 1984; Metcalf & Shervais, 2008). There is also evidence for high-temperature hydrothermal alteration of the layered gabbros (and also the overlying foliated gabbros) by the interaction of hot fluids (Gregory & Taylor, 1981; VanTongeren et al., 2008). The petrofabric within the layered gabbros is believed to be of a magmatic origin given the absence of extensive crystal plastic deformation, whilst the fabric itself is defined by the preferential alignment (i.e. shape preferred orientation) of plagioclase and clinopyroxene long axes as well as elongated olivine aggregates that aligned as they grew within a magmatic flow (Pallister & Hopson, 1981; Coogan et al., 2002a; MacLeod & Yaouancq, 2000; Nicolas et al., 2008). The layered gabbros are interpreted to have formed either by successive in situ sill intrusions and subsequent cumulate crystal accumulation or by crystallization at the dyke-gabbro transition within a high-level melt lens and subsequent subsidence; these two theories are the end-members of a
number of accretionary models that are described in full in section 3.4 (Pallister & Hopson, 1981; Boudier et al., 1996; Kelemen et al., 1997; MacLeod & Yaouancq, 2000; Coogan et al., 2002a).

3.3.4.2 The foliated gabbros

The upper-third of the plutonic suite (approximately 0.5-1 km in thickness) is composed of a non-layered, medium to fine-grained, foliated (or laminated) gabbro facies formed of gabbros and +/- occurrences of olivine-gabbros with a sub-vertical (Moho perpendicular) foliation (Figure 3.10a) (e.g. Pallister &
Hopson, 1981; Boudier et al., 1996; MacLeod & Yaouancq, 2000). The foliated gabbros are cumulate in nature and dominated by olivine, clinopyroxene and plagioclase feldspar with +/- amounts of orthopyroxene and hornblende as interstitial phases (e.g. Hopson et al., 1981; Kelemen et al., 1997; MacLeod & Yaouancq, 2000). It has also been reported by MacLeod & Yaouancq (2000), from their own petrological studies, that primary oxides are not present within the foliated (or layered) gabbros: they infer that all oxides within the layered and foliated gabbros are secondary and formed as a result of the serpentinization of olivine grains. In appearance, the foliated gabbros are generally homogenous and display little modal variation but do exhibit an obvious magmatic foliation and lineation that are defined by a very strong preferred alignment of elongate minerals (e.g. plagioclase, clinopyroxene and to some extent olivine aggregates) (Figure 3.10b) (Boudier et al., 1996; Kelemen et al., 1997; MacLeod & Yaouancq, 2000; Yaouancq & MacLeod, 2000; France et al., 2009). The boundary between the foliated gabbros and the layered gabbros occurs through an approximately 200 m thick transitional zone (at the base of the foliated gabbros) in which the clear horizontal layering disappears and the foliation gradually rotates from being sub-horizontal (Moho parallel) to sub-vertical (Moho perpendicular) (Figure 3.10c) (Boudier et al., 1996; MacLeod & Yaouancq, 2000; Yaouancq & MacLeod, 2000). Kelemen et al. (1997), however, did not observe this gradual swing of the foliation from sub-horizontal to sub-vertical at the base of the foliated gabbros but instead report a more abrupt transition separating the two igneous facies. Magmatic folding and boudinaged foliation planes are also prevalent, especially in the transitional zone (Figure 3.10d), with MacLeod & Yaouancq (2000) and Coogan et al. (2002a) both suggesting that folding took place in a crystal-liquid environment due to the lack
of crystal plastic deformation (Boudier et al., 1996). The strong sub-vertical foliation and mineral alignment observed within the foliated gabbros are interpreted as being produced either by the flow of melt upwards towards a high-level melt lens or the subsidence of crystals downward within a crystalline mush (see section 3.4 for full details) (Boudier et al., 1996; Kelemen et al., 1997; MacLeod & Yaouancq, 2000; Coogan et al., 2002a).

Figure 3.10: Foliated gabbros: (a) The foliated gabbros, as seen here in Wadi Abyad, are typically non-layered, medium to fine-grained gabbros (+/- olivine-gabbros) with a strong sub-vertical (Moho perpendicular) foliation (photo credit: C. MacLeod); (b) The foliation (indicated by the -S-) within the foliated gabbros is magmatic in origin and defined by the alignment of elongate plagioclase, clinopyroxene crystals (and to some extent olivine aggregates) (from MacLeod & Yaouancq, 2000); (c+d) An approximately 200 m thick transitional zone between the layered and foliated gabbros is marked by intense magmatic folding. In this zone the sub-horizontal (Moho parallel) foliation typical of the layered gabbros rotates and steepens to become sub-vertical (Moho perpendicular), consistent with the overlying foliated gabbros (photo credits: C. MacLeod; M. Anderson, respectively).
3.3.4.3 The varitextured gabbros and dyke-rooting zone (DRZ)

The uppermost 150-500 m (but typically about 200 m) of the plutonic suite is characterised by a unique gabbroic facies, the varitextured (or isotropic) gabbros, that display extreme structural, textural and compositional variability on a cm to m scale in comparison to the fairly texturally homogeneous underlying layered and foliated gabbros (Figure 3.11a) (MacLeod & Yaouancq, 2000; Coogan et al., 2002a; Nicolas & Boudier, 2011). However, throughout the varitextured gabbro horizon sporadic lenses and screens of more homogenous gabbroic material, with a recognizable foliation (that is parallel to the underlying foliated gabbros), can still be observed (Figure 3.11b) (MacLeod & Yaouancq, 2000). The varitextured gabbros also typically display much more evolved geochemical signatures than the layered and foliated gabbros (although some primitive signatures were also recorded), which suggests the presence of frozen liquids (i.e. trapped melts) at this level of the crust (Kelemen et al., 1997; MacLeod & Yaouancq, 2000; Coogan et al., 2002a). This horizon is believed to be either where cumulate crystal residue from the underlying lower crust accumulates before passing up to form the upper crust (MacLeod & Yaouancq, 2000) or where crystallization takes place before subsidence of a crystalline mush downwards forming the sub-vertical foliated gabbros (see section 3.4 for full details) (e.g. Boudier et al., 1996). The varitextured gabbros are commonly interpreted as the fossilised remains of a melt lens, equivalent to the magma chambers (i.e. the axial magma chambers) imaged in situ at this stratigraphic level by seismic data from the East Pacific Rise (e.g. Sinton & Detrick, 1992; MacLeod & Yaouancq, 2000; Coogan et al., 2002a; Nicolas & Boudier, 2011).

The mineralogy of the varitextured gabbros is dominated by plagioclase and clinopyroxene, with ± amounts of olivine and a dark amphibole (probably hornblende), whilst the presence of primary interstitial oxides (unlike the layered
and foliated gabbros) has also been reported (Yaouancq & MacLeod, 2000; Coogan et al., 2002a). The transition between the foliated gabbros and varitextured gabbros takes places over a very small vertical distance (only a few metres). The boundary is defined by an abrupt disappearance of the sub-vertical foliation, typical of the foliated gabbros, and is replaced by a very weak (if present at all) preferred crystallographic alignment and increasing heterogeneity (MacLeod & Yaouancq, 2000).

![Figure 3.11: Varitextured gabbros:](image)

The boundary between the varitextured gabbros and the overlying sheeted dyke complex, however, is variable and appears to change with levels of plutonic activity in the lower oceanic crust (MacLeod & Rothery, 1992). In areas of ‘normal’ plutonic activity a gradational transition over a vertical distance of between 50-100 m (average of 75 m) from <10% dykes to 100% dykes is most typical, in which dykes are observed to root within the varitextured gabbro
horizon, suggesting that they are fed directly from the melt lens, as is implied for the Wadi Abyad crustal section (Figure 3.12a) (Rothery, 1983; Lippard et al., 1986; MacLeod & Rothery, 1992; MacLeod & Yaouancq, 2000). Through this transitional zone, commonly referred to as the dyke-rooting zone (DRZ), the extreme heterogeneity that characterises the varitextured gabbros gradually decreases and is replaced by a more organised structure of weakly foliated gabbro/micro-gabbro screen (striking parallel to the overlying sheeted dykes) that are increasingly intruded by laterally extensive, but thin, doleritic dykes (again trending comparably with the sheeted dyke complex above) before becoming the dominant lithology and, thus, marking the start of the sheeted dyke complex (Lippard et al., 1986; MacLeod & Rothery, 1992; MacLeod & Yaouancq, 2000). In several localities within the Semail ophiolite (e.g. Wadi Khafifah), however, the base of sheeted dyke complex appears to be truncated by the high-level gabbros, suggesting that the DRZ has been partially removed as a result of continued plutonic activity (i.e. late-stage liquid intrusion) and subsequent assimilation (by stoping), of potentially hydrated material, back into the melt lens (i.e. the high-level gabbros) (Figure 3.12b,c+d) (Pallister, 1981; Rothery, 1983; MacLeod & Rothery, 1992; Coogan, 2003; France et al., 2009). Finally, in some areas of the ophiolite the high-level gabbros (i.e. the foliated and varitextured gabbros) are almost completely absent and the sheeted dykes instead penetrate laterally into (but do not root within) the layered gabbros, suggesting that these dykes originated distally further along strike from an area of high magma supply (i.e. located above mantle diapirs) and intruded into a zone of low magma supply (i.e. second-order axial discontinuities) (Rothery, 1983; MacLeod & Rothery, 1992; Nicolas & Boudier, 1995; Nicolas et al., 2000).
The sheeted dyke complex

The 0.5-2 km thick (generally 1.5 km) sheeted dyke complex of the Semail ophiolite is characterized by >95% parallel/sub-parallel, steeply dipping, doleritic and basaltic dykes with planar and parallel (frequently sharp) contacts between neighbouring dykes (Pallister, 1981; Rothery, 1983; Lippard et al., 1986). Throughout the ophiolite the sheeted dykes overlie the plutonic suite with a variable contact (see previous section), whilst the contact with the overlying extrusive sequence (that the dykes locally feed) is generally an abrupt transition.
over a few 10’s metres when covered by massive sheets flows to a more gradual boundary (of approximately 100 m) when pillow lavas dominate (Pallister, 1981; Lippard et al., 1986). The strike of the sheeted dyke complex generally mimics the trend of the ophiolite itself, with N-S striking dykes in the northern blocks (e.g. Aswad and Fizh) and NW-SE striking dykes in the central and southern ophiolite blocks (e.g. Haylayn to Samad-Wadi Tayin) (Pallister, 1981; Rothery, 1983; Lippard et al., 1986). Dykes located outside of the NW-directed propagator in the central and southern massifs, however, display more N-S strikes that are believed to be related to the slightly older NE-SW trending lithosphere that the propagating ridge opened in (MacLeod & Rothery, 1992; Nicolas & Boudier, 1995; Nicolas et al., 2000). Emplacement of individual dykes, which range from <10 cm to 5 m in thickness (but typically 0.5-1 m), predominately occurred along pre-existing dyke margins and other lines of weakness (i.e. fractures) resulting in some dyke-splitting (as evidenced by many dykes displaying only one chilled margin) to accommodate the intrusion of new dykes (Pallister, 1981; Rothery, 1983; Lippard et al., 1986). The sheeted dykes generally display equigranular medium to fine-grained ophitic, sub-ophitic and intergranular textures with plagioclase, clinopyroxene and Fe-Ti oxides dominating the unaltered/unmetamorphosed mineralogy (Pallister, 1981; Rothery, 1983; Lippard et al., 1986). Hydrothermal metamorphism, however, has replaced most of the primary mineralogy with a greenschist facies assemblage (e.g. quartz, epidote, ± chlorite and ± actinolite) (Rothery, 1983; Lippard et al., 1986).
3.3.6 The extrusive sequence and pelagic sedimentary cover

The upper part of the Semail ophiolite is dominated by a 0.5-2 km thick extrusive sequence that is locally interbedded with deep-water pelagic sediments and interpreted to represent submarine volcanism (and sedimentation) at or close to the Semail palaeo-ridge (e.g. Pearce et al., 1981; Lippard et al., 1986; Godard et al., 2003). On the basis of field geometry, petrology and geochemistry the extrusive sequence can be divided into the three main magmatic episodes as follows (from the stratigraphically lowest to uppermost): the Geotimes unit (or the V1 lavas of Ernewein et al., 1988); the (generally lower) Lasail, (upper) Alley and (uppermost) clinopyroxene-phyric units (collectively called the V2 lavas of Ernewein et al., 1988); and the Salahi unit (or the V3 lavas of Ernewein et al., 1988) (Pearce et al., 1981; Alabaster et al., 1982; Lippard et al., 1986). The V1 lavas represent approximately 60% (e.g. Nicolas et al., 2000) of the total extrusive rocks exposed within the ophiolite and are interpreted to be related to the accretion of the plutonic lower crust and sheeted dyke complex at the palaeo-ridge at around 94-95 Ma (Pearce et al., 1981; Alabaster et al., 1982; Lippard et al., 1986; Ernewein et al., 1988; Hacker et al., 1996). They consist of poorly-vesicular brownish pillow basalts and occasional massive sheet flows with a geochemical signature that relates them to mid-ocean ridge basalts (MORBs) (Godard et al., 2003). The V2 units account for about 35% (e.g. Nicolas et al., 2000) of the extrusive sequence and are thought to represent the progression of volcanism from arc-type (Lasail) to SSZ-type (Alley) magmatism during the early stages of ophiolite formation very soon after the eruption of the V1 lavas (<2 Ma; Hacker et al., 1996) (Boudier et al., 1988; Ishikawa, 2002). The V2 volcanics are grey-green, poorly-vesicular, small pillow lavas and massive sheets flows that are geochemically distinguishable from the V1 lavas by their low-Ti and incompatible element
contents, while also exhibiting lower HFSE/REE and HFSE/LILE ratios and high LILE/REE ratios, which are typical melts derived from a hydrous/depleted MORB source (Godard et al., 2003). The uppermost volcanic unit, the Salahi or V3 lavas, only represents 5% (e.g. Nicolas et al., 2000) of the exposed extrusive rocks and consists of alkaline to transitional within-plate basalts that are thought to have erupted from intraplate seamounts during volcanism related to the early stages of intraoceanic thrusting and obduction of the ophiolite onto the Arabian passive margin at around 79 Ma (Lippard et al., 1986; Warren et al., 2005).

Interbedded within and overlying the extrusive sequence of the Semail ophiolite is an approximately <500 m thick sedimentary cover sequence of metalliferous and pelagic sediments (Fleet & Robertson, 1980; Lippard et al., 1986). The interlava sediments are dominated by umbers and associated ochres with minor occurrences of volcaniclastic silts, lava breccias, laminated calcilutites and siliceous mudstones (Fleet & Robertson, 1980; Lippard et al., 1986). In general, the interlava sediments become thicker and more abundant up through the extrusive sequence as volcanism waned (Lippard et al., 1986). Overlying the interlava sediments are the fine-grained pelagic sediments of the Suhaylah Formation consisting of umbers, radiolarian mudstones, calcilutites, volcaniclastic sediments (siltstones and breccias) and occasional bands of chert (Fleet & Robertson, 1980; Lippard et al., 1986). Capping these sediments and, therefore, the ophiolite nappe are the lava breccias/conglomerates and volcaniclastic silt and sandstones of the Zabyat Formation (Fleet & Robertson, 1980; Lippard et al., 1986).
3.4 Lower oceanic crustal accretion

Formation of oceanic crust takes place at mid-ocean ridges (Vine, 1966), but the processes by which the lower oceanic crust develops are still not well constrained and are difficult to investigate in the modern oceans. Initial theories, pre-1990s, described large long-lived magma chambers beneath spreading centres that encompassed the entire lower oceanic crust and were essentially molten reservoirs that fed the higher crustal levels before crystallizing to form the foliated gabbros (by freezing on to the side walls of the magma chamber from the outside in) and the layered gabbros (by the accumulation of crystals from the base upwards) (Figure 3.13) (Cann, 1974; Pallister & Hopson, 1981; Sinton & Detrick, 1992). However, these concepts have now largely fallen out of favour, primarily as a result of geophysical experiments over fast-spreading ridges, such as the East Pacific Rise, that have demonstrated the presence of laterally extensive (500-1000 m wide) but very thin (approximately 10 m thick) melt lenses beneath the axis at the base of the inferred sheeted dyke complex (Detrick et al., 1987; Sinton & Detrick, 1992; Wilcock et al., 1992; Kent et al., 1990). Such melt lenses, often referred to as the axial magma chamber (AMC) or the high-level melt lens, were subsequently interpreted to be overlying large areas of hot and predominately solid material (i.e. a “crystal mush”) that extended down to the base of the crust (e.g. Crawford et al., 1999; Coogan et al., 2002a). The melt supplying the AMC is produced in the mantle and after initially pooling at the crust-mantle boundary (i.e. the Moho), in a lower melt lens (imaged using seafloor compliance studies¹ and 3D seismics), enters the crust and makes its way through the lower crust to feed the high-level melt lens (i.e. 

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¹ Seafloor compliance studies measure seafloor deformation under ocean surface wave loading to determine the shear velocity values in the oceanic crust (and upper mantle) to produce an ‘image’ of the subsurface (Crawford et al., 1999). Melt lenses within the crust (and upper mantle) are displayed particularly well when using this method as the signal is strongest over low shear velocity zones (Crawford et al., 1999).
AMC) at the base of the sheeted dyke complex (Crawford et al., 1999; Dunn et al., 2000; Singh et al., 2006). Very little, however, is conclusively known about the processes that are taking place beneath the AMC and, consequently, how accretion of the gabbroic lower oceanic crust takes place. However, a more recent seismic reflection survey over the Juan de Fuca Ridge has imaged a melt lens approximately 900 m above the MTZ, suggesting that melt may be distributed throughout the lower oceanic crust (Canales et al., 2009). An understanding of how melt moves through the lower oceanic crust is, therefore, vital to our understanding of how the lower oceanic crust forms.

Figure 3.13: Schematic illustrating the large long-lived magma chamber model for lower oceanic crustal accretion. The entire lower oceanic crust was once believed to essentially be a molten reservoir from which the foliated gabbro formed by freezing on to the side walls of the magma chamber and the layered gabbros formed by accumulation of crystals from the base of the magma chamber upwards (from Sinton & Detrick, 1992).
In addition to geophysical experiments on modern spreading ridges, the easily accessible and extensive exposures of lower crustal rocks that the Semail ophiolite provides have been (and still are) an important natural laboratory for field-testing experimental results/theories and research into lower oceanic crustal accretionary processes for well over 30 years (e.g. Browning, 1984; Sinton & Detrick, 1992; Boudier et al., 1996; Kelemen et al., 1997; Korenaga & Kelemen, 1997; MacLeod & Yaouancq, 2000; Garrido et al., 2001; Coogan et al., 2002a, 2002b; VanTongeren et al., 2008; Boudier & Nicolas, 2011; Nicolas & Boudier, 2011 amongst others). Many of the aforementioned studies have focused on the petrological characteristics of the inferred “crystal mush” zone (beneath the AMC) and found it lacked extensive crystal plastic deformation and was almost exclusively cumulate in nature, suggesting it formed in a semi-magmatic/crystal-liquid environment, whereas Macleod & Yaouancq (2000) were the first to recognise the varitextured gabbros, at top of the lower crustal sequence, as a fossil magma chamber, equivalent to the AMC imaged in geophysical experiments over active spreading ridges (Hopson et al., 1981; Pallister & Hopson, 1981; Sinton & Detrick, 1992; Kelemen et al., 1997; MacLeod & Yaouancq, 2000; Coogan et al., 2002a; Nicolas & Boudier, 2011). However, it is apparent from a review of the current literature that, at present, no clear consensus has been reached with regards to the fundamental processes involved in the formation of the gabbroic lower crust. Models of oceanic crustal accretion should be able to reproduce both petrological and geological observations from ophiolites and geophysical observation from active spreading ridges (Coogan et al., 2002a; Maclennan et al., 2004). Current models for the formation of the lower crust at fast-spreading ridges focus on two end-member models: the “gabbro glacier” model (e.g. Quick & Denlinger, 1993)
and the “multiple-sills” model (e.g. Kelemen et al., 1997), with a plethora of intermediate and hybrid models in-between (e.g. Boudier et al., 1996).

Additionally, MacLeod & Yaouancq (2000) have more recently proposed a new model that focuses on the transportation of melt through the crust, the so-called “melt migration” model. All models, however, are heavily influenced by the geological relationships observed in the Semail ophiolite.

3.4.1 The “gabbro glacier” model

The gabbro glacier model predicts that all melt pools within the high-level melt lens (located at the base of the sheeted dyke complex), partially crystallises (along the base of the melt lens) and subsides, by ductile flow, as a semi-solid crystalline mush downwards and away from the ridge axis to form the foliated and layered gabbros below (Figure 3.14) (e.g. Henstock et al., 1993; Nicolas et al., 1993; Phipps Morgan & Chen, 1993; Quick & Denlinger, 1993; Boudier et al., 1996; Nicolas et al., 2009). As it descends from the base of the AMC the subsiding crystalline mush rotates from being sub-horizontal (as a result of accumulation at the base of the melt lens) to become sub-parallel with the overlying sheeted dyke complex, thus, producing the sub-vertical foliation observed within the foliated gabbros; the foliation, therefore, reflects the trajectory of the descending crystal mush (e.g. Henstock et al., 1993; Phipps Morgan & Chen, 1993; Quick & Denlinger, 1993; Boudier et al., 1996; Nicolas et al., 2009). The underlying layered gabbros are then formed as the descending crystal mush levels out (i.e. to be Moho parallel) as it is stretched and drifts away from the ridge axis by forced ridge spreading driven by flow within the underlying mantle (Quick & Denlinger, 1993; Boudier et al., 1996; Nicolas et al., 2009).
How the modal layering typical of layered gabbros is produced at depth within a homogenous crystalline mush, however, is not well constrained but it is tentatively suggested (by e.g. Boudier et al., 1996) that the modal layering is initially generated within the AMC (as a result of crystal settling at the base of the melt lens) and somehow preserved during subsidence (Kelemen et al., 1997). However, this implies that there is no large-scale rigid body rotation during progression of the mush downwards to the base of the crust and that the sub-vertical foliation within the foliated gabbros is associated with overall horizontal shortening, whilst the sub-horizontal foliation of the layered gabbros is related to vertical shortening towards the base, which seems counter-intuitive, given that the sheeted dykes above the melt lens represent horizontal extension.

Figure 3.14: The gabbro glacier model of lower oceanic crustal accretion. The lower oceanic crust is predicted to form from a single high-level melt lens at the base of the sheeted dyke complex, followed by ductile flow of a semi-solid crystalline mush downwards and away from the ridge generating the rest of the lower oceanic crust (after Quick & Denlinger, 1993).
Many aspects of the gabbro glacier model are, however, (inferred to be) supported by petrological, geochemical, geophysical and thermomechanical evidence. Geophysical experiments over fast-spreading ridges have proven the existence of a single high-level melt lens at the base of the sheeted dyke complex that overlies an area of semi-solid material (i.e. a crystal mush) (e.g. Detrick et al., 1987). However, as previously discussed further studies have revealed melt lenses at the base of the crust (e.g. Crawford et al., 1999) and possibility within the lower crust between the AMC and Moho (e.g. Canales et al., 2009). Petrological evidence reveals that bulk whole rock values are similar throughout the lower crust, suggesting a fairly homogeneous composition to the gabbros and, therefore, a common petrogenesis in an open-system, whilst geochemical data has shown that primitive melts have crystallized within the AMC, which is taken to indicate that little or no crystallization/differentiation of primary mantle melt occurs en route to the AMC within the lower crustal levels (Coogan et al., 2002a, 2002b). Finally, thermomechanical models of ridge axes also indicate that it is difficult to remove sufficient heat by shallow ‘on-axis’ hydrothermal circulation alone and would, therefore, require significant levels of deep ‘off-axis’ hydrothermal circulation into the lower levels of the oceanic crust to allow in situ crystallization to take place (Chen, 2001; Coogan et al., 2002a, 2002b). The degree of deep ‘off-axis’ hydrothermal circulation within the oceanic crust is still widely debated, suggesting that the AMC must play a significant role in the removal of heat from the lower crustal system (via extraction from the high-level melt lens) so as to promote crystallization, and must, therefore, supply the majority of material from which the lower oceanic crust forms (Chen, 2001; Coogan et al., 2002a, 2002b).
3.4.2 The multiple-sills model

Whereas the gabbro glacier model involves crystallization within a single high-level melt lens, the multiple-sills model envisages crystallization taking place throughout the lower crust by the ‘on-axis’ intrusion of many sub-horizontal sill-like lenses so that most of the crust crystallizes in situ with little or no requirement for significant vertical transport of crystals (Figure 3.15) (e.g. Kelemen et al., 1997; Korenaga & Kelemen, 1997; Garrido et al., 2001). This model is based on the observed geochemical differences (e.g. of Mg# and REE abundances) between the upper (i.e. varitextured and foliated) and lower (i.e. layered) gabbros that are believed to be the result of a two-stage crystallization process (e.g. Kelemen et al., 1997; Korenaga & Kelemen, 1997; MacLeod & Yaouancq, 2000; VanTongeren et al., 2008).

Figure 3.15: The multiple-sills model of lower oceanic crustal accretion. Formation of the layered gabbros takes place by direct emplacement of many sub-horizontal sill-like melt lenses into the lower crust so that the entire layered series is formed in situ by multiple intrusive episodes without need of significant vertical transportation of crystallization products. Note the upper gabbros are predicted to form within a mini gabbro glacier-type system in order to produce the observed sub-vertical foliation (after Keleman et al., 1997).
The layered gabbros are interpreted to have formed by the emplacement of mantle derived melts directly into the lower crust that then partially crystallize/fractionate out between 10-50% of their liquid mass (e.g. Kelemen et al., 1997). Individual layers within the layered gabbros are characterised by sharp lower and upper contacts with ubiquitous modal banding, defined by olivine-rich bases that grade upwards into plagioclase-rich tops, that is taken to represent crystal settling and accumulation at the base of sill-like melt lenses (e.g. Kelemen et al., 1997). The cyclic repetition of these sill-like cooling units, that are typical of the layered gabbro sequences of many ophiolites (such as the Semail, Troodos and Annieopsquitch ophiolites), provides convincing field evidence for in situ sill emplacement (Kelemen et al, 1997; Korenaga & Kelemen, 1997; Lissenberg et al., 2004). Gabbroic sills within the crust-mantle transitional zone, described in detail by Korenaga & Kelemen (1997), are found to be compositionally and texturally similar to the overlying layered gabbros, suggesting they possibly share a common petrogenesis as well as providing further evidence supporting in situ crystallization within the lowermost crust (see also Kelemen et al., 1997). In addition, as previously mentioned, geophysical studies above active spreading ridges have not only imaged melt lenses at the base of the sheeted dyke complex but also within the lower crust and at the MTZ (e.g. Detrick et al., 1987; Crawford et al., 1999; Canales et al., 2009). In situ fractional crystallization within the MTZ and of the lower gabbros, however, requires mantle-derived melts to be temporarily “trapped” within the lower crust (including the MTZ) and cooled sufficiently to promote crystallization. It is suggested by Kelemen et al. (1997) that in situ emplacement is achieved by melts ascending through the lower crust (to feed sills) by porous flow pooling beneath permeability barriers when porosity and permeability (that reduces with
vertical distance) drops enough to block further ascent. Deep ‘off-axis’
hydrothermal circulation down to Moho depths is then required to remove
enough latent heat from the lower crust to allow in situ crystallization to take
place (Kelemen et al., 1997; Chen, 2001; Cherkaoui et al., 2003).

The residual melt remaining after formation on the lower gabbros is
subsequently ejected (by hydrostatic fracturing) and rises to feed the uppermost
melt lens (i.e. the AMC at the base of the sheeted dyke complex) where it
crystallises and subsides within a mini gabbro glacier-type system to produce
the sub-vertical foliation of the foliated gabbros, any remaining melt then
freezes to form the varitextured gabbros or continues to ascend to form the
dykes and lavas (e.g. Kelemen et al., 1997). The upper gabbros, dykes and
lavas display much more evolved geochemical signatures with respect to the
lower gabbros, whilst the dykes and lavas are again more evolved than the
upper gabbros, suggesting a progression of fractional crystallization from the
base of the crust upwards (e.g. Kelemen et al., 1997; MacLeod & Yaouancq,
2000). However, some primitive signatures have been recorded within the
highest level gabbros suggesting that some mantle-derived melts must make it
all the way to the AMC without undergoing any crystallization en route (e.g.
Kelemen et al., 1997; MacLeod & Yaouancq, 2000; Coogan et al., 2002a,
2002b).

3.4.3 The dual-feeding model

The multiple sills model developed by Kelemen et al. (1997) was slightly
preceded by the dual-feeding model of Boudier et al. (1996) and Chenevez &
Nicolas (1997) that has subsequently been referred to in the literature as the
“hybrid” model because it amalgamates the fundamental principles of the two
end-member models described above (Figure 3.16). This dual accretionary model involves a combination of both downward ductile flow of a crystalline mush from the AMC and sill intrusion within the lower crust (Boudier et al., 1996; Chenevez & Nicolas, 1997).

Figure 3.16: The dual-feeding model of lower oceanic crustal accretion. A hybrid model that combines the fundamental principals of the two end-member models. Both downward ductile flow from a high-level melt lens at the base of the sheeted dyke complex and in situ sill emplacement directly into the lower crust play a part in forming the gabbroic sequence. The bulk of crystallization is still predicted to take place in the AMC with subsidence of this material forming the sub-vertical foliation seen in the upper gabbros, in situ intrusion of sill-like melt lenses into the lower crust then accentuating the sub-horizontal foliation generated within the crystalline mush as it is forced away from the ridge axis (after Boudier et al., 1996; Chenevez & Nicolas, 1997; figure from Coogan et al., 2002a).

In detail, the majority of crystallization is once again inferred to take place within the high-level melt lens at the base of the sheeted dyke complex where
melt (directly fed from a melt lens at the MTZ) partially crystallizes and then subsides as a crystalline mush to produce the sub-vertical upper gabbros (as described in the gabbro glacier model) as well as the bulk of the lower gabbros (Boudier et al., 1996; Chenevez & Nicolas, 1997). The lower gabbros (that are predominately the product of the subsiding crystal mush) are then intruded by multiple sills (again fed from melt pooling at the MTZ) accentuating the sub-horizontal foliation generated within the crystalline mush as it is forced away from the ridge axis as result of the drag on the gabbroic mush produced by the flow of the underlying mantle (Boudier et al., 1996; Chenevez & Nicolas, 1997). The heterogeneous crust-mantle transition zone is, therefore, the product of sill-like lenses pooling at the base of the crust within mantle material before it ascends into the crust to feed either the high-level melt lens at the top of the gabbroic sequence or sills within the lower gabbros (Boudier et al., 1996; Chenevez & Nicolas, 1997)

3.4.4 The melt migration model

In an alternative model developed by MacLeod & Yaouancq (2000), the so-called “melt migration” model, the steep sub-vertical foliation observed within the upper gabbros is considered to be the result of the upward flow of melt feeding the high-level melt lens (i.e. the AMC) rather than the subsidence downwards of a crystalline mush as predicted by both the gabbro glacier and multiple sills models (Figure 3.17) (e.g. MacLeod & Rothery, 1992; Quick & Denlinger, 1993; Kelemen et al., 1997).
In this model, therefore, the sub-vertical foliated gabbros are interpreted to have formed in situ, directly beneath the AMC, by partial crystallization along the margins of a focused conduit of ascending melt rising to replenish the high-level melt lens (Figure 3.18) (MacLeod & Rothery, 1992; MacLeod & Yaouancq, 2000). As the partially crystallized mush moves ‘off-axis’ (due to spreading) the supply of melt is reduced (due to increased distance from the main melt conduit) leading to further cooling that results in crystals becoming locked into the sub-
vertical flow direction (as evidenced by the vertical l-s crystal shape fabrics) 
(see inset 1 in Figure 3.18a) (MacLeod & Yaouancq, 2000). MacLeod & Yaouancq (2000) argue that the case for the foliation within foliated gabbros representing the trajectory of melt upwards, instead of a descending crystalline mush, is supported by several key field relationships observed within many crustal sections of the Semail ophiolite:

Figure 3.18: Schematic detailing the processes inferred to be taking place beneath the melt lens at a fast-spreading ridge and key predictions of the melt migration model of MacLeod & Yaouancq (2000): (a1) As partially crystallized mush moves ‘off-axis’ further cooling results in crystals becoming locked into the sub-vertical flow direction as highlighted by vertical l-s crystal shape fabrics; (a2) After crystallization of the lower and upper gabbros, residual (interstitial) melt rises through the crystalline mush towards the high-level melt at the base of the sheeted dyke complex by intergranular (i.e. grain-supported) porous flow; (a3) No progressive steepening of the gabbro fabric at the varitextured-foliated gabbro transition is observed, implying that the little or no crystal subsidence from the melt lens down into the underlying gabbros is taking place; (a4) Residual melt after in situ crystallization of the layered and foliated gabbros accumulates within the high-level melt lens giving the varitextured gabbros their evolved geochemistry similar in composition to the overlying upper crust; (b) Anticipated magmatic architecture of thre lower oceanic crust as suggested by the melt migration model (from MacLeod & Yaouancq, 2000).
1. At the varitextured-foliated gabbro transition (i.e. at the base of the AMC) no progressive steepening of the gabbro fabric is observed (see inset 3 in Figure 3.18a) (e.g. Lippard et al., 1986; MacLeod & Yaouancq, 2000). The gabbro glacier model implies this zone should be characterised by a horizontal fabric (formed as a result of cumulus layering at the base of the AMC) that progressively steepens as it descends into the lower crust. However, what is seen at this critical level is the steep sub-vertical fabric juxtaposed right up against the contact, suggesting that the fabric within the foliated gabbros formed prior to the contact zone and is, therefore, a primary crystallization feature (MacLeod & Yaouancq, 2000).

2. The absence of modal layering within the upper (foliated and varitextured) gabbros (MacLeod & Yaouancq, 2000). If the striking compositional layering of the lower gabbros is the result of cumulus crystallization along the base of the high-level melt lens that is subsequently preserved during subsidence, as the gabbro glacier model proposes (e.g, Boudier et al., 1996), why then is this modal layering not seen within the upper gabbros (that are typically fairly homogenous) but is observed within the lower gabbros (MacLeod & Yaouancq, 2000)?

3. The upward flow of melt and the descent of a crystalline mush would oppose each other (MacLeod & Yaouancq, 2000). In order for all crystallization to take place within the AMC to form a gabbro glacier-type system a constant flow of melt from the underlying mantle would be required to provide sufficient material to keep the system running, whilst a continuous flow downwards of the crystallization products
would be needed to form the lower crust (MacLeod & Yaouancq, 2000). As a result, the melt feeding the high-level melt lens would effectively become blocked due to the mass of the subsiding crystal-laden mush (and vice versa) (MacLeod & Yaouancq, 2000).

The accretion of the underlying lower (layered) gabbros is also predicted to occur in situ according to the melt migration model by the intrusion of thin laterally extensive sill-like lenses throughout the lower part of the lower crust in broad agreement with the multiple sills model of Kelemen et al. (1997) (MacLeod & Yaouancq, 2000). MacLeod & Yaouancq (2000), however, favour ‘off-axis’ intrusion of melt extending away from the spreading ridge axis either side of the focused ‘on-axis’ conduit of ascending melt; injection of melt is, therefore, inferred to take place within a largely solid and cooler ‘off-axis’ zone that promotes crystallization due to more effective heat extraction (Figure 3.18b). Any melt remaining (i.e. interstitial melt) after partial crystallization of the lower and upper gabbros (excluding the varitextured gabbros that are inferred to represent the AMC) then rises inwards and upwards, just ‘off-axis’, through the crystalline mush towards the high-level melts at the base of the sheeted dyke complex by intergranular (i.e. grain-supported) porous flow (see inset 2 in Figure 3.18a) (MacLeod & Yaouancq, 2000). As previously mentioned, the varitextured gabbros are considered by MacLeod & Yaouancq (2000) and others (e.g. Coogan et al., 2002a; Nicolas & Boudier, 2011) to represent a fossilised melt lens (i.e. high-level melt lens or AMC) where melt fed from the underlying lower crust and/or the mantle accumulates. In the melt migration model this melt is believed to be dominated by the residue of cumulate crystallization within the underlying lower crust and is similar in composition to
the overlying upper crust (i.e. it is evolved) (see inset 4 in Figure 3.18a). However, occasional primitive melt signatures recorded within the varitextured gabbros suggests that melt direct from the mantle can still reach the AMC during periods of high magma supply (MacLeod & Yaouancq, 2000; MacLeod pers. comm.). Finally, the melt migration model infers that emplacement and crystallization of the lower crust takes place *in situ* directly beneath and slightly ‘off-axis’ of the high-level melt lens, therefore, requiring no downward transportation of crystallization products (as predicted by the gabbro glacier model and to some extent the multiple sills model). However, in order for *in situ* crystallization to be achieved sufficient amounts of latent heat must still be removed from the entire lower crust. As suggested by Kelemen *et al.* (1997) for the multiple sills model, the most effective way to do this is to allow deep ‘off-axis’ hydrothermal circulation down to Moho depths. MacLeod & Yaouancq (2000) agree with Kelemen *et al.* (1997) and also present evidence from thermodynamic modelling (e.g. McCollom & Shock, 1998) and the EPR (e.g. Manning *et al.*, 2000) that documents the penetration of seawater down into the lower crust at temperatures up to 900°C to support this hypothesis (Chen, 2001; Cherkaoui *et al.*, 2003).

### 3.5 Testing lower oceanic accretionary models

The models presented above aim to reproduce petrological, geological and geochemical observations from ophiolites as well as geophysical observations from active spreading ridges. However, each model places a greater emphasis on a particular set of observations and, therefore, predicts fundamentally different geological processes taking place beneath the high-level melt lens that results in specific petrological, geochemical and geophysical
consequences for the lower oceanic crust. Additionally, a critical consideration for any accretionary model is the thermal structure of the lower oceanic crust, which has profound implications for the nature of heat and mass transfer between the Earth's interior and exterior at constructive plate margins (Figure 3.19) (Maclennan et al., 2004).

Figure 3.19: The variation with depth through the lower crust of several key quantifiable parameters as predicted by the gabbro glacier (orange line) and multiple-sills (green dashed line) models (modified from a figure designed and drafted by Roz Coggan). See text for discussion.

Viable methods of lower oceanic crustal accretion must take into account the need to extract the latent heat released by crystallization so as to sufficiently cool the newly forming crust to be fully solidified within approximately 1 km of the ridge axis (Vera et al., 1990; Maclennan et al., 2004). Heat is principally lost from the lower crust either by passive conduction of heat away from the ridge system or (and far more efficiently) by hydrothermal cooling as a result of the penetration of cold fluids (e.g. seawater) into the lower oceanic crust (e.g.
Maclennan et al., 2004). The effectiveness of the hydrothermal circulatory system on cooling the lower crust is, however, dependent upon the depth of penetration (see hydrothermal fluid flux curves in Figure 3.19). In the gabbro glacier model, only shallow 'on-axis' hydrothermal circulation within the upper crust is required to successfully cool the high-level melt lens to enable crystallization to take place and also cool to some extent the upper gabbros, while heat remaining in the lower gabbros would be lost slowly by conduction (Chen, 2001). Large-scale in situ crystallization within the lower crust would, however, be impossible, as conductive cooling alone could not remove sufficient heat (Chen, 2001; VanTongeren et al., 2008). Melt emplaced in situ throughout the lower oceanic crust beneath the high-level melt lens, therefore, requires deep (most likely 'off-axis') hydrothermal fluid circulation down into the lower crust (possibly to the Moho) in order to remove sufficient heat to allow in situ crystallization to take place (Garrido et al., 2001; Cherkaoui et al., 2003; Maclennan et al., 2004; VanTongeren et al., 2008).

Deep hydrothermal circulation would theoretically lead to a near constant cooling rate with depth, whilst shallow hydrothermal circulation would lead to an exponential decrease in the rate of cooling with increasing depth into the lower crust (see latent heat release curves in Figure 3.19) (Garrido et al., 2001; Coogan et al., 2002b; Cherkaoui et al., 2003; Maclennan et al., 2004; VanTongeren et al., 2008). These trends led several workers (e.g. Garrido et al., 2000; Coogan et al., 2002b; VanTongeren et al., 2008) to attempt to use the cooling rates of calcium in olivine and plagioclase in the lower oceanic crust as a proxy for testing accretionary models (see cooling rate curves in Figure 3.19). However, Coogan et al. (2002b) identified a considerable decrease in the cooling rate with depth (consistent with a gabbro glacier-type system), whereas
VanTongeren et al. (2008) found no variation in cooling rate with depth (consistent with the multiple-sills model). VanTongeren et al. (2008) concluded that the disparity between the two studies must be due to the geochemical, geological and architectural differences in the crustal sections sampled (Coogan et al. (2002b) sampled in Wadi Abyad, whereas VanTongeren et al. (2008) collected samples from the Wadi Tayin massif) as the same analytical methods were used. Garrido et al. (2001) and Cherkaoui et al. (2003) (who used numerical modelling), however, also recognised a decrease in the cooling rate with depth at the ridge axis, but their work led them to envisage deep ‘off-axis’ hydrothermal cooling of crustal material outside a narrow conduit of stacked ‘on-axis’ sills (melt lenses) directly beneath the ridge (also see McCollom & Shock, 1998; Manning et al., 2000; Chen, 2001). Once these sills (filled with melt and mush) move ‘off-axis’ hydrothermal fluids cool them, with any sills actually intruded ‘off-axis’ cooling considerably faster (Garrido et al., 2001; Cherkaoui et al., 2003). The observations made by Garrido et al. (2001) and Cherkaoui et al. (2003) on: (i) cooling rates; (ii) deep ‘off-axis’ hydrothermal circulation and; (iii) a narrow conduit of melt, lends support to the group of models that proposes in situ crystallization throughout the lower oceanic crust and particularly the model of MacLeod & Yaouancq (2000) for the formation of the lower gabbros: one of a focused, narrow zone of ascending melt with sills intruding into a mush/solid zone either side and slightly ‘off-axis’ of the ridge making use of the hydrothermal circulation to aid cooling and promote crystallization.

The presence or absence of geochemical trends in whole-rock and mineral chemistry has also been used to distinguish between accretionary models (see bulk Mg# curves in Figure 3.19) (Coogan et al., 2002a; Kelemen et
al., 1997; MacLeod & Yaouancq, 2000). If the majority of crystallization takes place within the high-level melt lens, as predicted by the gabbro glacier model and to some extent the dual-feeding model, then there should be little or no geochemical variation with depth and the lower crust should be of a fairly homogeneous composition (e.g. Coogan et al., 2002a). In contrast, in situ crystallization throughout the lower oceanic crust either by ‘on-axis’ or ‘off-axis’ intrusion (i.e. the multiple sills and melt migration models respectively) would result in a geochemical transition from generally more primitive signatures at the base of the crust to more evolved values within the high-level melt lens (e.g. Kelemen et al., 1997; MacLeod & Yaouancq, 2000). The literature is, therefore, divided and the occurrence of primitive melt signatures within the varitextured gabbros (i.e. the high-level melt lens) also complicates interpretation of geochemical trends and their reliability as a proxy for testing models of crustal accretion.

The variations in hydrothermal fluid flux, latent heat release, cooling rates and whole-rock geochemical compositions have all, therefore, been utilised in an attempt to determine the mechanism of lower oceanic crustal accretion and the feasibility of models by a number of authors but debate still continues. Furthermore, many of these previous studies have endorsed their preferred accretionary model after investigating key lower crustal sections, typically the same sections, of the Semail ophiolite. Little attempt, however, has been made to use magnetic fabric (anisotropy) analyses to quantify the variation in crystalline fabrics in these well-exposed lower crustal sections even though each model predicts the development of very different petrofabrics (both tectonic and magmatic in origin) in various parts of the gabbro section.
For example, crystallization and subsequent subsidence of a semi-solid crystalline mush from a single high-level melt lens in a gabbro glacier-type model would theoretically lead to a near-exponential increase in strain with depth in the lower crust, whilst little or no increase in strain would imply minimal downwards transportation as favoured by the *in situ* crystallization-type models (see strain curves in Figure 3.19) (e.g. Henstock *et al.*, 1993; Phipps Morgan & Chen, 1993). The gabbro glacier model, therefore, predicts that as the partially crystallised material descends from the high-level melt lens, the weight of the overlying crust on the mush would increase, resulting in continuous and ever-increasing strain due to progressive and persistent ductile flow (Nicolas *et al.*, 2009). Material at the base of the lower oceanic crust would, consequently, be more deformed (i.e. exhibit more strain) than material just beneath the high-level melt lens. This increase in strain would potentially result in stronger shape and lattice fabrics (leading to an increase in the degree of preferred crystallographic alignment) with depth into the lower crust that are measureable with magnetic fabric (anisotropy) analyses such as AMS (i.e. anisotropy of magnetic susceptibility) because the magnitude of anisotropy is controlled by the intensity of the fabric.
Additionally, sills emplaced directly into the lower crust might have distinctive microfabrics along their margins resulting from the imbrication (tiling) of crystals during magmatic flow (Figure 3.20a). Hence AMS might provide information relevant to primary magmatic processes in the lower crust. Such fabrics have previously been used to determine emplacement directions of dykes in the Troodos ophiolite (Figure 3.20b) (Staudigel et al., 1992).

This research, therefore, involved systematic sampling at different pseudo-stratigraphic levels in multiple locations (e.g. Wadi Abyad, Wadi Khaffifah, Somrah, Wadi Nassif and Tuf) within the gabbroic sequence (as well as the MTZ and the overlying DRZ) of the Semail ophiolite followed by extensive magnetic fabric (anisotropy) analyses in an attempt to distinguish between potential accretionary models. In addition to using magnetic fabric analyses to investigate crustal accretion, the same samples were used to perform palaeomagnetic analyses, aimed at establishing the nature of magnetic remanences in the Semail lower crustal gabbros in order to quantify the extent of tectonic rotation and/or remagnetization of the southern ophiolitic massifs.

Figure 3.20: Schematic diagrams of the distinctive microfabrics that might form as a result of sill/dyke emplacement: (a) Illustration of the arrangement of magnetic fabric axes that may result from sill injection. The obliquity of fabrics relative to sill margins may be used to infer the magma flow vector; (b) Such imbricated fabrics have previously been used by Staudigel et al. (1992) to determine the emplacement directions of dykes in the Troodos ophiolite.
3.6 Previous palaeomagnetic studies of the Semail ophiolite

Previous palaeomagnetic studies of both extrusive and intrusive rocks in the Semail ophiolite have consistently revealed a large declination anomaly between the northern and southern massifs of the ophiolite (Figure 3.21). Palaeomagnetic directions from the northern massifs typically give E to SE-directed magnetizations with low inclinations, while remanence directions from the southern massifs generally display N to NW-directed magnetizations also with low inclinations.

Figure 3.21: Summary of published palaeomagnetic data from the Semail ophiolite showing the large declination anomaly between the northern and southern massifs (modified from Weiler, 2000; base map modified from e.g. Lippard et al. 1986; Nicolas & Boudier, 2011). Arrows = mean tilt corrected remanence directions. Sources: L = Luyendyk & Day (1982); S = Shelton (1984); T = Thomas et al. (1988); P = Perrin et al. (1994, 2000); F = Feinberg et al. (1999); W = Weiler (2000).
The early palaeomagnetic studies carried out by Luyendyk & Day (1982) on the gabbros and upper peridotites from Wadi Kadir and by Luyendyk et al. (1982) on the sheeted dyke complex near Ibri (within the Samad-Wadi Tayin massif from the southern part of the Semail ophiolite) primarily focused on determining whether or not these crustal units could provide a significant contribution to marine magnetic anomalies. As an additional result, these studies provided detailed information related to the magnetic and palaeomagnetic characteristics of the gabbros and dykes and have, therefore, formed the basis of many subsequent palaeomagnetic studies (see “L” data in Figure 3.21). Luyendyk and Day (1982) observed that the plutonic section and mantle sequence directly underlying it are normally magnetized and give approximately similar remanence directions between N and NW with shallow (positive) inclinations. They established that this remanent magnetization was carried by both primary exsolved magnetite (in the upper gabbros) and secondary magnetite (in the lower gabbros) produced from the serpentinization of olivine at elevated temperatures most probably close to the spreading ridge but possibly (although the less favoured option) during obduction (Luyendyk and Day, 1982). Remanence directions within the overlying sheeted dyke complex also gave comparable results with predominately positive N-directed magnetizations with low inclinations; however, several antipodal, reserved polarity directions were also identified even though the magnetization is believed to have been acquired during the Cretaceous Normal Superchron (114-83 Ma) (Luyendyk et al., 1982; Thomas et al., 1988). The N/NW remanences with positive inclinations are reported to be carried by primary exsolved magnetite, whereas the reversed, antipodal magnetizations are carried by secondary hematite formed as a result of late-stage hydrothermalism
(Luyendyk et al., 1982). These data led Luyendyk and Day (1982) and Luyendyk et al. (1982) to conclude that the southern part of the ophiolite formed at equatorial latitudes at an approximately N-S oriented spreading ridge that was subsequently obducted onto Arabia (whilst Arabia was still joined to Africa) following 10° southward transportation, after which it moved northward with Afro-Arabia with little or no net rotation taking place between formation and emplacement.

Following on from the work of Luyendyk & Day (1982) and Luyendyk et al. (1982) in the southern massifs, Shelton (1984) carried out a palaeomagnetic study on the gabbros (and some dykes) from the northern massifs (see “S” data in Figure 3.21). He found them to have acquired a strong present-day overprint that partially masked their primary magnetization. However, when isolated this primary remanence was directed towards the SE, not the N/NW as found in the southern massifs (Shelton, 1984; Thomas et al., 1988). Results from the northern (e.g. Shelton, 1984) and southern (e.g. Luyendyk & Day, 1982; Luyendyk et al., 1982) massifs were subsequently confirmed by Thomas et al. (1988) following their palaeomagnetic investigation of the metalliferous sediments (e.g. umbers and radiolarites) interbedded within the volcanic sequence and gabbros from across the ophiolite (see “T” data in Figure 3.21). Their investigation also revealed the presence of a NNE-directed magnetization within the umbers and radiolarites of the Fizh (northern domain) and Birquat (southern domain) massifs that, like the dykes from the southern massifs, displayed antipodal directions, whilst a SE-directed remanence from a metamorphosed radiolarite sample from the Birquat massif was also identified (see Figure 17 in Thomas et al., 1988) (Luyendyk et al., 1982; Thomas et al., 1988). These new data, when combined with data from the previous three
studies, allowed Thomas and co-workers to produce a comprehensive summary of palaeomagnetic directions for the entire ophiolite that clearly highlighted the declination anomaly between the northern and southern massifs (Figure 3.21). Thomas et al. (1988), however, considered the NW and SE-directed magnetizations (predominately carried by the gabbros) to be contemporaneous and, therefore, antipodal to each other (acquired during unconfirmed reversals within the Cretaceous Normal Superchron), whilst they considered NNE-directed remanences (with SSW-directed antipodes) recorded in the metalliferous sediments and the N-directed magnetization (with S-directed antipodes) seen in the dykes to represent two further magnetizations (with the NNE-directed magnetization being a remagnetization) acquired later and separately (but prior to emplacement as evidenced by a positive fold test) during coherent rotation of the entire ophiolite. Whole nappe rotation of the ophiolite, as opposed to differential rotation between the northern and southern massifs, was favoured by Thomas et al. (1988) because of the along-strike continuity of the sheeted dyke complex and the fact that NNE-directed magnetizations and NW-SE directions are recorded (although sporadically) in the both the northern (Fizh) and southern (Birquat) massifs. Thomas et al. (1988), therefore, proposed two scenarios to account for the acquisition of the differing remanence directions recorded in the Semail ophiolite. The first predicts continuous magnetization/remagnetization during a large 145° clockwise rotation of the entire ophiolite followed by an anticlockwise rotation of 15° together with Afro-Arabia once emplacement was complete (Thomas et al., 1988). The second, and favoured scenario of the authors, involves two opposing rotations before emplacement and subsequent anticlockwise rotation (of approximately 15°) with Afro-Arabia (Figure 3.22) (Thomas et al., 1988). In
this model, the gabbros and metalliferous sediments (i.e. the umbers) acquire the (now) NW-SE orientated magnetizations first, prior to intraoceanic detachment of the (NE-SW orientated) palaeo-ridge between 95 and 105 Ma (Thomas et al., 1988). Intraoceanic thrusting, at approximately 90 Ma, then results in an anticlockwise rotation of up to 75°, during which the (now) N-S directed dyke magnetization is most likely acquired and after which remagnetization of the majority of the sediments take places giving them their present-day NNE-SSW orientated magnetization (Thomas et al., 1988). Finally, obduction and emplacement of the ophiolite onto the Arabian continental margin at around 75 Ma leads to a 40° clockwise rotation (Thomas et al., 1988).

Figure 3.22: Schematic reconstruction detailing how successive magnetization/remagnetization events during whole nappe rotations led to the recorded remanence directions seen in the Semail ophiolite and how these rotations relate to intra-oceanic thrusting and ophiolite obduction (modified from Thomas et al., 1988). See text for discussion.
In an attempt to further constrain the rotation history of the Semail ophiolite, Perrin et al. (1994) conducted the first palaeomagnetic study on the volcanic sequence (within the Salahi area of the northern domain Hilti massif) because it was hoped that the progression from ridge-axis volcanism (V1) to arc-type volcanism (V2) and finally intraplate seamount volcanism (V3) identified within the volcanic sequence would potential allow a continuous record of the rotation history of the ophiolite through time to be acquired. Perrin et al. (1994) found the V1 and V2 lavas to be exclusively magnetized towards the E/SE with low positive inclinations (V1: 146/20; V2: 106/18), whereas the V3 lavas recorded a N-directed magnetization, again with low and predominately positive inclinations (see “P” data in Figure 3.21).

These results led Perrin et al. (1994) to describe the rotation of the Salahi area (which they took to be representative of the entire ophiolite, thus, disregarding internal differential rotations) as a single clockwise rotation of approximately 150° (analogous with Thomas et al., 1988) between 100-70 Ma, after intraoceanic detachment and thrusting and prior to obduction and emplacement,
around a rotation (Euler) pole located close to the present-day northern tip of the ophiolite (Figure 3.23). This model implies that the V1 and V2 lavas were erupted at the beginning of this rotation event prior to and immediately after intraoceanic detachment and thrusting, whilst the eruption of the V3 lavas took place just before final emplacement (Perrin et al., 1994).

In order to confidently apply the results obtained from the Salahí area to the whole Hilti massif, all the northern massifs and possibly the entire ophiolite, Perrin and co-workers conducted additional palaeomagnetic sampling within the volcanic sequence (see “P” data in Figure 3.21). (Perrin et al., 2000). In this study they focused their sampling efforts on V1 and V2 lavas within the northern massifs but also sampled a single V1 lava flow from the southern domain in the Samad-Wadi Tayin massif and additionally used a V2 lava equivalent radiolarite sample from Thomas et al. (1988) (from the Samad-Wadi Tayin massif) for comparison purposes (Perrin et al., 2000). Palaeomagnetic directions provided by the V2 lavas from the northern massifs are fairly consistent and are taken by Perrin et al. (2000) to indicate that by the time of V2 lava emplacement and magnetization the entire northern part of the ophiolite was acting as a single tectonic unit. However, given the scattering of magnetization directions obtained from the earlier V1 lavas, differential rotations between individual massifs within the northern domain are inferred to have taken place prior to V2 lava emplacement and during eruption of the V1 lavas (Perrin et al., 2000). Additionally, results from the limited sampling within the southern Samad-Wadi Tayin massif revealed an approximately 100° declination disparity between the magnetization directions of southern and northern massifs, with N/NNE-directed magnetizations observed in the southern massif as opposed to predominately SE-directed magnetizations in the northern massifs (Perrin et al., 2000). The
inconsistency between the magnetization directions recorded in the northern and southern parts of the ophiolite is taken by Perrin et al. (2000) to indicate a large rotation between the northern and southern massifs.

Following the work of Perrin et al. (1994, 2000) on predominately volcanic rocks within the northern massifs, and given the paucity of data from the southern massifs, Weiler (2000) attempted to resolve the declination anomaly between the northern and southern massifs by extensively sampling lower crustal layered gabbroic rocks due to their excellent and widespread exposure throughout the entire ophiolite. Magnetization directions observed by Weiler (2000) from gabbros sampled in the southern massifs gave comparable results to those reported by Luyendyk and Day (1982) from the Ibra area, with N/NW-directed remanences, whilst gabbros sampled in the northern massifs gave E/SE-directed magnetizations, consistent with palaeomagnetic directions described by Perrin et al. (1994, 2000) for the V1 and V2 lavas (see “W” data in Figure 3.21). Additionally, Feinberg et al. (1999) conducted palaeomagnetic sampling within the lowermost layered gabbros and upper mantle peridotites of the Samail and Samad-Wadi Tayin massifs, whilst also investigating the sub-ophiolitic autochthonous to parautochthonous pre-Permian basement rocks (collectively termed as the continental metabasites) exposed within the Saih Hatat tectonic window. Palaeomagnetic directions obtained by Feinberg et al. (1999) from the gabbros, peridotites and metabasites are also comparable with previous results from palaeomagnetic investigations in the southern massifs, with almost exclusively NW to NNE-directed magnetizations (see “F” data in Figure 3.21). However, as Luyendyk and Day (1982) first recognised, the remanent magnetization within the lower layered gabbros and upper mantle rocks of the southern massifs is carried predominately by secondary magnetite.
(and is, therefore, a secondary magnetization or remagnetization) produced by the serpen tinization of olivine during hydrothermal alteration either at the ridge (i.e. an early remagnetization as proposed by Weiler, 2000) or as a result of an obduction-related hydrothermal event (i.e. a late remagnetization as proposed by Feinberg et al., 1999). The timing of the N/NW-directed magnetization/remagnetization of the southern massifs and how this relates to the E/SE-directed magnetization of the northern massifs is, therefore, a subject of much debate and has fundamental implications on the rotation history of the Semail ophiolite (e.g. Feinberg et al., 1999; Perrin et al., 2000; Weiler, 2000).

Although significant uncertainty surrounding the timing of magnetization acquisition within the southern massifs has been established, Feinberg et al. (1999) argue that given the well-documented age of ophiolite formation within the Cretaceous Normal Superchron (114-83 Ma) the contemporaneous nature of the NW-SE orientated magnetizations, as proposed by Thomas et al. (1988), should be rejected due to the lack of any evidence suggesting short-lived reversals during this time (Luyendyk & Day, 1982). Furthermore, early remagnetization at the spreading ridge is not favoured because it is assumed by Feinberg et al. (1999) that hydrothermal activity driven by volcanism could not affect the lowermost crust and uppermost mantle. Finally, and most significantly, the palaeomagnetic directions recorded in the metabasites are similar to those measured in the overlying mantle and lowermost crustal rocks of the ophiolitic nappe-stack, which is taken to imply that the ophiolite and underlying metabasites acquired their magnetization at the same time during a single late remagnetization event (Feinberg et al., 1999). In order to account for these findings, Feinberg et al. (1999) developed a model in which the remagnetization of both the ophiolite (from the bottom upwards to at least the
layered gabbros) and underlying continental margin takes place during obduction-related serpentinization triggered by a hydrothermal wave composed of high-temperature fluid expelled from beneath the ophiolite during emplacement (Figure 3.24). This obduction-related remagnetization (resulting in the N/NW-directed magnetizations) is only observed within the southern massifs, whilst the northern massifs are believed to retain an earlier magnetization (i.e. the E/SE-directed magnetizations) possibly associated with intraoceanic detachment and subsequently rotated during intraoceanic thrusting (Feinberg et al., 1999). Feinberg et al. (1999) also noted that the decreasing metamorphic grade between the southern and northern massifs likely resulted in lower temperature and, therefore, less intense hydrothermalism beneath the northern massifs allowing for the retention of an earlier, possibly primary, magnetization.

Figure 3.24: Model for obduction-related remagnetization of the southern massifs. Remagnetization of both the ophiolite (from the bottom up) and underlying continental margin is inferred to have taken place during obduction-related serpentinization that was triggered by a hydrothermal wave composed of high-temperature fluids expelled from beneath the ophiolite during emplacement. This remagnetization event is believed to have only affected the southern massifs due to more intense hydrothermalism beneath the southern portion of the ophiolite than beneath the northern massifs. The northern massifs are, therefore, considered to retain an earlier magnetization (the E/SE-directed remanences), while the southern massifs hold the obduction-related remagnetization (i.e. the N/NW-directed remanences). Note also the decrease in metamorphic grade from south to north that mimics the inferred decrease in hydrothermalism (from Feinberg et al., 1999).
Late obduction-related remagnetization of the only southern massifs, however, is not favoured by Weiler (2000) who highlights a number of issues that contradict the conclusions of Feinberg et al. (1999), these include:

1. Positive fold test. The decrease in the dispersion of site mean magnetization directions when the tilt of gabbros is removed suggests that the magnetization was acquired before emplacement-related tilting occurred, ruling out syn/post-emplacement remagnetization (Weiler, 2000).

2. Pre-emplacement hydrothermalism. The main hydrothermal event has been dated to have occurred prior to 90 Ma and, therefore, before emplacement but likely during intraoceanic detachment (Weiler, 2000).

3. Importance of secondary magnetite within the northern massifs. The magnetization within the gabbros from the Hilti massif is predominately carried by secondary magnetite produced by the serpentinization of olivine (Weiler, 2000).

4. Similar magnetization directions between gabbros and overlying volcanics. The mean magnetization directions of the gabbros and V2 lavas within the Hilti massif (northern domain) and the Samail massif (southern domain) are consistent with each other (Weiler, 2000).

5. Primary and secondary magnetite grains with the gabbros display the same remanence directions. Both primary (magmatic) and secondary (hydrothermal) magnetite grains from throughout the ophiolite give similar magnetization directions, implying a high-temperature hydrothermal event completely remagnetized the primary magnetite.
whilst additionally forming (by the serpentinization of olivine) the secondary magnetite (Weiler, 2000).

6. Unknown affinity of the Saih Hatat window. The high-grade metabasites sampled by Feinberg et al. (1999) within the Saih Hatat window cannot be irrefutably proven to have been beneath the ophiolite during emplacement (Weiler, 2000).

Weiler (2000), therefore, proposes that the remanence carried by the gabbros (in both the primary magmatic magnetite and secondary hydrothermal magnetite) and majority of the overlying upper crustal units (i.e. sheeted dykes, V1 and V2 lavas and metalliferous sediments) throughout the ophiolite is a remagnetization acquired during early high-temperature hydrothermal alteration near to the ridge but shortly after intraoceanic detachment. Furthermore, declination anomalies observed between the magnetization directions within gabbros of the southern massifs are taken to indicate accretion either inside or outside of the 1-3 m.y. younger NW-trending ridge propagator first identified by Nicolas et al. (1990) (Nicolas & Boudier, 1995; Nicolas et al., 2000; Weiler, 2000).

In order, therefore, to account for the disparity between the magnetization directions recorded in the northern massifs from those found in the southern massifs (a difference of approximately 130°), Weiler (2000) developed a tectonic model based upon the geometry of an EPR-type rapidly rotating microplate such as the Juan Fernandez or Easter microplate systems. In his model, Weiler (2000) highlights the similarities between the EPR microplates and the Semail ophiolite (as originally proposed by Boudier et al., 1997) after restoration of the palaeomagnetic declinations for each massif back
to north (Figure 3.25). This back-rotation reveals a NE-SW tending palaeo-ridge composed of the northern massifs and the Balah massif, whilst the southern and central massifs (Haylayn, Rustaq, Samail and Samad-Wadi Tayin), that are believed to have been predominately formed within the younger NW-trending ridge propagator, align NW-SE at a high angle to the orientation of the northern massifs.

Figure 3.25: Palaeomagnetically restored geometry of the Semail ophiolite shows similarities to the Juan Fernandez microplate on the EPR: (a) Back-rotation of individual massif mean magnetization directions to restore them to north results in a palaeo-geometry similar to that of the EPR microplate systems. In this scenario a NE-SW trending palaeo-ridge composed of the northern massifs (120° CCW rotation) and the Balah massif (70° CW rotation) is cut, at a high angle, by a younger NW-trending ridge propagator formed of the central and southern massifs (20° CW rotation); (b) simplified sketch of the Juan Fernandez microplate system (modified from Weiler, 2000).
These results are taken to suggest that spreading along the (older) NE-SW trending ridge must have become unfavourable with subsequent spreading taken-up by a new NW-SE trending ridge (i.e. the ridge propagator) that intersected the older ridge at an oblique angle (Figure 3.26a) (Weiler, 2000). As accretion along the NW-SE trending propagating ridge continued, the older ridge was forced to rotate (clockwise), resulting in the formation of a rotating microplate containing the older ridge segment, similar in geometry to the Juan Fernandez (JF) and Easter (E) microplates along the EPR with the propagating ridge analogous to the Endeavour Deep (JF) and Pito Deep (E) propagators (Figure 3.26b-c; compare with Figure 3.25b) (Boudier et al., 1997; Nicolas et al., 2000; Weiler, 2000). The rotation of the microplate takes place as the pole of the rotation axis (i.e. the Euler pole) of the microplate (located at the tip of the propagating ridge) is forced to migrate around the boundary of the microplate, leading to a clockwise rotation of the older ridge segment by approximately 120° (Figure 3.26b-d) (Weiler, 2000). In contrast, the younger propagating ridge is relatively unrotated, rotating anticlockwise by only 20°, whilst the large clockwise rotation of the older ridge segment brings it almost parallel with the younger propagating ridge within 6 m.y. (assuming a 20° m.y. average rotation rate and a 75 km m.y. average half-spreading rate for the propagating ridge) after rotation began (Figure 3.26d) (Weiler, 2000). However, the rate of rotation can be significantly increased if differential rotations between the northern massifs (i.e. within the older ridge) took place, as initially proposed by Perrin et al. (2000), during emplacement of the V1 lavas as a result of the initiation of intraoceanic detachment (Perrin et al., 1994; Weiler, 2000). Eruption of the V2 lavas followed shortly afterwards, after approximately 40° of further rotation, by which time the northern massifs were acting as a single tectonic unit (Perrin et
al., 1994, 2000; Weiler, 2000). Finally, the eruption of the V3 lavas, approximately 6-7 m.y. later, marks the end of rotation and the beginning of obduction on to the Arabian continental margin, with final emplacement completed shortly after leaving the ophiolite in its present-day geometry (Figure 3.26e) (Perrin et al., 1994, 2000; Weiler, 2000).

![Figure 3.26: Semail ophiolite rotation explained using an EPR-type rapidly rotating microplate model](image)

- (a) Spreading along the (older) NE-SW ridge becomes unfavourable with subsequent spreading taken-up by a new NW-SE trending ridge that intersects the older ridge at a high angle;
- (b-c) Continued accretion along the new NW-SE trending ridge results in rapid clockwise rotation of the older ridge;
- (d) The large rotation (of approximately 120°) of the older ridge eventually brings it almost parallel with the younger propagating ridge within 6 m.y. after the beginning of rotation;
- (e) present-day geometry of the Semail ophiolite after emplacement onto the Arabian continental margin (modified from Weiler, 2000). See text for full discussion.
The palaeomagnetic results and accompanying rotation models reviewed here reveal that many uncertainties regarding the rotation history, the timing of remanent magnetization acquisition and the degree of remagnetization still exist. Therefore, in order to continue to unravel these issues further clarification of nature of the remanent magnetization carried by the lower crustal gabbros of the Semail ophiolite is required, as this will potentially provide insights into the timing of remanence acquisition and the pattern of remagnetization within these lower crustal rocks that may lead to a reconciliation of the contradictory magnetization directions reported in the literature. Palaeomagnetic experiments on samples collected from various localities throughout the southern massifs of the ophiolite were, therefore, conducted in order to determine whether rotations and/or remagnetization of crustal blocks had occurred, either during seafloor spreading or during later emplacement of the ophiolite, and to additionally place the magnetic fabric data into an initial geographic framework.

3.7 Geology of the sampled sections

Wadi Abyad and Wadi Khafifah represent two complete, relatively uncomplicated and representative sections through the lower oceanic crust from the Moho (and underlying mantle) to the DRZ in the case of Wadi Abyad and to the highest-level gabbros in Wadi Khafifah. In addition to sampling throughout the extent of the gabbros to characterise the development and variation of crystalline fabrics through the lower oceanic crust, two transects perpendicular to the strike of the measured foliation were extensively sampled within the upper foliated gabbros directly beneath the varitextured gabbros. The 50 m Wadi Abyad transect consisted of 19 sub-sites (WA18-37) that sampled two broadly different cumulate gabbroic facies identified at the locality: a fine-
grained gabbro and a coarse-grained granular gabbro, whereas the 70 m Wadi Khafifah transect of 15 sub-sites (KF12-27) sampled an equigranular fine-grained gabbro and a coarser-grained olivine-rich gabbro. These two transects provide a detailed comparison across this pseudo-stratigraphic level from two different crustal sections and have been used to produce a large magnetic fabric and petrological dataset of this critical zone directly beneath the high-level melt lens aimed at testing key aspects of several accretionary models. The melt migration model of MacLeod & Yaouancq (2000) predicts that this pseudo-stratigraphic level should show evidence of melt migrating upwards through a crystalline mush feeding the AMC, whereas models that predict subsidence of a crystalline mush forming the upper gabbros (i.e. the gabbro glacier, multiple sills and dual-feeding models) suggest that a generally homogenous mass of gabbro with a uniform sub-vertical fabric would be observed. Distinguishing between these two groups of accretionary models using magnetic anisotropy is potentially achieved by examining in detail the magnetic fabric in relation to the petrofabrics observed in the field and highlighting the variability (if any) of fabric development across this zone.

The excellent exposure of multiple sub-horizontal layers in layered gabbros near the village of Somrah provided an ideal locality to assess in detail the development of fabrics related to the strong compositional layering that is ubiquitous within the layered gabbro sequence of the Semail ophiolite. A layered gabbro cliff section NW of the village was, therefore, extensively sampled to determine the variation in anisotropy throughout individual layers in an effort to determine the mineralogical control on the magnetic fabric and highlight any possible crystal imbrication along the top and bottom of the layer as predicted by the in situ sill emplacement group of accretionary models.
Additionally, crystal settling of a static non-flowing melt within a magma chamber would also result in a characteristic magnetic fabric signature that is also detectable by AMS analysis.

Similar questions relating to the development of layering within the lower crust were also evaluated in Wadi Nassif. Sampling of a gabbroic sill within the well-exposed MTZ also took place to assess how the magnetic fabric compared with the layered gabbros above the Moho. Kelemen et al. (1997) and Maclennan et al. (2004) describe how gabbroic sills within the MTZ and the lower layered gabbros must share a common petrogenesis due to their textural and petrological similarities (Boudier et al., 1996). A series of sites were, therefore, sampled from below the Moho (in the MTZ) up into the lower part of the layered gabbros in order to track the variation of fabric through the crust-mantle boundary and into the lower oceanic crust.

Additionally, limited sampling also took place in Tuf (a 20 m thick gabbroic sill within the MTZ), however, results (both palaeomagnetic and magnetic fabric) from these sites were poor. Therefore, these data, along with a description of the geology and sampling activities that took place, are only presented briefly in Appendix B.

3.7.1 Wadi Abyad

Located around the oasis village of Al Abyad, the lower crustal rocks exposed within Wadi Abyad progress from mantle peridotites that pass up through the Moho into a roughly 2.6 km thick plutonic sequence of layered, foliated and varitextured gabbros that is overlain by a dyke-rooting zone, which itself is capped by sporadic outcrops of sheeted dykes (Figure 3.27) (e.g. MacLeod & Yaouancq, 2000).
The Wadi Abyad section cuts approximately north-south through the eastern part of the Rustaq tectonic block in the central part of the Semail ophiolite and presents a roughly 5 km continuous (gabbroic sequence and
semi-continuous lowermost sheeted dyke complex) cross section through the lower oceanic crust (e.g. MacLeod & Yaouancq, 2000). Located approximately 3 km SW of Khatum (Khatum itself being 25 km south on the Abyad Road off the Oman Route 1 Highway), the Wadi Abyad section is an easily accessible, and considered by many to be a classic ophiolitic oceanic crustal sequence owing to its quality and extent of exposure.

Sampling for this study took place throughout the extent of the exposed lower oceanic crustal section, from: close to the Moho (sites WA09 and WA10); within the lower and mid-level layered gabbros (WA11 and WA12 respectively) and the higher-level layered gabbros (WA04A+B); throughout the foliated gabbros (WA05+06, WA01-03, WA08, WA38, WA39+40, WA07 and the transect WA18-37); from the varitextured gabbros (WA13 and WA15) along with some dykes (WA14 and WA16) that intrude the gabbros at this horizon; and finally the dyke-rooting zone (WA17) (Figure 3.27). Multiple measurements of local structural features (e.g. layering/foliation, lineations and dyke margins) were also made at each sampling site (along with measurements in-between); a representative selection of which are plotted on Figure 3.27 (see map and stereonet inset). The structural data recorded in the Wadi Abyad section reveals the steepening nature of the petrofabric up through the gabbros from being initially Moho parallel (black circle on the stereonet) within the layered gabbros (blue circles) then sub-vertical and perpendicular to the Moho in the foliated gabbros (red circles). A zone of variability at the base of the foliated gabbros (the layered-foliated gabbro boundary) between sites WA05+06 to WA08 can also be picked out where the fabric changes rapidly between sub-horizontal and sub-vertical and is interpreted as the transitional zone described by MacLeod & Yaouancq (2000). In the very north and, therefore, stratigraphically highest...
outcrops of the crustal section, the exposed dykes (both within the varitextured gabbros and the dyke-rooting zone) generally strike NE-SW, with dykes farther to the northwest trending N-NNW. The NE-SW trending dykes may possibly be related to the older NE-SW trending ridge, but the presence of N-NNW striking dykes nearby implies an interaction with the younger NW-SE propagating ridge during the formation of the Wadi Abyad crustal sequence. While, WNW-ESE trending lineations recorded in the layered gabbros suggests a flow of material roughly perpendicular to either the NE-SW orientated ridge or the NW-SE propagating ridge.

3.7.2 Wadi Khafifah

Wadi Khafifah provides an almost complete section through the lower oceanic crust, from mantle peridotites to the uppermost gabbros and the base of the DRZ (Figure 3.28) (e.g. Pallister & Hopson, 1981; Garrido et al., 2001). An approximately 3.7 km thick crustal section is preserved cutting north-south through the central part of the southern most Samad-Wadi Tayin ophiolitic block in the steeply dipping northern limb of the east-west trending Ibra syncline (e.g. Pallister & Hopson, 1981). Previous workers (e.g. Pallister & Hopson, 1981; Garrido et al., 2001; VanTongeren et al., 2008; France et al., 2009; Rioux et al., 2012) have noted the area’s excellent and easily accessible outcrops, simple structural and accretionary history, plus a more MORB-like geochemical signature than is recorded in Wadi Abyad, making it an ideal location to test models of lower oceanic crustal accretion. Located approximately 15 km NW of Ibra on the Oman Route 23 Highway; along an 8 km stretch of the Oman Route 25 Highway from the village of Khafifah southwards, the stratigraphy of the
oceanic crustal section within Wadi Khafifah is similar (except in thickness) to that of Wadi Abyad.

Figure 3.28: Geological map of the Wadi Khafifah plutonic sequence with location of sampling sites (blue stars). Lower and middle gabbros (olive green) include both layered and foliated gabbros, while upper gabbros (yellow and black) incorporate the highest-level foliated gabbros, varitextured gabbros and remnants of the DRZ. Inset map shows location of Wadi Khafifah within the Samad-Wadi Tayin block in the southern part of the Semail ophiolite. Inset stereonet presents representative structural data (i.e. poles to plane of layering/foliation and trend and plunge of lineations) from Wadi Khafifah (map modified from Garrido et al., 2001).
Sampling in Wadi Khafifah included the collection of samples from sub-horizontal layered gabbros at sites KF04-06 and KF08 close to the Moho, whereas KF03 sampled a transitional layered-foliated gabbro facies with a moderately dipping foliation and steep, almost vertical lineation, that again shows the steepening of the gabbro foliation up through the plutonic sequence (Figure 3.28). Sites KF01, KF02 and the transect KF12-27 sampled the upper foliated gabbros, whereas the stratigraphically highest gabbroic units of the Khafifah section were sampled either side of the Ibra Road at sites KF10, KF11 and Khafifah South (Figure 3.28). The Khafifah South section is taken to represent the highest-level foliated gabbros (very close to the AMC) and shows clear evidence of late-stage liquid intrusion into the magmatic system whilst the host-rock (i.e. the foliated gabbros) was still in a semi-rigid state, resulting in the partial removal of the base of the DRZ. As a result, inclusions of doleritic xenoliths were also located within the Khafifah South section, which were thought either to represent dropstones of dyke material stoped from the overlying sheeted-dyke complex or the remains of the assimilated lower DRZ.

Dykes sampled in the Khafifah area (KF07 in the lower gabbros and KF09 in the upper gabbros) strike approximately NW-SE suggesting that the Khafifah crustal section, unlike the Wadi Abyad section, formed within the younger NW-SE orientated propagating ridge system. Lineations within the layered and foliated gabbros support this theory as they are orientated E-W and, thus, approximately perpendicular to orientation of the inferred propagating ridge, implying flow orthogonal to the ridge axis (Figure 3.28).
3.7.3 Somrah

The wonderfully exposed layered gabbro sequence to the NW of the village of Somrah (Samra’) is believed to have formed within the younger NW-SE trending propagating spreading ridge system and is now located in the central part of the Samail tectonic block (Figure 3.29) (e.g. Nicolas & Boudier, 1991; Juteau et al., 2000; Nicolas et al., 2008). The village of Somrah itself is at the end of a 9 km metalled road west off Oman Route 23 Highway (Ibra Road) approximately 40 km south along the highway from Bid Bid.

Figure 3.29: Schematic geological map of Somrah layered gabbro cliff section with location of sampling sites (blue stars). Inset map shows location of Somrah within the Semail block in the southern part of the Semail ophiolite. Inset stereonet presents representative structural data (i.e. poles to plane of layering and trend and plunge of lineations) from Somrah.

Sites SR01-05 sampled four layers within this layered gabbro sequence (Figure 3.30). These layers were designated from the lowermost layer upward as: Layer A (SR01-02), Layer B (SR03), Layer C (SR04) and Layer D (SR05). The sampling strategy employed for each layer involved the collection of a
number of samples from the base and top of each layer and also from within the layer itself. The lowermost layer, Layer A, was sampled at a very fine scale (~5 cm sampling intervals) to provide a semi-continuous coverage of one of the layers, so that a high-resolution dataset could be produced. Similar to the Wadi Abyad and Khafifah layered gabbros, there is a strong compositional control on the layering with clinopyroxene-rich bases (with ± olivine) that grade upwards into plagioclase-rich tops. Generally, the layering/foliation (defined by elongate plagioclase and clinopyroxene crystals) is sub-horizontal; a sub-horizontal mineral lineation, roughly trending W-E, can also be identified, again picked out by the alignment of elongate plagioclase and clinopyroxene crystals. The gabbroic rocks have subsequently been cut by a series of dykes that trend NW-SE (parallel to the trend of the inferred propagating ridge), cutting the gabbro layering at a high angle (~70°). Sites SR06 and SR07 sampled two of these dykes to assess the nature of the preserved remanent component of magnetization to see if it differs from the magnetization direction preserved in the layered gabbros.

Figure 3.30: A sequence of four layered gabbros (left) were sampled from within the layered gabbro sequence near the village of Somrah, sites SR01-05. The four layers were labelled from the lowermost layer upward as: Layer A (SR01-02; blue), Layer B (SR03; green), Layer C (SR04; yellow) and Layer D (SR05; orange). The lowermost layer, Layer A, was sampled at a very fine scale (~5 cm sampling interval) to provide a semi-continuous coverage so that a high-resolution dataset could be produced.
3.7.4 Wadi Nassif

The Wadi Nassif section exposes uppermost mantle peridotites that pass up into a roughly 2 km thick oceanic crustal sequence composed of layered gabbros (~1 km thick) and foliated gabbros (~1 km) (e.g. Boudier et al., 1996; Nicolas et al., 1996) (Figure 3.31). Accessible via an 8 km dirt track north off the Oman Route 23 Highway (Ibra Road) the section cuts north-south through the central part of the Samad-Wadi Tayin ophiolitic block.

![Schematic geological map of the Wadi Nassif lower layered gabbro crustal section with location of sampling sites (blue stars). Inset map shows location of Wadi Nassif within the Samad-Wadi Tayin block in the southern part of the Semail ophiolite. Inset stereonet presents representative structural data (i.e. poles to plane of layering) from Wadi Nassif.](image)

Site WN01 was sampled from within an intrusion of gabbroic material within the MTZ (approximately 20 m below the Moho), which is dominated by

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intrusions of foliated gabbroic material into massive mantle peridotite rocks. Above the Moho, the first 70 m of the gabbro sequence is characterised by a similar transitional zone where sills, dyke-lets and lenses of peridotite material intrude into the gabbros, the quantity and size of these intrusions decrease with increasing distance from the Moho. Site WN02 marks the end of this crustal transitional zone. The layered gabbros of the Wadi Nassif sections, however, display only a weak compositional layering in contrast to other sampled sections, but instead exhibit a very strong and regular Moho-parallel layered foliation defined by elongate clinopyroxene and plagioclase (with ± olivine) crystals. However, some rare outcrops of more compositionally layered material can be observed (WN03). Patches of peridotite can also still be seen 200 m (near WN03) stratigraphically above the Moho, appearing as dykes/lens that cut the gabbro foliation intruding laterally and up-section. These peridotite outcrops represent the final appearance of mantle material within the gabbroic sequence and mark the end of a protracted zone of sporadic peridotite intrusions. Continuing further up-section, the more, layered foliated nature of the lower gabbros continues along with evidence of magmatic folding near site WN04. Finally, site WN05, approximately 800 m stratigraphically above the Moho, sampled a section of a 30 m high cliff section composed of a whole series of steeply dipping (+70°) +50 cm thick layers of layered foliated gabbro that again only showed a weak compositional layering. As can been seen in Figure 3.31, the dip of the layered foliation in Wadi Nassif increases rapidly with increasing distance from the Moho: a feature that is also seen in cross sections constructed by Nicolas et al. (1996) (see Figure 2 cross section number 23 in Nicolas et al., 1996).
Chapter 4: Rock magnetic experiments

4.1 Introduction

The strength of the magnetic fabric detected by anisotropy of magnetic susceptibility (AMS) analysis and the ability of a rock to retain a remanent magnetization are wholly dependent upon the magnetic mineralogy of the rock itself. The types of ferromagnetic (s.l.) minerals present within a sample will determine whether the rock has the capability of retaining a stable remanent magnetization over geological time.

The response of a sample’s remanent magnetization to either thermal or AF demagnetization can help broadly characterise the magnetic minerals present within a sample, but in order to gain further insights into the magnetic mineralogy of the lower oceanic crustal rocks of the Semail ophiolite a number of additional rock magnetic experiments were performed. These included: (i) isothermal remanent magnetization (IRM) acquisition (see Chapter 2, section 2.9.1) to determine the coercivity spectra of ferromagnetic minerals present along with back-field IRM acquisition to determine the coercivity of remanence; and (ii) determination of the variation of magnetic susceptibility with temperature (see Chapter 2, section 2.9.2) in order to determine the Curie temperature(s) of ferromagnetic (s.l.) minerals present.

4.2 Rock magnetic experiments

4.2.1 AF demagnetization and median destructive field determination

Alternating field (AF) demagnetization is used to effectively randomise the remanence of grains with coercivities at or below the peak applied field strength, cancelling their net contribution to the magnetization (Butler, 1998;
Tauxe, 2009). The coercivity, that is, the resistance of a magnetic grain to demagnetization depends upon its grain-size; SD magnetite has a higher coercivity and, therefore, stronger resistance to demagnetization than MD magnetite, with PSD grains falling somewhere in between. Haematite and goethite, however, do not fully demagnetize until fields in excess of 150 mT are reached.

![AF demagnetization curves for a range of magnetite grain sizes of 0.1 mT TRM.](image)

Figure 4.1: AF demagnetization curves for a range of magnetite grain sizes of 0.1 mT TRM. The changing shape of the demagnetization curves reflects an increase in the grain size of magnetite from SD (sigmoidal curve) to MD (exponential curve) (from Dunlop and Özdemir, 1997, after Argyle et al., 1994).

AF demagnetization can, therefore, be used to acquire general information about the remanence-carrying minerals within samples; this is done by the calculation of the median destructive field (MDF) (i.e. the field strength, in
mT, required to reduce the remanence of the sample to 50% of its initial NRM value) as well as by looking at the overall characteristics of the demagnetization intensity curve (Tauxe, 2009). High MDF values suggest fine-grained magnetic minerals, whilst low MDF values are representative of coarser grained magnetic minerals, and by comparing standard magnetite AF demagnetization curves of Argyle et al. (1994) (Figure 4.1), the approximate grain sizes of magnetite present can be inferred.

AF demagnetization of layered gabbro samples produced weakly sigmoidal remanence decay paths from which a mean MDF value (for the layered gabbros) of 19 mT (with a range of 9 to 26 mT) could be calculated (Figure 4.2). Overall, similar responses to AF demagnetization are seen in all the layered gabbro sites with MDFs of below 30 mT, with only 10% (or less) of the initial remanent magnetization remaining after 80 mT. The foliated gabbros from Wadi Abyad also displayed similar decay paths with a mean MDF of 22 mT (range of 6 to 32 mT), whilst the foliated gabbros from Wadi Khafifah tend to display either a much more exponential or a weakly sigmoidal decay giving a mean MDF value of just 11 mT (range of 4 to 17 mT). The Wadi Abyad foliated gabbros are far more resistant to AF demagnetization with between 10-20% of initial magnetization still remaining after 70 mT (Figure 4.3a), whereas in the Wadi Khafifah foliated gabbros the majority of samples have lost ~90% of their remanence by 50 mT (Figure 4.3b). The varitextured gabbros (only sampled in Wadi Abyad) show similar characteristics as the layered and foliated gabbros of the Wadi Abyad section with a mean MDF of 18 mT (range of 17 to 19 mT). After 100 mT (the maximum demagnetization field used) the residual remanence for the majority of gabbro samples was less (sometimes significantly less) than 10% of the initial value. Along with an overall mean MDF value of 17
mT (range of 3 to 40 mT) and given the weakly sigmoidal nature of the remanence intensity decay curves (and along with reference to Figure 4.1) the gabbros are most likely dominated by low to medium coercivity magnetic minerals with the range of MDF values likely reflecting subtle differences in grain size. These results, therefore, suggest that fine-grained pseudo-single-domain (PSD) magnetite grains (~2 µm in size) are the dominant magnetic carrier in the gabbro samples. The presence of larger PSD to fine-grained MD magnetite grains (~5 to 10 µm) is also likely in several samples, especially in the foliated gabbros of the Wadi Khafifah area. Dyke samples from Wadi Abyad show about 10-20% residual remanence left after the 100 mT demagnetization step and give a mean MDF value of 32 mT (Figure 4.4a). Dykes sampled at Khafifah (Figure 4.4b) and Somrah (Figure 4.4c) display a range of MDF values between 8 and 42 mT. However, in general they have a much “harder” magnetic carrier than the gabbros, as demonstrated by their higher mean MDF value of 28 mT, which suggests that medium to high coercivity SD magnetite grains (~0.6 µm) are the likely remanence carriers in the dykes.
Figure 4.2: Normalised magnetization intensity curves against field strength during AF demagnetization with calculated median destructive field (MDF) values for layered gabbros from: (a) Wadi Abyad; (b) Wadi Khafifah; (c) Somrah; (d) Wadi Nassif.

Figure 4.3: Normalised magnetization intensity curves against field strength during AF demagnetization with calculated median destructive field (MDF) values for foliated gabbros from: (a) Wadi Abyad; (b) Wadi Khafifah.
4.2.2 Magnetization unblocking temperature windows derived from thermal demagnetization

Thermal demagnetization is based on the inverse relationship between relaxation time and temperature: as the temperature increases the relaxation times of the magnetic grains present within the sample decrease until the unblocking temperature of each grain is reached, at which point they lose their stable remanence and become superparamagnetic. In a natural sample, unblocking may occur across a distribution of temperatures (up to the Curie temperature) or over a narrow, discrete temperature range, depending on ferromagnetic grain-size distributions (Butler 1998; Tauxe, 2009). The Curie temperature of pure magnetite, for example, is 578°C, whilst pure hematite is 680°C: other important ferromagnetic (s.l.) minerals have lower Curie
temperatures, for example, the Curie temperature of greigite is 350°C, whilst pyrrhotite is 320°C and goethite 127°C (Tarling & Hrouda 1993; Butler 1998; Tauxe, 2009). The addition of titanium (Ti) to the atomic structure of pure magnetite (Fe$_2$O$_4$) is also a common cause for a lowering of the Curie temperature, with greater concentrations of Ti leading to lower Curie temperatures (Tarling & Hrouda 1993; Butler 1998; Tauxe, 2009). First-order magnetic mineralogical information can, therefore, be obtained from the estimation of maximum unblocking temperatures from thermal demagnetization data.

Thermal demagnetization revealed three distinct unblocking temperature windows where remanence intensities were seen to clearly decrease:

1. High-temperature window – unblocking between approximately 500-590°C;
2. Intermediate-temperature window – unblocking between approximately 200-400°C, focused around 300-350°C;
3. Low-temperature window – unblocking between approximately 100-200°C, focused at 150°C.

The layered gabbros predominately display unblocking within the high-temperature window (500-590°C) in which approximately 80% of the initial remanence is lost (Figure 4.5), however, remanence was also lost from a number of layered gabbro samples from Wadi Khafifah (Figure 4.5b) within the intermediate-temperature window between 200-400°C (focused around 300-350°C) followed by further remanence loss within the high-temperature window, resulting in a noticeably stepped intensity decay curve. Several of the Wadi Khafifah layered gabbros also displayed remanence loss within the low-temperature window (100-200°C). Similar responses to thermal
demagnetization were also observed within the foliated gabbros from Wadi Abyad where the majority of samples lost approximately 80% of their initial remanence within the high-temperature unblocking window (Figure 4.6a). In contrast, a large number of the Wadi Khafifah foliated gabbros (Figure 4.6b) displayed stepped intensity decay curves, with remanence loss (on average 45-50%) taking place within the intermediate unblocking temperature window. The majority (about 85%) of remanence lost during demagnetization of the varitextured gabbros took place within the high-temperature unblocking window with full demagnetization taking place before 550°C (Figure 4.7a). Within the gabbroic units (i.e. layered, foliated and varitextured gabbros) the high-temperature unblocking window (between 500-590°C) is clearly the most significant in terms of remanence loss with, on average, over 75% of initial magnetization lost within this interval. Unblocking temperatures between 500-590°C are indicative of magnetite/titanomagnetite with the range of maximum unblocking temperatures most likely reflecting a mixture of differing grain sizes of the magnetic minerals. The higher maximum unblocking temperatures (~550-590°C) can be further constrained to be representative of magnetite, whilst the lower temperatures (~500-550°C) are most likely the result of increased titanium content within the magnetite grains (i.e. titanomagnetite). The intermediate-temperature unblocking window (200-400°C), that is focused around 300-350°C, possibly indicates the presence of pyrrhotite, while the low-temperature window may represents the unblocking of the iron oxyhydroxide mineral goethite. However, given the lack of pyrrhotite and goethite signatures in IRM (see section 4.2.3) and susceptibility vs. temperature runs (see section 4.2.4) the intermediate and low-temperature unblocking windows most likely represent the unblocking of discrete (coarser-grained) grain-size populations of
magnetite/titanomagnetite. The dykes displayed a similar range of unblocking temperatures to the gabbros with remanence loss taking place within all three unblocking temperature windows (Figure 4.7). Sites WA14 and WA16 show significant decrease in remanence within the intermediate-temperature unblocking window, while the low-temperature unblocking window is also present and sometimes significant within the Khafifah dyke sites. The maximum unblocking temperatures of these sites, however, falls within the high-temperature window above 500°C. Sites WA17, WA38 and SR06 do not display the stepped decay paths clearly seen in the previously described sites but show a gradual loss of remanence up to 500°C, indicative of a large ferromagnetic grain-size distribution, and then fully unblock within the high-temperature unblocking window. Site SR07, however, loses almost 80% of its remanence before 150°C, which is suggestive of a large contribution of possibly goethite (or another low unblocking mineral) to the initial remanence. Remanence loss at this site then stabilizes after 150°C and is followed by a gradual decrease in remanence towards a maximum unblocking temperature close to that of magnetite. A similar magnetic mineralogical assemblage to the gabbroic units can, therefore, be inferred for the dykes. However, due to the more significant presence of the low and intermediate-temperature unblocking windows (suggestive of coarser-grained magnetite/titanomagnetite) it can be hypothesized that the dykes (as well as a number of the Khafifah foliated gabbros) were subjected to more intense alteration as a result of the interaction with hot hydrothermal fluids.
Figure 4.5: Normalised magnetization intensity curves against temperature showing characteristic responses to thermal demagnetization for layered gabbros from: (a) Wadi Abyad; (b) Wadi Khaffaf; (c) Somrah; (d) Wadi Nassif.

Figure 4.6: Normalised magnetization intensity curves against temperature showing characteristic responses to thermal demagnetization for foliated gabbros from: (a) Wadi Abyad; (b) Wadi Khaffaf.
4.2.3 Isothermal remanent magnetization (IRM) experiments

As described in Chapter 2, section 2.9.1, isothermal remanent magnetization (IRM) acquisition results from the application of a strong magnetizing field in a stepwise manner and at a constant temperature until magnetic saturation is achieved. As the magnetizing field is increased, the ferromagnetic (s.l.) grains with coercivities equal to or less than the field will be forced to align themselves in the orientation of the applied field resulting in a net increase of magnetization (Butler, 1998; Tauxe, 2009). Once magnetically saturated, further increases in net magnetization do not occur as all ferromagnetic (s.l.) grains are already aligned with the magnetizing field and the sample is said to hold a saturation IRM (SIRM). The value of SIRM (in mT) and the shape of the IRM acquisition curve are wholly dependent upon the type,
concentration and domain state of the magnetic minerals present within the sample. SD magnetite, for example, reaches a maximum SIRM by approximately 300 mT while MD magnetite will saturate in fields of only a few 10’s mT, and hematite requires far stronger magnetizing fields, between 1.5-5.0 T, in order to reach magnetic saturation. Once magnetically saturated, back-field IRM acquisition can be used to remagnetize half the ferromagnetic (s.l.) grains in the opposite direction allowing for the calculation of the coercivity of remanence (Hc), which can be taken as the magnetizing equivalent of the MDF and is itself dependent upon the grain-size/domain state of the ferromagnetic (s.l.) minerals present. The field required to achieve saturation and the coercivity of remanence can, therefore, provide useful information on the ferromagnetic (s.l.) mineralogy of a sample as well as the concentration and domain state of these minerals.

IRM saturation of the gabbroic rocks is generally achieved by 300 mT with some upper foliated and varitextured gabbros not fully saturating until 500 mT (Figure 4.8, 5.9 & 5.10a). These IRM curves are characterized by initial rapid acquisition up to 100 mT followed by more a gradual increase in isothermal remanent magnetization up to saturation. Discrete dykes sampled from throughout the plutonic section (from Wadi Abyad, Khafifah and Somrah) tend to saturate in magnetizing fields between 300-500 mT (with the exception of site WA38 which saturates by 100 mT). Some samples, however, require fields greater than 500 mT before fully saturating (e.g. sites WA17 and KF09) (Figure 4.10). With saturation of the gabbros generally achieved by 300 mT and of the dykes by 500 mT, these data suggest the presence of low to medium coercivity fine-grained magnetic carriers, most likely SD to PSD magnetite, with no indication of high coercivity magnetic minerals such a haematite. This is
supported by average coercivity of remanence (Hc) values of 25.4 mT for the layered gabbros, 24.8 mT for the foliated gabbros and 32.8 mT for the varitextured gabbros. These Hc values are consistent with the presence of fine-grained PSD magnetite in the layered and foliated gabbros, whilst the increase in Hc value seen in the varitextured gabbros may be the result of a slight reduction in grain size of the magnetic carrier but still within the PSD (magnetite) range. A mean Hc value of 32.9 mT was calculated for the dykes, which again is consistent with magnetite grain sizes in the PSD range.

Figure 4.8: Representative normalised IRM acquisition and back-field IRM acquisition curves for layered gabbros from: (a) Wadi Abyad; (b) Wadi Khaffah; (c) Somrah; (d) Wadi Nassif.
Figure 4.9: Representative normalised IRM acquisition and back-field IRM acquisition curves for foliated gabbros from: (a) Wadi Abyad; (b) Wadi Khafifah.

Figure 4.10: Representative normalised IRM acquisition and back-field IRM acquisition curves for varitextured gabbros and dykes sampled in: (a) Wadi Abyad; (b) Wadi Khafifah; (c) Somrah.
4.2.4 Variation of magnetic susceptibility with temperature and the determination of the Curie temperature

The variation of magnetic susceptibility with temperature (as described in Chapter 2, section 2.9.2) can be used to obtain useful information regarding the magnetic mineralogy of a sample as well as the specific contributions of these minerals to the net susceptibility. The variation of susceptibility with temperature also enables an estimation of the Curie points (i.e. the temperature at which ferromagnetic (s.l.) material behaves paramagnetically) of the minerals that are contributing to the susceptibility. Much like thermal demagnetization, thermomagnetic susceptibility experiments are based on the inverse relationship between temperature and relaxation time. Once the Curie temperature of the magnetic grains within a sample is reached a drastic reduction of the magnetic susceptibility will take place due to the significantly lower susceptibility of paramagnetic materials compared to ferromagnetic (s.l.) materials. Several methods for the estimation of the Curie point (Tc) from thermomagnetic curves have been developed with most authors applying the two-tangent method of Grommé et al. (1969). However, in their paper on Tc determination, Petrovsky & Kapička (2006) discuss how this method often overestimates the Tc, placing it at a temperature when the substance is in fact behaving paramagnetically. Petrovsky & Kapička (2006), therefore, with reference to the work of Hrouda et al. (2005), developed a new method that uses the linear relationship of inverse susceptibility and its dependence on temperature after the Tc to determine the temperature at which the substance starts to behave paramagnetically and defined this as the Curie temperature (see Petrovsky & Kapička (2006) for full details of the inverse susceptibility method). Given the possible errors involved with using the two-tangent method, this alternative method has been used in this study. In addition, when a sharp
Hopkinson peak was observed in the thermomagnetic curve, the temperature at which the maxima of the peak occurred was also noted, whilst if a wide susceptibility peak was present before rapid susceptibility loss, the inflection point on the decreasing curve was taken to define the onset of paramagnetic behaviour.

Prior to the Tc, the majority of the gabbroic samples displayed a noticeable bump in susceptibility on their thermomagnetic curves during heating that peaked between 250-300°C (e.g. WA1008H, Figure 4.11a). This peak is taken to represent the alteration of titanomagnetite to titanomaghemite as a result of low-temperature oxidation (below 300°C), whilst the subsequent decrease in susceptibility above 300°C is believed to be associated with high-temperature oxidation (of the titanomaghemite) to magnetite (and possibly ilmenite) (Ozima & Sakamoto, 1971). The remaining thermomagnetic curves, however, either display a steady increase, a very gradual decrease or no change in magnetic susceptibility with increasing temperature (Figures 5.11 and 5.12). At the Tc for all the gabbroic samples an increase in susceptibility is often observed due to the Hopkinson effect that can result in sometimes very pronounced Hopkinson peaks, which are related to the superparamagnetic behaviour of magnetic particles (e.g. WA1203H, Figure 4.11a). These results are consistent with magnetite/titanomagnetite being the dominant magnetic carrier within the gabbros. This is supported by unblocking temperatures of approximately 566°C in the layered gabbros, 565°C in the foliated gabbros and at 578°C in the varitextured gabbros, resulting in an overall mean unblocking temperature of 566°C for the gabbros. Several samples from the upper foliated and varitextured gabbros of the Wadi Abyad section also displayed some initial loss of susceptibility between 500-530°C followed by a step at approximately
565-570°C after which continued (and more rapid) susceptibility loss was observed (Figures 5.12 and 5.13a). These “stepped” susceptibility decay thermomagnetic curves are possibly related to the presence of titanomagnetite within these samples. The majority of the heating and cooling curves from the gabbro samples are found to be non-reversible and show an increase, sometimes a significant increase, in susceptibility during cooling. This results from the formation of new, more strongly magnetic phases of magnetite/titanomagnetite during the heating cycle (up to ~700°C).

Thermomagnetic curves for dyke samples tend to display a wide variety of characteristics. Dyke sites WA14 and WA16 (from within the varitextured gabbros) display a gradual loss of susceptibility as temperature increases followed by a sharp rise in susceptibility around 450°C to form a Hopkinson peak due to the creation of new magnetic phases, after which (at 560°C) a rapid loss of susceptibility occurs as the Tc is reached (Figure 4.13a). Sites WA38 and WA17 on the other hand display similar SD magnetite/titanomagnetite characteristics as seen in the gabbros (Figure 4.13a). The cooling curves for the Wadi Abyad dykes (except WA38, see below), like the gabbros, display a susceptibility increase (i.e. are non-reversible). The cooling curve for sample WA3801A at first displays significantly lower susceptibility values until approximately 400°C at which point it become reversible, this is taken to represent the destruction of magnetite during heating. Two dykes sampled from Khafifah and Somrah (from sites KF09 and SR06) display magnetite/titanomagnetcite characteristics with small titanomaghemite peaks, whilst site KF07 has very low concentrations of ferromagnetic (s.l.) material and, therefore, produces a very noisy thermomagnetic curve. The cooling curves from these dyke sites are also non-reversible and display an increase in
magnetic susceptibility. Site SR07 (sample SR0701B, Figure 4.13c), however, displays paramagnetic behaviour (as evidenced by the parabolic heating curves). These characteristics, as well as an average unblocking temperature for the dykes of approximately 550°C, are assumed to indicate the presence of magnetite/titanomagnetite, most likely titanomagnetite, as the dominant magnetic carrier with the dykes.

Figure 4.11: Characteristic normalised thermomagnetic curves for layered gabbros from: (a) Wadi Abyad; (b) Wadi Khafifah; (c) Somrah; (d) Wadi Nassif. Solid lines represents heating curves up to 700°C, whilst dashed lines represents the cooling curves from 700°C back to ambient temperature.
The inverse susceptibility method of Petrovsky & Kapička (2006) was used to accurately calculate the Curie temperature. Tc values of 590°C, 592°C and 601°C for the layered, foliated and varitextured gabbros respectively (collective mean Tc of 592°C) were calculated with a Tc of 559°C calculated for the dykes. These Tc values are systematically higher than the unblocking

Figure 4.12: Characteristic normalised thermomagnetic curves for foliated gabbros from: (a) Wadi Abyad; (b) Wadi Khafifah.

Figure 4.13: Characteristic normalised thermomagnetic curves for varitextured gabbros and dykes sampled in: (a) Wadi Abyad; (b) Wadi Khafifah; (c) Somrah.
temperatures inferred from the point on the thermomagnetic susceptibility of rapid susceptibility loss or, if present, the zenith of the Hopkinson peak by approximately 25°C (for the gabbros and 14°C for the dykes). The calculated Tc values for the gabbroic units are also higher than would expected for magnetite (Tc = 578°C) or titanomagnetite (Tc = <578°C). Above the Curie temperature magnetic susceptibility will rapidly fall to (almost) zero due to the fact that the material will now only display paramagnetic susceptibility. The point as which this sharp decrease in magnetic susceptibility (i.e. unblocking temperatures) is observed, therefore, provides a far more accurate point for when ferromagnetic (s.l.) behaviour ceases and paramagnetic behaviour begins and can be considered as a good approximation of Curie temperature. These calculated Tc values shall, therefore, be disregarded, as they do not aid in the classification of the magnetic mineralogy. In summary, the unblocking temperatures (approximate Tc’s) for the layered and foliated gabbros are suggestive of magnetite (maybe with some titanium impurities as the value is below the Tc of pure magnetite), whilst the slightly higher unblocking values recorded in the varitextured gabbros suggest a more pure form of magnetite. The lower unblocking temperatures recorded in the dykes, however, imply a more Ti-rich magnetite or titanomagnetite being the dominant magnetic carrier.
Figure 4.14: Temperature dependence of inverse magnetic susceptibility for a selection of thermomagnetic curves from sites sampled within: (a) the layered gabbros (LG); (b) the foliated gabbros (FG); (c) the varitextured gabbros (VG); (d) the discrete dykes (D). The Curie temperature (Tc) has been calculated using the inverse susceptibility method of Petrovsky & Kapička (2006).

4.3 Summary of magnetic mineralogy of the lower oceanic crust of the Semail ophiolite

Rock magnetic experiments suggest that SD to fine-grained PSD secondary magnetite/titanomagnetite (approximately 1-2 µm in size) is the dominant magnetite carrier within the gabbros and dykes. The presence of larger more coarse-grained PSD to fine-grained MD secondary magnetite (~5 to 10 µm) is also implied, especially in the foliated gabbros of the Wadi Khafifah area, as a result of more intense hydrothermal alteration. In addition, pyrrhotite and goethite (formed due to hydrothermal alteration) may also be present but are not considered to be a significant magnetic carriers. These results are in agreement with previous rock magnetic studies on the gabbros and lowermost
dykes from the Semail ophiolite (e.g. Luyendyk et al., 1982; Luyendyk & Day, 1982; Thomas et al., 1998; Feinberg et al., 1999; Weiler, 2000; Yaouancq & MacLeod, 2000; Kawamura et al., 2005; Usui & Yamazaki, 2010) as well as the results from other Tethyan ophiolites (e.g. Inwood et al., 2009) and the gabbros from Hess Deep (Richter et al., 1996). These studies revealed that predominately PSD secondary magnetite grains associated with serpentinized olivines and hydrothermally altered clinopyroxenes are the main remanence carrier within these rocks. Both Weiler (2000) and Yaouancq & MacLeod (2000) reported the presence of pyrrhotite within the higher-level gabbros (i.e. upper foliated and varitextured gabbros) related to more intense hydrothermal alteration. Finally, Yaouancq & MacLeod (2000) also identified that the magnetite found within varitextured gabbros and dyke-rooting zone was mostly primary magnetite (but secondary magnetite was still observed to be associated with olivine and clinopyroxene).
Chapter 5: Magnetic fabrics

5.1 Introduction

As outlined in Chapter 2, the Semail ophiolite is an important field laboratory for testing models of oceanic crustal accretion. Previous studies on oceanic crustal accretion models have focused on petrological, geochemical and structural techniques to determine accretion processes through both *in situ* scientific ocean drilling and fieldwork in ophiolites. The use of magnetic fabric (anisotropy) analyses as a tool for quantifying the variation of crystalline fabrics in ophiolitic sections has, however, been poorly exploited. To this end, and with a focus on the lower oceanic crust, sampling for magnetic fabric (and palaeomagnetic) analyses at different pseudo-stratigraphic levels was undertaken in gabbroic sequences at four localities within the Semail ophiolite (Wadi Abyad, Wadi Khafifah, Somrah and Wadi Nassif). In this chapter, bulk susceptibility and the relationship between anisotropy of magnetic susceptibility (AMS) and anisotropy of partial anhysteretic remanent magnetization (ApARM) are used to determine the source of the AMS signal in the lower crust. The relationship between AMS principal axes and both macroscopic layering/foliation (measured in the field) and microscopic preferred crystal alignments (from observations of orientated thin sections) is discussed, and variations in fabric development are used to test alternative models of magmatic crustal accretion. Additional magnetic fabric data from individual discrete dykes cutting the gabbroic sequences and from a pilot study of a gabbroic sill within the Tuf area are presented in Appendix A and B (respectively for completeness.

The AMS (and ApARM) data are presented graphically in the form of equal area stereographic projections and bi-plots of anisotropy parameters.
Most data are considered at the site level, but some sites sampled geographically close together (e.g. WA04A+B, KF04-06 and SR01-05) have been combined, as they gave generally comparable results and can, therefore, be regarded as single sampling sites.

5.2 Mineralogical source of the AMS signal

Mean (or bulk) magnetic susceptibility can provide basic information on the carriers of AMS signals as it is primarily controlled by composition and concentration of different mineral phases. Figure 5.1 shows the variation of susceptibility due to different percentage concentrations of common rock-forming dia-, para- and ferromagnetic minerals (Tarling & Hrouda, 1993). Rocks dominated by paramagnetic phases (e.g. pyroxenes) may reach maximum susceptibilities of ~$10^{-4}$ [SI], but magnetite will dominate the susceptibility signal if present in concentrations $>~0.1\%$ by weight.

Figure 5.1: Variation of low field magnetic susceptibility with concentration of various common rock forming minerals (from Tarling & Hrouda, 1993).
The histograms shown in Figure 5.2 show the distributions of mean susceptibilities of samples from each of the pseudo-stratigraphic levels sampled in the lower crust of the ophiolite, plotted using log\(_{10}\) values and accompanied by associated geometric mean (log\(_{10}\)) values. Site-level mean susceptibility values are listed in Table 5.1. Log\(_{10}\) susceptibility values are generally symmetrically distributed around the geometric mean values, i.e. are characterized by approximately log-normal distributions, consistent with most large magnetization data sets (Tarling, 1983; Gee & Kent, 2007). The geometric mean susceptibilities (taking anti-logs) and ranges are:

- **Layered gabbros**: mean = 2.09 \(\times\) 10\(^{-3}\), min = 0.089 \(\times\) 10\(^{-3}\), max = 83.0 \(\times\) 10\(^{-3}\) [SI]
- **Foliated gabbros** = 4.90 \(\times\) 10\(^{-3}\), min = 0.45 \(\times\) 10\(^{-3}\), max = 73.0 \(\times\) 10\(^{-3}\) [SI]
- **Varitextured gabbros** = 4.07 \(\times\) 10\(^{-3}\), min = 0.098 \(\times\) 10\(^{-3}\), max = 10.9 \(\times\) 10\(^{-3}\) [SI]

These mean values exceed the maximum susceptibility of the major silicate minerals in these rocks and require a contribution from ferrimagnetic phases. Rock magnetic analyses (Chapter 4) indicate that magnetite is present in all of the gabbros studied here. If all of the susceptibility was due to magnetite, the

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1 It is interesting to note that these data are entirely consistent with values (c. 0.1 \(\times\) 10\(^{-3}\) to 40 \(\times\) 10\(^{-3}\) [SI]) reported from the first drill core samples of layered gabbros recovered by scientific ocean drilling (Gillis et al., 2014) from the world’s oceans (East Pacific Rise-generated lower crust sampled by IODP Expedition 345 to Hess Deep).
mean susceptibility data would be consistent with magnetite concentrations of ~0.1 wt% (which is highly plausible in gabbroic rocks; Figure 5.1).

However, estimated modal compositions of the gabbros suggest that 20-40% of these lithologies consists of clinopyroxene, which, as a paramagnetic mineral with typical susceptibilities in the range of 0.5 – 5.0 x 10^-3 [SI] (Rochette et al., 1992), may make significant contributions in more magnetite-poor samples. A composite paramagnetic + ferrimagnetic source in such samples is consistent with the first-order guide of Tarling & Hrouda (1993), who suggest that rocks with mean susceptibilities in the range 5 x 10^-4 to 5 x 10^-3 [SI] typically indicate combined paramagnetic and ferrimagnetic contributions.

Table 5.1: Summary of site-level magnetic susceptibility and AMS parameters.
Further evidence for the important contribution of magnetite to the AMS signal in the layered and foliated gabbros is provided by (i) high temperature susceptibility experiments that show magnetite Curie temperatures (see Chapter 4, section 4.2.4), and (ii) a close agreement between the orientations of principal axes of AMS and ApARM ellipsoids in these rocks and by thin section analyses, as presented elsewhere in this chapter. ApARM, a form of remanence anisotropy, reflects the distribution of only the ferromagnetic (remanence carrying) minerals present in a sample (magnetite in these rocks; see Chapter 4). AMS and ApARM fabrics are shown to be coaxial (see Figure 5.5 and section 5.4.3), and thin sections cut parallel to AMS principal axes show that AMS $K_{\text{max}}$ axes are aligned with the preferred orientation of silicate crystals (defined by crystal long axes) (see section 5.3.3 and 5.4.4). These observations suggest that ferrimagnetic and paramagnetic fabrics are parallel and hence that the distribution of magnetite is controlled by silicate fabrics (e.g. as secondary grains formed aligned along fractures in olivine crystals and as exsolved inclusions in plagioclase and/or clinopyroxene crystals). It is likely that the AMS signal of the magnetite represents a distribution anisotropy (Hargraves et al., 1991) resulting from an uneven distribution of grains that are close enough to interact magnetostatically. Several theoretical models for distribution anisotropy (Stephenson, 1994; Cañon-Tapia, 1996; Muxworthy & Williams, 2004) show that when grains become close enough to interact magnetostatically then their distribution rather than their individual orientations dominates AMS signals. The dual contribution of magnetite and silicate phases to the observed AMS fabrics is returned to briefly below in relation to mean susceptibility and anisotropy degree data from the layered gabbro sites.
5.3 Magnetic fabrics in layered gabbros

5.3.1 Anisotropy of magnetic susceptibility results

Layered gabbros were sampled in Wadi Abyad, Wadi Khafifah, at Somrah and in Wadi Nassif, and corresponding site mean susceptibility and AMS parameters are listed in Table 5.1. The shape of individual specimen AMS ellipsoids is described by the parameter $T$, with $0 < T \leq 1.0$ for oblate ellipsoids and $-1.0 \leq T < 0$ for prolate ellipsoids. The majority of specimens exhibit oblate AMS ellipsoids with a maximum $T$ of 0.95 (Figure 5.3), although a significant number of specimens with prolate ellipsoids occur. Strength of anisotropy is described by the corrected anisotropy degree, $P_J$ (Jelinek, 1981), and ranges from 1.02 to 1.46 for individual specimens (Figure 5.3), with a modal value of 1.24 (i.e. 24% anisotropy). There is no preferred relationship between $P_J$ and $T$, although the majority of samples with prolate ellipsoids have $P_J$ values < 1.15.

There is, however, a broad positive correlation between $P_J$ and mean susceptibility (Figure 5.3), suggesting that the degree of anisotropy in these rocks is dependent on ferrimagnetic concentration. This is typically encountered (Rochette et al., 1992) in rocks where susceptibility is due to more than one mineral with coaxial normal fabrics but different anisotropy ratios. If the proportion between different contributions changes between specimens then the overall anisotropy will show variations related to composition (rather than, for instance, differences in strain; Rochette et al., 1992). These data, therefore, support the inference that AMS in these gabbros is carried by both paramagnetic silicate phases and magnetite.
At a site level, clustering of $K_{\text{max}}$ and $K_{\text{min}}$ axes defines the magnetic lineation and the pole to the magnetic foliation respectively. Oblate fabrics are characterised by clusters of $K_{\text{min}}$ axes orthogonal to great circle girdle distributions of $K_{\text{max}}$ and $K_{\text{int}}$ axes, whereas prolate fabrics are characterised by clusters of $K_{\text{max}}$ axes orthogonal to girdle distributions of $K_{\text{int}}$ and $K_{\text{min}}$ axes. In triaxial fabrics, specimen principal susceptibility axes form three distinct groups. Orientations of the principal AMS axes at each layered gabbro site are compared to macroscopic modal layering (determined in the field) in the stereographic equal area projections of Figure 5.4, presented in geographic coordinates. Mean ellipsoid principal axes and corresponding confidence

Figure 5.3: Summary of AMS parameters for layered and foliated gabbros sampled in this study. Left: variation of corrected anisotropy degree with mean susceptibility; Right: Jelinek plot of corrected anisotropy degree against shape parameter.
ellipses (Jelinek, 1978) are also shown in these projections. In the majority of cases a clear correlation of at least one AMS principal axis and the measured petrofabric is seen. Some slight mismatches between magnetic foliations/lineations and orientations of macroscopic structures observed in the field (e.g. at sites WA12 in Wadi Abyad and WN05 in Wadi Nassif; Figure 5.4) may be due to local variability at the point of sampling compared to the average orientation of layering measured over a larger area.

Figure 5.4: AMS data from layered gabbros sampled in the Semail ophiolite, showing orientations of specimen AMS principal axes, with associated site mean axes (larger symbols) and Jelinek (1978) confidence ellipses. WA = Wadi Abyad, KF = Wadi Khaffah area, SR = Somrah and WN = Wadi Nassif. Black great circles = orientation of macroscopic layering observed in the field; red diamonds = orientation of lineations defined by preferred alignments of crystal long axes observed in the field; red great circles = orientation of fault planes observed at site WA04A+B.
On the basis of the relationships between AMS principal axes and macroscopic structures, the AMS results in the layered gabbros may be divided into three groups:

1) Sites where $K_{\text{max}}$ and $K_{\text{int}}$ axes lie in or close to the plane of macroscopic layering and where $K_{\text{min}}$ axes lie at or close to the pole to layering; i.e. there is a close match between magmatic and magnetic foliations (sites: WA09, 12; KF03, 04-06, 08; SR01-05; WN01, 04, 05). This is consistent with layering-parallel, normal, oblate fabrics at these sites.

2) Sites where $K_{\text{max}}$ axes lie in or close to the plane of macroscopic layering and where $K_{\text{int}}$ and $K_{\text{min}}$ axes form (semi-)girdle distributions outside of the plane of layering (sites: WA10, WA11, WN03). This pattern is consistent with prolate fabrics at site-level, with ellipsoid long-axes lying within the magmatic foliation.

3) Sites with complex fabrics that are only partly related to the observed macroscopic layering (sites: WA04A&B, WN02). In the case of site WA04A+B, samples were collected close to significant faults that cut the magmatic layering (orientations of which are indicated by the orange great circles in the stereonet for this site in Figure 5.4). $K_{\text{int}}$ axes are strung out along the steeply-dipping N-S-striking fault suggesting an anomalous composite magmatic and tectonic fabric at the site. At site WN02, two clusters of $K_{\text{max}}$ axes are present, and the dominant cluster is consistent with presence of a prolate fabric of type (2) at the site.
5.3.2 Anisotropy of partial anhysteretic remanence (ApARM) results

Anisotropy of partial anhysteretic remanent magnetization (ApARM) was conducted on a number of specimens from the layered gabbro section sampled at site KF08 to allow comparison with the AMS results. Whereas AMS measures the bulk anisotropy of magnetic susceptibility of the whole specimen (i.e. the dia-, para- and ferromagnetic (s.l.) fractions), ARM experiments focus only on the remanence-carrying ferromagnetic (s.l.) minerals, providing information solely on the preferred alignment of this mineral fraction. ApARM measurements were undertaken at the Ludwig-Maximilians-Universität München (LMU), Department of Earth and Environmental Sciences in Munich, Germany, on the “SushiBar” system. A biasing DF of 0.05 mT within a coercivity window of 60-90 mT was used, so that only the fine-grained, higher coercivity SD and PSD magnetite grains would acquire a magnetization.

Clear consistency can been seen between the AMS and ApARM data sets (Figure 5.5), with the majority of, for example, ApARM maximum susceptibility axes (i.e. pARM$_{\text{max}}$) deviating less than 20° from the corresponding AMS maximum susceptibility axes (i.e. K$_{\text{max}}$) (Table 5.2). These results indicate that the AMS signal is either dominated by magnetite or that magnetite and paramagnetic (silicate) fabrics are coaxial in lower susceptibility (more magnetite-poor) samples. Moreover, the “normal” (Rochette et al., 1992), oblate arrangement of principal anisotropy axes observed in the AMS data at this site is also observed in the ApARM data, suggesting that little or no SD magnetite is present, as this would produce inverse AMS fabrics (with interchange of K$_{\text{max}}$ and K$_{\text{min}}$ axes; Potter and Stephenson, 1988) resulting in K$_{\text{max}}$ axes that correlate with pARM$_{\text{min}}$ axes.
Figure 5.5: AMS (left) and ApARM (right) data from layered gabbros at site KF08 at Khafifah. Note the excellent correlation between these two forms of anisotropy, demonstrating that AMS data are not affected by single-domain-related inverse fabrics.

Table 5.2: Comparison of the orientations of $K_{\max}$ and $pARM_{\max}$ axes at site KF08, with calculated angular differences. Dec. = declination; Inc. = inclination.
5.3.3 Petrographic analyses

Thin-section analyses were carried out on samples from the layered gabbros to support field mineralogy observations and to directly observe, if possible, the magnetic minerals and also to establish the mineralogical controls on the magnetic fabrics identified from AMS analyses. Thin sections were cut in the plane of the $K_{\text{max}}$-$K_{\text{min}}$ AMS principal axes to allow direct comparison with the magnetic fabric data.

In the field the layered gabbros are characterized by well-defined cm to m scale layers controlled by the modal proportions of plagioclase feldspar, pyroxene (most likely clinopyroxene) and olivine. Pyroxene and ± olivine dominate the base of layers, whereas plagioclase is the dominant crystallization product towards the top of layers. In order to quantify the variation of the three main igneous phases through a typical gabbro layer, point-counting (400 counts) was conducted on a set of thin sections prepared from core samples collected from Layer A of the Somrah layered gabbro section, representing a near perfect example of modal layering. This provided an almost continuous view through an individual layer and showed in great detail the petrofabric features and crystallization products along with their percentage change through the ~60 cm thick section.

Initial analysis of thin sections validated field observations verifying the presence of pyroxene (which was identified as clinopyroxene), plagioclase feldspar and olivine, which had undergone variable degrees of serpentinization (Figure 5.6a+b). The serpentinization of olivine is characterised by heavy fracturing (that is commonly aligned parallel with the $K_{\text{max}}$ axis) and veins (also aligned) filled with opaques. Alteration rims of very fine-grained acicular tremolite, chlorite (as well as other fine-grained clay minerals) and additional opaques surround the olivine crystals (Figure 5.6c+d). Rare equant opaques
Figure 5.6: Photomicrographs of thin sections of layered gabbros (left-hand side = plane polarised light; right-hand side = under crossed polars), showing: (a, b) magmatic textures defined by interlocking, undeformed crystals of plagioclase, clinopyroxene and olivine; (c, d) the typical nature of serpentinization of olivine crystals and presence of alteration halos; and (e, f) equant opaque minerals, assumed to be magnetite, distributed along cleavage planes in clinopyroxene crystals. Ol = olivine, Cpx = clinopyroxene, Pl = plagioclase feldspar, Mag = magnetite. Mineral symbols in accordance with Kretz (1983).
could also be observed under high magnification as inclusions within the cleavage planes of some clinopyroxene grains (Figure 5.6e+f). Although formal identification of opaque minerals cannot be conducted using a polarising microscope, the opaques associated with the olivines, as well as those within clinopyroxene cleavage planes, are interpreted to be secondary magnetite. Magnetite is a common by-product of the serpentinization of olivine, whereas magnetite within pyroxenes may form by exsolution during cooling (Maffione et al., 2014; Renne et al., 2002; Butler, 1998). Primary interstitial oxide minerals are, however, notably lacking from these layered gabbro samples. The proportion of magnetite observed within the fracture planes of olivine or as inclusions in the cleavage planes of clinopyroxenes was, therefore, also noted during point-counting, whilst zones of abundant alteration were also recorded to assess how alteration varied with height through the layer. The results from point-counting are presented in Figure 5.7, and may be used to further subdivide the layer into zones (Ai-Aiii) based on modal composition.

Figure 5.7 shows that significant proportions of both clinopyroxene (~40%) and plagioclase (~35%) crystallized at the base of Layer A, with olivine contributing about 8% of the total composition. The relative proportions of clinopyroxene and plagioclase remain consistent throughout zone Ai (base to ~10 cm up-layer), whereas olivine crystallization reduces. Throughout the lower part of Layer A the long axes of elongate clinopyroxene and plagioclase (plus to some extent the olivine) grains are observed to be aligned along the measured $K_{\text{max}}$ axis (Figure 5.8a+b). Dramatic changes in the proportions of clinopyroxene and plagioclase, however, take place within zone Aii (~10 cm up to ~20 cm) with the amount of plagioclase increasing to a peak of almost 70% of the total, whilst clinopyroxene reduces to just 20% at the transition from zone Aii to zone
Figure 5.7: Results of point-counting analysis of thin sections from Somrah Layer A (400 counts), showing the variation of the three main igneous phases (clinopyroxene, plagioclase and olivine) as well as the percentages of magnetite and alteration through a typical gabbro layer.

Aiii (Figure 5.8c+d). Olivine percentage continues to decrease up through the layer (zones Ai and Aii) before effectively disappearing altogether midway through the layer in zone Aiii (Figure 5.7). Throughout the remaining ~35 cm of Layer A (all within zone Aiii) a gradual reduction in plagioclase is observed that is countered by an increase in clinopyroxene. The alignment of crystal long axes relative to \( K_{\text{max}} \) axes continues throughout the layer, with the crystallographic alignment clearly of a magmatic origin and showing very little to
no signs of crystal plastic deformation (Figure 5.8e+f). Significant grain boundary migration, as a consequence of a long cooling-history, has also resulted in rounded and embayed interlocking mineral edges (Figure 5.8g+h). Small rounded inclusions of both plagioclase and clinopyroxene (and sometimes olivine) grains can be observed within the clinopyroxenes (i.e. poikilitic/ophitic texture) (Figure 5.8g+h), whereas several plagioclase grains contain fresh clinopyroxene inclusions and there are very rare cases where olivine grains are included in larger plagioclase grains. A crystallization order of olivine followed by predominately syn-crystallization of clinopyroxene and plagioclase is, therefore, inferred suggesting the possibility of both SSZ-type (hydrous) and MORB-type (normal) characteristics, which are typical of SSZ-type magmatic suites (Pearce et al., 1984; Dilek & Furnes, 2011). Signs of alteration of all three of the primary phases (but clearly focused around olivine crystals) are also observable throughout the layer but especially noticeable in thin sections from the base and top, with point-counting confirming the extent to which the intensity of alteration varied throughout the layer (Figure 5.7). Commonly, a greenschist facies alteration assemblage of tremolite (alteration product of olivine and clinopyroxene), epidote/clinozoisite (alteration product of plagioclase feldspar) and chlorite (further alteration of olivine and clinopyroxene and also tremolite) can be identified (Figure 5.8i+j). A clear association between olivine and the presence and quantity of magnetite is also highlighted from the point-counting results (Figure 5.7). The greatest concentrations of magnetite are found where the amounts of olivine are highest (i.e. at the base of the layer), and as olivine percentage decreases up-layer so does the amount of magnetite.

A large thin-section was also prepared from a hand sample collected from the boundary between two layers (Layers A and B) and cut perpendicular
Figure 5.8: Photomicrographs of thin sections of layered gabbros from Somrah (lefthand side = plane polarised light; righthand side = under crossed polars), illustrating changes in mineralogy through a single layer. See text for details. Ol = olivine, Cpx = clinopyroxene, Pl = plagioclase feldspar.
to the trend of the foliation and, therefore, the layer boundary, in order to look in greater detail at the nature of the contact. This revealed a rather diffuse boundary between the two layers defined by a mixing zone where, most noticeably, olivine grains appear to have sunk down from the base of Layer B in to the top of Layer A making it hard to define the contact precisely. Also, across the contact no significant change in the size and/or amount of plagioclase was seen; however, a marked increase in the proportion and size of clinopyroxene grains was used to estimate its location. The diffuse nature of the contact is taken to suggest that both layers (i.e. Layers A and B) were still hot and in a semi-solid state (i.e. crystalline mush) when they formed. Throughout the thin-section a clear alignment of the long axes of elongate plagioclase and clinopyroxene grains (and also olivine when present) could be observed mimicking the orientation of the boundary between the layers, whilst thin veins of alteration, which are common throughout the thin-section but more abundant in the vicinity of the contact, and fractures within the olivines also run parallel with the predominate crystallographic alignment. The abundance of alteration veins close to the contact (both within the top of Layer A and the base of Layer B) is believed to be a result of hydrothermal fluids exploiting the weakness between the two layers. The route of the hydrothermal fluids through these rocks has, therefore, resulted in foliation-parallel veining in the plagioclase and clinopyroxene, exsolution of the clinopyroxenes forming (titano)magnetite within the cleavage planes and magnetite filled fractures (also aligned with the foliation plane) within the olivines due to serpentinization.

In summary, the petrographic analysis of samples from Somrah confirms that AMS fabrics in the layered gabbros, which are dominated by the signal from magnetite, are parallel to silicate fabrics defined by the long-axes of the
major phases seen in oriented thin sections. This suggests that both ferrimagnetic and paramagnetic contributions to AMS are co-axial, resulting from a strong silicate crystallographic control on the distribution of magnetite (formed both as secondary magnetite resulting from serpentinization of olivine and as exsolved inclusions in clinopyroxene (and potentially plagioclase) crystals).

5.3.4 Vertical variations in anisotropy parameters within a single unit of layered gabbro

Layer A from the Somrah layered gabbro section was extensively sampled (at approximately 5 cm intervals over its ~60 cm thickness) to produce a high-resolution dataset of how AMS parameters (Km, Pj and T) and orientation of AMS principal axes change through the layer in response to vertical changes in composition across the layer, defined principally by modal percentages of clinopyroxene, plagioclase and olivine determined from thin sections (Figure. 5.7).

Figure 5.9 displays the variation of mean susceptibility, corrected anisotropy degree (Pj), shape parameter (T), and inclination of $K_{\text{max}}$ axes with height through Layer A. Mean susceptibilities are highest in zone Ai at the base of the layer, with significantly lower values in the more leucocratic zone Aiii at the top. Thin-section analysis show that the higher susceptibility values correspond to the presence of significant serpentinized olivine crystals containing secondary magnetite within fractures and veins. The very low amounts of olivine found within the upper part of the layer (Figure 5.7) results in low mean susceptibility values due to other phases in zone Aiii (Figure 5.9). Within this zone there is a clear co-variation of mean susceptibility and anisotropy degree (Figure 5.9), which becomes more apparent when these
parameters are plotted against each other (Figure 5.10). This reveals a power law relationship (due to the log-normal nature of susceptibility) that suggests that the anisotropy is controlled by the presence of two minerals with co-axial fabrics (Rochette et al., 1992), as seen in the larger dataset from all of the layered gabbro sites (see Figure 5.3). Thin section analysis (Figure 5.7) shows that this zone exhibits inverse co-variation in the proportions of clinopyroxene and plagioclase, suggesting that varying contributions from magnetite inclusions in these phases and from the paramagnetic signal of the clinopyroxene controls the signal. Hence the strong compositional control on the mean susceptibility in this layer results from the fractional crystallization and cumulate crystal accumulation that is a characteristic of the layered gabbros, which in turn

Figure 5.9: Graphs showing the variation of mean susceptibility and AMS parameters through Layer A at Somrah, with subdivisions (Zones Ai-iii) based on analysis of thin section point count data.
governs the distribution of secondary magnetite/titanomagnetite (due to the serpentinization and hydrothermal alteration of olivine and clinopyroxene grains respectively). There is also a pronounced variation in fabric type through Layer A, with predominantly prolate AMS ellipsoids in zone Ai and the upper two-thirds of zone Aiii and a region of oblate ellipsoids in the central part of the layer (Figure 5.9). Another distinctive feature is a variation in the inclination of $K_{\text{max}}$ axes, which are steeper in zone Ai than elsewhere in the layer (Figure 5.9). The significance of these variations will be explored in section 5.6.1 below.
5.4 Magnetic fabrics in foliated gabbros

5.4.1 General anisotropy of magnetic susceptibility results

Foliated gabbros were sampled in Wadi Abyad and Wadi Khafifah, and corresponding site mean susceptibility and AMS parameters are listed in Table 5.1. The majority of specimens exhibit oblate AMS ellipsoids with a maximum $T$ of 0.96 (Figure 5.3), although a significant proportion of specimens with prolate ellipsoids occur. Corrected anisotropy degree, $P_J$ (Jelinek, 1981), ranges from 1.01 to 1.27 for individual specimens (Figure 5.3), with a modal value of 1.07 (i.e. 7% anisotropy) that is significantly lower than that in the layered gabbro sections. Again there is no preferred relationship between $P_J$ and $T$, although the majority of samples with prolate ellipsoids have $P_J$ values < 1.15. Unlike the layered gabbros, there is no correlation between $P_J$ and mean susceptibility (Figure 5.3) suggesting no significant compositional control on variations in anisotropy (consistent with petrographic observations outlined in section 5.4.4 that show little variation in modal compositions within the foliated gabbros).

 Orientations of the principal AMS axes at each foliated gabbro site are compared to macroscopic foliation (determined in the field) in the stereographic equal area projections of Figure 5.11 presented in geographic coordinates. Mean ellipsoid principal axes and corresponding confidence ellipses (Jelinek, 1978) are also shown in these projections. As seen in the layered gabbros, the majority of magnetic fabrics within the foliated gabbros are consistent with measured petrofabrics recorded at the sampling sites. As in the layered gabbros, some slight mismatches between magnetic foliations/lineations and orientations of the macroscopic foliation (e.g. at sites WA07 in Wadi Abyad and KFS01 at Khafifah; Figure 5.11) may be due to local variability at the point of sampling compared to the average orientation of the foliation measured over a larger area.
On the basis of the relationships between AMS principal axes and macroscopic foliations, the AMS results in the foliated gabbros may be divided into three groups:

1) Sites with normal fabrics where $K_{\text{max}}$ and $K_{\text{int}}$ axes lie in or close to the plane of macroscopic foliation and where $K_{\text{min}}$ axes lie at or close to the pole to layering, i.e. there is a close match between magmatic and magnetic foliations. In some cases, AMS principal axes form discrete clusters representing triaxial fabrics at the site-level (sites: WA01-03, WA07, WA39-40 and within individual sites in transects across the strike of foliation at Wadi Abyad [sites WA18-37] and Khafifah [sites KF12-27]; see below). Other sites display (semi-)girdle distributions of $K_{\text{max}}$ and $K_{\text{int}}$ axes (sites: WA08, KF02, KFS01) consistent with foliation-parallel, normal, oblate fabrics.

2) Sites where $K_{\text{max}}$ axes lie in or close to the plane of macroscopic foliation and where $K_{\text{int}}$ and $K_{\text{min}}$ axes form (semi-)girdle distributions outside of the plane of layering (sites: KF01, KF10-11). This pattern is consistent with prolate fabrics at site-level, with ellipsoid long-axes lying within the magmatic foliation.

3) Sites with complex fabrics that are only partly related to the observed macroscopic layering (site KFS04) or scattered, poorly defined fabrics (site KFS02). The former displays a predominantly foliation-parallel oblate AMS fabric, but some axis inversion can also be recognised where the $K_{\text{min}}$ axis switches with either $K_{\text{max}}$ or $K_{\text{int}}$. 
Outside of this broad subdivision, site WA05-06 displays a unique fabric with a well-defined magnetic foliation (oblate in appearance) that is orthogonal to the macroscopic magmatic foliation. This site, however, comes from within the transitional zone between sub-horizontal layered gabbros and sub-vertical foliated gabbros) in Wadi Abyad. The petrofabric within this zone is characterized by extreme variability over very small distances displaying complex folding patterns with both moderately (NW plunging) and steeply dipping (E-W striking) limbs and was consequently hard to precisely measure.
This magnetic fabric may, therefore, relate to the steeply dipping limb of the fold structure within the transitional zone.

The magnetic fabrics displayed by the foliated gabbros are generally steeper than those seen in the layered gabbros, reflecting the sub-vertical nature of the petrofabric that dominates this level of the ophiolite. The application of tectonic tilt corrections to these data based on restoring the regional orientation of the Moho to the horizontal results in a further steeping of both the petrofabrics and magnetic fabrics (see below). Mineral lineations also generally correspond well with the trend and plunge of the $K_{\text{max}}$ axes, again highlighting the link between mineral alignment and AMS fabrics.

5.4.2 Detailed variability of AMS fabrics within foliated gabbros sampled at the same structural level (Wadi Abyad and Wadi Khafifah)

In an effort to test predictions related to the formation of the foliated gabbros made by models of lower crustal accretion, detailed sampling of the uppermost foliated gabbros, directly beneath the varitextured gabbros, was undertaken in both Wadi Abyad and Wadi Khafifah (Figure 5.12). In Wadi Abyad, MacLeod & Yaouancq (2000) recognized that the varitextured gabbros at the sheeted dyke-gabbro transition zone represent the fossilized remains of a melt lens (i.e. the axial magma chamber; AMC). Detailed analysis of the foliated gabbros from this critical zone, therefore, aimed to establish the processes involved in magmatic fabric development immediately beneath the melt lens. This is predicted to involve upwards movement of melt in the melt migration model or downward flow in models involving crystal subsidence from the AMC.
Figure 5.12: Simplified geological maps showing the location of the transects sampled within foliated gabbros (immediately beneath the varitextured gabbros/fossil axial magma chamber horizon) at Wadi Abyad and Khaffah.

In order to collect a sufficiently representative quantity of material from both localities, sampling was conducted along two transects perpendicular to the strike of the measured foliation and at the same pseudo-stratigraphic level (Figure 5.13 and 5.14). The Wadi Abyad foliated gabbro transect (Figure 5.13), sites WA18-37, was along an approximately 50 m section sampling two different gabbroic facies identified at the locality: a fine-grained gabbro and a coarse-grained granular gabbro (Figure 5.15a). The Wadi Khaffah foliated gabbro transect (Figure 5.14), sites KF12-27, was an approximately 70 m section sampling two distinct gabbroic facies: an equigranular, fine-grained gabbro and a coarser-grained, olivine-rich gabbro (Figure 5.15b).
Figure 5.13: Map showing the location of sampling sites along a transect across the foliated gabbros in Wadi Abyad. Foliation orientation = 72/254 (measured at site WA21, but consistent across the sampled area). Green sites = coarse-grained facies; red sites = fine-grained facies. Grid lines every 10 m.

Figure 5.14: Map showing the location of sampling sites along a transect across the foliated gabbros in Wadi Khafifah. Foliation orientation = 35/277 (measured at site KF25, but consistent across the sampled area). Green sites = coarser-grained olivine-rich gabbro; red sites = equigranular fine-grained gabbro. Grid lines every 10 m.
Figure 5.15: Field photographs of foliated gabbro facies sampled along: (a) the Wadi Abyad and; (b) the Khaffah transects. KF20 = coarser-grained, olivine-rich gabbro; KF21 = equigranular, fine-grained gabbro.

AMS results from the individual sampling sites within the Wadi Abyad transect show consistency across the transect, with southerly plunging $K_{\text{max}}$ axes (average orientation = 40/185) that cluster on (or very near to) the foliation plane, $K_{\text{int}}$ axes within or close to the plane of the foliation, and $K_{\text{min}}$ axes that cluster near to the pole of the foliation plane (Figure 5.16). Foliation-parallel, “normal”, magnetic fabrics dominate with a predominately triaxial nature, with the exception of site WA36 (Figure 5.16b) that shows a prolate fabric. Sites with green labels in Figure 5.16 were sampled in the coarse-grained gabbro facies, whereas those with red labels come from the fine-grained facies. However, no variation in the site-level magnetic fabrics can be recognised between the two facies.
Figure 5.16a: AMS data from foliated gabbros across the transect in Wadi Abyad, showing orientations of specimen AMS principal axes. Dashed great circles = orientation of macroscopic foliation observed in the field. See Figure 5.13 for base map showing relative locations of sites.
Figure 5.16b: AMS data from foliated gabbros across the transect in Wadi Abyad, showing orientations of specimen AMS principal axes. Dashed great circles = orientation of macroscopic foliation observed in the field. See Figure 5.13 for base map.
Figure 5.17a: AMS data from foliated gabbros across the transect in Wadi Khaffah, showing orientations of specimen AMS principal axes. Dashed great circles = orientation of macroscopic foliation observed in the field. See Figure 5.14 for base map showing relative locations of sites.
Figure 5.17b: AMS data from foliated gabbros across the transect in Wadi Khaffifah, showing orientations of specimen AMS principal axes. Dashed great circles = orientation of macroscopic foliation observed in the field. See Figure 5.14 for base map showing relative locations of sites.
Magnetic fabric data from the Khafifah transect also reveals $K_{\text{max}}$ and $K_{\text{int}}$ axes that are well clustered and plot within the foliation plane at an individual site-level (Figure 5.17), but the plunge and azimuth of these axes varies considerably across the length of the transect such that the individual site clusters of $K_{\text{max}}$ and $K_{\text{int}}$ axes define a girdle distribution along the magmatic foliation plane at the transect-level. The $K_{\text{min}}$ axes, however, are well clustered at both site- and transect-level, consistently plotting as the pole to the plane of the foliation and plunging about 30° to the east (Figure 5.17). All sampling sites display foliation-parallel, “normal” magnetic fabrics, predominately with triaxial fabric patterns. However, sites KF21 and KF24 (Figure 5.17b) show a more oblate fabric with semi-girdle (KF21) and girdle (KF24) distributions of the $K_{\text{max}}$ and $K_{\text{int}}$ axes. As in the Wadi Abyad transect, there is no relationship between the observed magnetic fabrics from each sampling site and the different gabbroic facies (red labelled sites = equigranular fine-grained gabbro and green = coarser-grained olivine-rich gabbro).

The significance of the different styles of variation in AMS fabrics across these two transects will be addressed in section 5.6.3 below.

5.4.3 Anisotropy of partial anhysteretic remanence (ApARM) results

ApARM measurements were also carried out on a selection of samples from both foliated gabbro transects to allow a comparison with AMS fabrics. Experimental work was again carried out on the SushiBar system at LMU, Germany and also at Plymouth University using a biasing DF of 0.05 mT within a coercivity window of 60-90 mT (Figure 5.18). The pARM fabrics recorded across both transects are clearly consistent with the AMS fabrics described above (compare Figure 5.18 to Figure 5.16 and 5.17), which again suggests
that: (i) the AMS signal is either dominated by magnetite or that magnetite and paramagnetic (silicate) fabrics are coaxial in lower susceptibility (more magnetite-poor) samples; and (ii) that little or no SD magnetite grains are present within these rocks given the absence of inverse fabrics, suggesting that the magnetic carriers within the foliated gabbros are either PSD to MD magnetite grains.

Figure 5.18: Equal area stereographic projections showing principal axes of ApARM ellipsoids of samples collected across transects in the foliated gabbros at (a) Wadi Abyad and (b) Khafifah.

5.4.4 Petrographic analyses

AMS results from both the Wadi Abyad and Khafifah foliated gabbro transects, sampled in the highest-level foliated gabbros directly beneath the varitextured gabbros, revealed sub-vertical to moderately dipping magnetic fabrics at this pseudo-stratigraphic level. A focused magnetic fabric, consistent across the entire outcrop was observed in the Wadi Abyad transect, whereas a more diffuse fabric was recorded within the Khafifah transect that not only showed variability between each of the individual transect sampling sites but also within sampling sites themselves. These results, therefore, provide
interesting evidence regarding magmatic and melt transportation processes taking place beneath the high-level melt lens (now represented by the varitextured gabbros). Thin-sections (14 in total) were, therefore, prepared from samples collected from both foliated gabbro transects (five thin-sections from Wadi Abyad and nine from Khafifah consisting of both fine and coarse-grained facies type) in order to provide a direct comparison between the petrofabric seen in the field, the silicate fabric observed under the microscope and the magnetic fabrics measured during AMS analysis. The orientation of elongate minerals, such as plagioclase and clinopyroxene (and sometimes olivine), relative to the orientation of the $K_{\text{max}}$ axis was also measured in an effort to distinguish between the focused fabric of the Wadi Abyad transect and the more diffuse/variable fabric of the Khafifah transect.

Overall, the foliated gabbros were observed to be of similar appearance and mineralogical composition to the layered gabbros from Somrah. However, the percentages of plagioclase, clinopyroxene and to some extent olivine remained more consistent across the transects due to the lack of compositional layering. A far more homogenous mineral composition was, therefore, observed for this level in the crustal sequence with a ratio of approximately 60% plagioclase, 40% clinopyroxene and 10% olivine (±10% for all three phases). The olivines are highly fractured as a result of serpentinization and filled with magnetite. The fractures themselves commonly run parallel to the silicate fabric (as defined by the predominant crystallographic alignment) and are, therefore, usually parallel to the long axes of the olivine grains. Very small, more cubic-shaped opaques can also be seen within the cleavage planes of the clinopyroxenes.
Figure 5.19 and 5.20 present photomicrographs of four swathes across thin-sections from the Wadi Abyad and Khafifah transects respectively. These slices have been photographed orientated relative to the $K_{\text{max}}$ axis of the individual core (in specimen coordinates) from which the thin-section was prepared. A strong alignment of elongate plagioclase and clinopyroxene grains (and the olivines) running parallel with the long edge of the photomicrographs (i.e. the orientation of $K_{\text{max}}$) can clearly be observed in the majority of these thin-sections. This fabric, defined by the alignment of the silicate phases, is clearly of magmatic origin and was formed in melt-present conditions as indicated by the lack of crystal plastic deformation, the presence of magmatic microstructures and the strong evidence of flow between grains, as seen most strikingly in sample KF2303B (Figure 5.20 & Figure 5.21).

A number of grains are, however, orientated slightly oblique (and in some extreme cases almost perpendicular) to the dominant crystallographic alignment and also, therefore, to the orientation of the maximum susceptibility axis. The angle of individual silicate grains, relative to the trend of the $K_{\text{max}}$ axis, was measured in the eight thin sections presented in Figure 5.19 (Wadi Abyad) and Figure 5.20 (Wadi Khafifah) in order to quantify the dispersion of these grains relative to the $K_{\text{max}}$ axes (and hence to the average crystallographic alignment).

The results (Figure 5.22; Table 5.3) reveal significant differences in the variability of silicate grain orientations in the foliated gabbros from the two transects. On average, median angular deviations in the Wadi Khafifah samples ($25^\circ$) are ~70% higher those in the Wadi Abyad samples ($15^\circ$), with 75% of grains deviating by up to $41^\circ$ compared to up to $27^\circ$ in Wadi Abyad. These variations in silicate grain orientations at thin section (i.e. individual sample)
scale are averaged out by AMS measurements on single specimens, since the
AMS fabrics represent the net, average orientation of all mineral grains.
However, the overall pattern of AMS fabrics also varies between localities at the
transect-scale, with clustered principal axes in Wadi Abyad that are consistent
both between samples at each site and between different sites, in contrast to
clustered $K_{\text{max}}$ and $K_{\text{int}}$ axes at each site in Wadi Khafifah that define a girdle
distribution at the transect-level. The significance of these differences for
models of crustal accretion is discussed in section 5.6.3 below.
Figure 5.19: Photomicrograph of swathes across four thin-sections from the Wadi Abyad transect, with $K_{\text{max}}$ aligned along the length of the images.
Figure 5.20: Photomicrograph of swathes across four thin-sections from the Khafifah transect, with $K_{\text{max}}$ aligned along the length of the images.
Figure 5.21: Photomicrographs of thin section from sample KF2303B at Wadi Khafifah, showing details of magmatic, anastomosing flow fabrics in foliated gabbros.
Figure 5.22: Obliquity of grain long axes relative to the orientation of $K_{\text{max}}$ axes from: (left) the Wadi Abyad transect; and (right) the Wadi Khafifah transect.

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<td>KF2401Ci</td>
<td>23°</td>
<td>40.5°</td>
</tr>
<tr>
<td>KF2303B</td>
<td>20°</td>
<td>35°</td>
</tr>
<tr>
<td>KF2201B</td>
<td>28°</td>
<td>40°</td>
</tr>
<tr>
<td>KF1403B</td>
<td>30°</td>
<td>48°</td>
</tr>
<tr>
<td><strong>Mean values:</strong></td>
<td><strong>25.25°</strong></td>
<td><strong>40.9°</strong></td>
</tr>
</tbody>
</table>

Table 5.3. Summary of statistics relating to angular deviation of silicate crystal long-axes from $K_{\text{max}}$ axes in oriented thin sections of foliated gabbros
5.5 Magnetic fabrics in varitextured gabbros and the dyke-rooting zone

The varitextured gabbros were sampled in the northern part of the Wadi Abyad crustal section at two sites, WA13 and WA15, and although only weak magmatic fabrics could be seen in the field, clear magnetic fabrics are present (Figure 5.23 and Table 5.1). Site WA13 has a triaxial magnetic fabric with a steeply dipping cluster of $K_{\text{max}}$ axes that plunges towards the NW, similar to the orientation of a weak, steeply plunging mineral alignment seen in the field. A steeply dipping, NE-SW striking, magnetic foliation can be defined by a great circle that intersects the $K_{\text{max}}$ and $K_{\text{int}}$ axes. A triaxial magnetic fabric is also seen at site WA15; here a weak, steeply dipping NE-SW striking foliation was recorded and both the $K_{\text{max}}$ and $K_{\text{int}}$ axes lie within this plane with the $K_{\text{min}}$ axes

![Figure 5.23. Summary of AMS results from the varitextured gabbros, associated dykes and the dyke-rooting zone in Wadi Abyad (n.b. the foliation was not measurable at site WA13).](image-url)
plotting orthogonally as the pole to the plane. A steep (~80°) mineral lineation measured at site WA15 also corresponds well to the steeply plunging $K_{\text{max}}$ axes at this site. Additionally, after the application of a tectonic tilt correction that returns the Moho to the horizontal, the steep dip of the magnetic fabrics in the varitextured gabbro is maintained.

The varitextured gabbros are in places cut by discrete dolerite dykes, sampled as sites WA14 and WA16, and pass up into the dyke-rooting zone (MacLeod and Yaouancq, 2000), from which a single dolerite dyke was sampled at site WA17 (Figure 5.23 and Table 5.1). These dykes have NNE-NE strikes and all have very similar AMS fabrics (Figure 5.23), characterised by clusters of $K_{\text{min}}$ axes that lie close to the dyke poles and $K_{\text{max}}$ and $K_{\text{min}}$ axes that define semi-girdle distributions along or close to the dyke planes.

The significance of these fabric relationships is discussed in section 5.6.3 below.

5.6 Implications of fabric analyses for lower crustal accretion processes

5.6.1 Insights into magmatic processes in layered gabbros

AMS fabrics in the layered gabbros are dominated by the signal from magnetite, with significant contributions from paramagnetic clinopyroxene crystals only in those samples with low susceptibilities (<5.0 x 10^{-3} SI). Interstitial magnetite in these rocks is very rare, and most magnetite is either of secondary origin along serpentinized fractures in olivine crystals or exists as exsolved inclusions in clinopyroxene and plagioclase. However, it has been shown that the distribution anisotropy of the magnetite grains is coaxial with the preferred orientation of the primary silicate crystals and therefore may be used as a proxy for the silicate magmatic fabric. Similar relationships have been
previously reported in the limited literature on AMS in Oman gabbros, from Wadi Abyad (Yaouancq and MacLeod, 2000) and Wadi Sadm in the northern part of the ophiolite (Kawamura et al., 2005), but also in similar rocks outside of Oman. For example, Abelson et al. (2001) reported coaxial magnetic and silicate fabrics in the layered gabbros of the slow-spreading rate Troodos ophiolite and used these to infer that magma flow in the lower crust was directed along the spreading axis towards the South Troodos Transform Fault Zone. Similarly, Richter et al. (1996) established parallelism between magnetic fabrics and macroscopic foliations in high-level gabbros recovered by scientific ocean drilling at Hess Deep. Coaxial magnetic and magmatic fabrics have also been reported from non-oceanic layered gabbros, for example in the Bushveld Complex of South Africa, where Feinberg et al. (2006) found AMS fabrics developed during syn-magmatic deformation caused by a distribution anisotropy in plagioclase-hosted magnetite inclusions.

Formation of the layered gabbros is considered to either occur by *in situ* intrusion of sills in the lower crust (i.e. the multiple-sills, hybrid/dual-feeding, and melt migration models; Kelemen et al., 1997; Boudier et al., 1996; MacLeod & Yaouancq, 2000) or by crystal settling and accumulation in the high-level axial magma chamber (AMC) followed by wholesale crystal subsidence downwards (i.e. the gabbro glacier model; e.g. Quick & Denlinger, 1993). Formation of a lower crustal sill complex may be expected to produce planar, margin-parallel magnetic fabrics in constituent, individually emplaced melt layers (or potentially fabrics that are imbricated against layer margins, similar to those documented in dykes; Knight & Walker, 1988; Staudigel et al., 1992; Tauxe et al. 1998). Crystal settling at the base of the AMC in the absence of magmatic flow should form oblate, planar fabrics, analogous to the depositional fabrics seen in sedimentary
rocks, with clustered $K_{\text{min}}$ axes orthogonal to the layering and $K_{\text{max}}$ and $K_{\text{int}}$ axes forming girdle distributions parallel to layering with no preferred azimuth. Subsequent downward subsidence of crystals in a gabbro glacier system should modify these fabrics along flow trajectories that vary systematically with depth. Hence, the petrofabric proxy provided by magnetic anisotropy may provide insights into the applicability of the various accretion models in terms of their predicted magmatic and deformation fabrics.

Layered gabbros analysed in this study have been shown to have magnetic fabrics that are broadly layer-parallel, with maximum principal susceptibility axes that are parallel to the preferred orientation of non-equant silicate crystals. Examination of Figure 5.4 shows that $K_{\text{max}}$ axes within and between sites are not randomly oriented, but instead appear to be consistent in broad orientation, even between sampled sections that are geographically separated on length-scales of up to 100 km. This consistency becomes more apparent when AMS principal axes from all layered gabbro samples are plotted on the same stereographic projections (Figure 5.24). These show a strong ENE-WSW clustering of $K_{\text{max}}$ axes (Figure 5.24a, left hand stereonet), that becomes even more pronounced when data are rotated into a common horizontal plane of magmatic layering by applying site-level standard tilt corrections (Figure 5.24a, right hand stereonet). $K_{\text{min}}$ axes also show a regional consistency in this reference frame (Figure 5.24b) This consistency of $K_{\text{max}}$ axes at the regional-level is inconsistent with localised controls on the formation of layering (for example, convective circulation or deposition by density currents on to the base of a magma chamber or melt lens), suggesting that either large-scale controls on magmatic flow processes existed to produce regionally-
Figure 5.24. Equal area stereographic projections showing (a) AMS maximum and (b) minimum principal axes from all layered gabbros sites sampled in this study (Wadis Abyad, Khafifah, Somrah and Nassif). Tilt corrected data produced by restoring observed magmatic layering at each site to the horizontal using a standard tilt correction. Primitives on right hand projections therefore correspond to the plane of magmatic layering. Note the consistent ENE-WSW orientation of $K_{\text{max}}$ axes across all sampling localities, suggesting regional-scale controls on AMS fabrics in the layered gabbros.
consistent, preferred mineral alignments or large-scale deformation processes operated to control or modify fabric development.

This regional pattern can be placed into context by considering the large-scale structure of the ophiolite deduced from field mapping and tectonic reconstructions. A synthesis of research conducted by the international community working on the Semail ophiolite was provided by Nicolas et al. (2000), resulting in development of an overall, integrative structural map of the ophiolite. After being reassembled in a best geometrical and structural fit, data from the various massifs revealed the presence of a 40-50 km-wide, over 200 km long, younger ridge segment in the southern ophiolite massifs which is presently oriented NW-SE and which propagated into 1-2 My older lithosphere oriented NE-SW (Nicolas et al., 2000; see Figure 3.3). All of the sections sampled in the present study lie within this domain of the ophiolite and formed by spreading along the NW-SE propagating ridge, and all lie to the NE of the inferred position of this ridge. The overall ENE-WSW preferred orientation of $K_{\text{max}}$ axes (Figure 5.24) is nearly orthogonal to this palaeo-ridge system, suggesting that it may be due to consistent magmatic flow away from the spreading axis, drag-induced shearing related to turn-over of the asthenospheric mantle beneath the spreading axis and divergence away from the ridge, or a combination of these mechanisms.

Further insights into potential modes of fabric development on a much smaller, individual layer-scale is provided by the detailed sampling in Layer A at Somrah, which reveals a striking variation in anisotropy parameters through the layer. Corrected anisotropy degree reaches a peak of 1.20-1.27 in the central part of the layer, with lower anisotropies of 1.05-1.15 near the base and top (see Figure 5.9). Moreover, there is a distinct pattern of oblate fabrics in
samples from the central portion flanked by prolate samples above and below (Figure 5.25), suggesting different mechanisms of fabric development. Oblate fabrics in the centre of the layer could either imply gravity controlled crystal accumulation or compaction related to overlying stress. Geochemical considerations indicate that compaction and expulsion of residual melt must have occurred in the layered gabbro sequences, as they represent cumulates that are in rare earth element and Mg-Fe equilibrium with the more evolved melt that formed the upper gabbros, sheeted dykes and lavas (Kelemen et al., 1997).

In contrast, the predominately prolate fabrics at the base (zone Ai) and top (upper part of zone Aiii) of the layer are arranged around the observed

![Graph](image)

Figure 5.25: Variation in the inclination (dip) of $K_{\text{max}}$ axes through Layer A of the layered gabbros at Somrah. Note that mean inclination is shown where more than one sample was collected at the same height.
macroscopic magmatic lineation at this site (Figure 5.26) and suggest a
crystallographic alignment produced by magmatic flow or by shearing.
Importantly, the inclination of \( K_{\text{max}} \) axes also varies across the layer (Figure 5.25), with prolate ellipsoids in zone Ai aligned at a significantly oblique angle (mean obliquity = 10°) to the plane of layering. AMS ellipsoids in the upper prolate zone are also subtly oblique (mean obliquity = 2°). These relationships are best visualised by applying sequential rotations to the AMS data to take the plane of magmatic layering to a vertical E-W orientation and the magmatic lineation to the vertical (Figure 5.26). This shows the clear obliquity of \( K_{\text{max}} \) axes near the base of the layer and the subtle yet consistent obliquity in the upper part. As discussed earlier, the orientation of the \( K_{\text{max}} \) axis within individual samples mimics the orientation of the predominant silicate fabric. Therefore, the

Figure 5.26: Equal area stereographic projections showing the orientation of \( K_{\text{max}} \) axes in samples from Layer A at Somrah, relative to magmatic layering and magmatic lineation. (left) In situ coordinates, showing distribution of \( K_{\text{max}} \) axes centred on the magmatic lineation and straddling the plane of magmatic (modal) layering; (right) same data (with samples from the middle of the layer omitted for clarity) after rotating the plane of layering and the magmatic lineation to vertical, i.e. looking down the lineation in the direction of the \( K_{\text{max}} \) axes. Note systematic difference between \( K_{\text{max}} \) axes from the lower and upper parts of the layer, suggesting obliquity of crystal orientations relative to the layer margins (especially in near the base of the layer).
obliquity of the $K_{\text{max}}$ inclinations within the upper and lower zones relative to the magmatic layering can be interpreted in terms of obliquity of silicate crystals along the margins of the layer.

These types of fabrics cannot be produced by downward subsidence and deformation of crystals formed in an axial magma chamber, as envisaged by the gabbro glacier model. However, the observed obliquity may potentially be explained if Layer A formed by emplacement in a sill complex, consisting of a sequence of layers, each of which may be considered to represent a discrete magmatic event. For instance, imbrication of crystals along the margin of a layer emplaced by magmatic flow may occur and be preserved during crystallization, with opposing obliquity along the margins providing information on flow direction (Figure 5.27). This relationship has been detected by AMS analysis in several studies of dykes (including ophiolitic sheeted dykes in Troodos; Staudigel et al., 1992). However, in dykes the zone of imbricated fabrics usually extends only a few centimetres in from the dyke margins (Tauxe et al., 1998),

![Diagram](image.png)

Figure 5.27: Schematic illustration of the theoretical arrangement of magnetic fabric axes that may result from emplacement of a melt layer. The obliquity of fabrics relative to layer margins as well as the trend of the $K_{\text{max}}$ axes may potentially be used to infer emplacement direction (Morris, unpublished diagram).
whereas the prolate fabrics displaying significant obliquity in Layer A are much thicker (Figure 5.25). In addition, any model for the distribution of fabrics in this layer must also account for the mineralogical grading, with olivine crystals concentrated near the base and more plagioclase and clinopyroxene-rich upper zones.

Three potential scenarios for producing obliquity (principally at the base of a layer) are illustrated in Figure 5.28. All require emplacement of a magma that already contains olivine phenocrysts to be delivered from a melt reservoir within the mantle transition zone or some other deep melt lens. The first involves solid state, simple shear deformation after gravitational settling of olivine at the base and complete solidification of the layer (Figure 5.28, upper diagram). Shearing could be driven by mantle drag from below, resulting in transposition of fabrics in the overlying gabbros. This could create shear fabrics oblique to the margins of a layer or package of layers, but with crystals at opposing margins sharing similar dip directions. However, the absence of pervasive shear deformation in Layer A, the lack of crystal deformation within the layer and the presence of oblate fabrics in its centre all exclude this mechanism. Shearing (e.g. due to large-scale mantle drag) could also result in a marked difference in fabric orientations along opposite margins of a layer in a sill complex that, following initial melt emplacement and gravitational settling of olivine phenocrysts, consists of a lower zone with a partially solidified crystal framework and an upper melt-dominated zone of low viscosity (Figure 5.28 middle diagram). In this scenario, drag at the base of the layer may reorient olivine crystals to form a significantly oblique fabric in the lower zone, while only weak or no imbrication would occur at the top. In this model, the dip direction of the oblique fabric at the base would indicate the direction of drag from below.
In the case of Layer A, this would indicate a bottom to the east displacement. Shearing-induced reorientation of silicate crystals would need to take place in the presence of significant residual melt, however, there is no thin section evidence for deformation of individual crystals. The lack of grain-scale deformation is also evidenced by unpublished pilot electron backscatter diffraction analyses of thin sections from Layer A (Moyle, 2015), which show complete lack of sub-grain development. Following formation of shear-induced obliquity of crystals at the base of the layer, subsequent gravitational flattening and/or compaction (pure shear shortening) could then induce oblate fabrics in the centre of the layer which would be the last zone of the layer to crystallize and may remain weak because of presence of residual interstitial melt. Finally, a scenario similar to that illustrated in Figure 5.27 involving imbrication of crystals during magmatic flow associated with layer emplacement could also
result in the fabrics seen in Layer A, but would require near simultaneous accumulation and imbrication of early-formed olivine crystals, followed by less pronounced imbrication of dominantly clinopyroxene and plagioclase crystals along the upper margin, possibly as flow slowed following initial injection (Figure 5.28 lower diagram). In the case of Layer A, the opposing obliquities adjacent to lower and upper margins would indicate magmatic flow to the west. Subsequent compaction could again result in the observed oblate fabrics in the layer centre. In both of the latter models, compaction-related vertical shortening might lead to expulsion of residual interstitial melt that would likely follow the pre-existing silicate framework and hence maintain a preferred orientation of magmatic and magnetic lineations.

With the important caveat that only limited significance can be placed on study of a single layer, the sense of magmatic flow or basal shear inferred from these models when applied to Layer A might provide a means of distinguishing between alternate large-scale, spreading-related processes discussed in relation to the regional consistency of \( K_{\text{max}} \) axes. Imbrication along the margins of the layer during emplacement-related magmatic flow would imply flow inwards towards the palaeoridge, whereas drag-induced shearing would imply displacement away from the ridge, which may be easier to reconcile with the regional spreading fabric. However, clearly more detailed sampling of many more layers would be needed to reach more definitive conclusions.

Opportunities to compare these results with published work are restricted, as only two previous magnetic fabric studies have been conducted in the Semail ophiolite. In the first study, Yaouancq & MacLeod (2000) examined the relationship between magnetic fabrics and petrofabrics in Wadi Abyad. They found that the majority of their layered gabbro specimens had “normal”
magnetic fabrics with $K_{\text{min}}$ axes perpendicular to the petrofabric and $K_{\text{max}}$ axes parallel to mineral lineation. This is consistent with the more extensive dataset reported here, demonstrating that normal magnetic fabrics dominate throughout the lower oceanic crust. Yaouancq & MacLeod (2000) also investigated the extent to which magnetic fabric was controlled by petrofabric by measuring the orientation of serpentine veins within serpentinized olivines. They found that a strong correlation exists between the orientation of the serpentine fractures and the whole rock fabric, again in agreement with the conclusions of this study. As in this study, they found that plagioclase crystallographic fabrics correlated strongly with AMS principal axes, but concluded that the magnetic fabric was dominantly controlled by the distribution of secondary magnetite along serpentinized fracture planes in olivine. This secondary fabric was shown to be controlled by the orientation of the primary magmatic petrofabric that results in strong crystallographic alignment of (predominately) plagioclase and clinopyroxene crystals (Yaouancq & MacLeod, 2000). The conclusions of Yaouancq & MacLeod (2000) are in full agreement with the findings of this investigation.

Kawamura et al. (2005) examined the origin of layering in cumulate gabbros in Wadi Sadm in the northern part of the ophiolite, using AMS across a limited section of gabbros in order to test whether they formed by pure shear, simple shear or by gravity controlled deposition. Kawamura et al. (2005) sampled a series of melanocratic layers (several tens of centimetres thick composed of olivine ± clinopyroxene and orthopyroxene) and leucocratic layers (several metres thick with clinopyroxene ± olivine-rich bases that grade upwards, becoming dominated by plagioclase), similar in appearance and mineralogy to the Somrah section. The bases of melanocratic layers are sharp, whereas the
contacts of leucocratic layers with the underlying melanocratic layers are more gradual (Kawamura et al., 2005). AMS results obtained by Kawamura et al. (2005) reveal vertical to sub-vertical $K_{\text{min}}$ axes and shallowly plunging $K_{\text{max}}$ and $K_{\text{int}}$ axes (after correction to horizontal of the local dip), which compares well with results obtained in this study from the Somrah layered gabbros (see result for sites SR01-05 in Figure 5.4). The mean susceptibility profiles produced by Kawamura et al. (2005) for the Wadi Sadm layered gabbros are also comparable with results from Layer A at Somrah, with higher mean susceptibility values in the olivine and clinopyroxene-rich lower melanocratic parts of layers (equivalent to zone Ai in Layer A) and lower susceptibility values in the plagioclase-rich upper leucocratic zones (equivalent to zone Aiii in Layer A). Kawamura et al. (2005) concluded that the orientation of the $K_{\text{max}}$ axes is strongly controlled by the rock fabric itself (i.e. the crystallographic alignment) as was previously established by Yaouancq & MacLeod (2000). Finally, Kawamura et al. (2005) suggest that the layered gabbros are formed primarily by the downward cyclic accumulation of cumulate crystals (olivine followed by clinopyroxene and then plagioclase) in sill-like magma chambers following in situ emplacement. Subsequent simple shear, as a result of either magmatic flow or ductile deformation and evidenced by gentle grain imbrication of plagioclase and clinopyroxene relative to the foliation plane, then generates the strong crystallographic alignment (that is aligned parallel to the shear direction) (Kawamura et al., 2005). The results and conclusions of Kawamura et al. (2005) regarding the formation of layered gabbros are, therefore, wholly consistent with those developed in this investigation.
5.6.2 Can magnetic anisotropy be used to determine strain variations in the lower oceanic crust?

A key prediction of the gabbro glacier model (e.g. Quick & Denlinger, 1993) is that the lower crust forms by subsidence of a crystal mush from the base of an axial magma chamber. Subsidence of a semi-solid crystalline mush following crystallization in a high-level melt lens at the base of the sheeted dyke complex would theoretically lead to a near-exponential increase in strain with depth in the lower oceanic crust (Figure 5.29) as a result of ductile flow down through the crust (e.g. Henstock et al., 1993; Phipps Morgan & Chen, 1993; Nicolas et al., 2009). Material at the base of the oceanic crust should, therefore, exhibit more strain and potentially display greater shape and crystallographic alignment than material just beneath the high-level melt lens. Testing this key prediction of the gabbro glacier subsidence-type model may potentially be possible using magnetic analyses because AMS responds systematically to

Figure 5.29: The variation with depth through the lower crust of several key quantifiable parameters as predicted by the gabbro glacier (orange line) and multiple-sills (green dashed line) models (modified from a figure designed and drafted by Roz Coggan).
strain, allowing the degree of anisotropy to act as a proxy for strain intensity in ideal circumstances. For example, Borradaile & Alford (1987) conducted deformation experiments on synthetic sandstone samples containing magnetite that demonstrated a clear linear relationship between measured strain and anisotropy degree (Figure 5.30). More recently, Ferré et al. (2014) reviewed the development of magnetic fabrics in ductile shear zones, and noted that a number of studies have shown a positive correlation between corrected anisotropy degree (P_J) and strain, suggesting that AMS has great potential as a tool for investigating large strain gradients.

In order to determine whether AMS provides information of relevance to testing the gabbro glacier model, P_J values obtained in this study across a range of vertical scales through the lower crust (layered and foliated gabbros) are presented in Figure 5.31. This figure shows the variation in P_J with: (1) height above the crust-mantle boundary in the Wadi Khafifah crustal section; (2) height above the crust-mantle boundary in the Wadi Abyad crustal section; (3) through Layers A-D of the Somrah layered gabbros; and (4) through Layer A at

![Figure 5.30: The relationship between strain and AMS during experimental deformation of a magnetite-bearing synthetic sandstone (Borradaile and Alford, 1987).](image-url)
Figure 5.31: Vertical variations in corrected anisotropy degree at a range of scales.
Somrah. In all cases and at all scales there is no systematic increase in anisotropy with depth in the lower oceanic crust, with $P_J$ ranging between ~1.05 and 1.3 both within a single layer and at the scale of the entire lower crust. This could be taken as evidence for absence of the exponential increase in strain with depth predicted by the gabbro glacier model. However, use of AMS as a strain marker is subject to a number of caveats discussed by Ferré et al. (2014). The most important of these in the context of the data presented in Figure 5.31 is that variations in $P_J$ cannot be solely attributed to variations in strain if there is a mineralogical control on the source of the AMS signal. At least in the case of the layered gabbros sampled here, this is evidently the case as $P_J$ has been shown to vary as a function of mean susceptibility (see Figure 5.3 and 5.10), making the separation of the mineralogical and strain controls on AMS impossible. The only way to use variations in $P_J$ to test strain variations in these mineralogically heterogeneous lower crustal lithologies would be to systematically collect samples through a crustal section from very plagioclase-rich, olivine-poor parts of layers (to minimise the effects of secondary magnetite produced during serpentinization). This is not practically feasible in the field because of the obscuring effects of desert varnish on exposed sections, but may be possible in fresh drill-core samples to be collected during the forthcoming Oman Drilling Project.  

5.6.3 Significance of fabrics within the varitextured and foliated gabbros for crustal accretion processes

The varitextured gabbros are now generally considered to be the fossilized remains of the axial magma chamber (AMC) and as such represent a

1 https://www.ldeo.columbia.edu/gpg/projects/icdp-workshop-oman-drilling-project
zone of melt accumulation at the top of the plutonic sequence (MacLeod & Yaouancq, 2000). In the gabbro glacier model (e.g. Quick & Denlinger, 1993), crystallization takes place within the AMC and accumulates at the base of the melt lens before subsiding to form the lower crust. If correct, this would imply development of planar, sub-horizontal fabrics in the varitextured gabbros. Alternatively, in the melt migration model of MacLeod & Yaouancq (2000), the AMC is solely a site of melt accumulation and mixing and the varitextured gabbros should, therefore, display either a random fabric or have steeply inclined fabrics due to melt migration to feed the overlying sheeted dykes and lavas. The steeply dipping NE-SW striking magnetic fabric displayed by the varitextured gabbros, along with the consistency of their $K_{\text{max}}$ axes, suggests the presence of a weak, steeply orientated macroscopic fabric in these rocks, with no evidence for sub-horizontal crystal fabrics as predicted by the gabbro glacier model. The orientation of these fabrics is sub-parallel to both the trend of the dykes in the overlying dyke-rooting zone and those sampled within the varitextured gabbros (see section 5.5). Together these uppermost gabbro and dyke sites share common, approximately NE-striking, moderate-steeply magnetic foliations. A similar situation has been reported from fast-spread, East Pacific Rise crust exposed on the seafloor of the Hess Deep Rift. Here, westward propagation of the Cocos-Nazca spreading axis towards the East Pacific Rise has resulted in faulting that exposes a complete crustal section that has been the subject of numerous submersible studies and two ODP/IODP drilling expeditions (Mével et al., 1996; Gillis et al., 2013). Richter et al. (1996) report magnetic fabric data from the uppermost gabbros of the Hess Deep section sampled at ODP Hole 894G during ODP Leg 147 (Mével et al., 1996). Steeply dipping magnetic foliations in the sampled gabbros were found to have
the same orientation as dykes in the overlying sheeted dyke complex (observed during Alvin dives; Karson et al., 1992). This analogy with in situ fast-spread crust in the Pacific indicates that the steep lineations and foliations in the varitextured gabbros in the Wadi Abyad section, that are parallel to the orientation of associated dykes, records the upward flow of melt at the top of the axial magma chamber into the base of a sheeted dike complex (Nicolas et al., 1988; MacLeod & Rothery, 1992).

The gabbro glacier, multiple-sills and hybrid model for lower oceanic crustal accretion (see Chapter 3) all infer that the foliated gabbros are produced by the near vertical downward ductile flow of a crystalline mush from the AMC, with the sub-vertical foliation reflecting the trajectory of the descending material. In contrast, the melt migration model of MacLeod & Yaouancq (2000) predicts that the foliated gabbros are formed in situ and their sub-vertical foliation is the result of upward melt migration towards the high-level melt lens (i.e. AMC) rather than the downward subsidence of a crystalline mush. In the absence of imbricated fabrics, magnetic anisotropy data cannot unequivocally determine directions of magmatic or shear-related flow, but other variations in the style of fabric development can provide insights into this problem. Thin-section analysis of the foliated gabbros again confirms a direct relationship between the orientation of magnetic fabrics and crystal alignments in these rocks, so that AMS $K_{\text{max}}$ axes provide information on the average preferred orientation of all crystals in a sample.

When plotted together, AMS results from the Wadi Abyad and Khafifah transects reveal two fundamentally different magnetic fabric types present at approximately the same pseudo-stratigraphic level of the crustal sequence (Figure 5.32).
There is a consistent magnetic fabric across the entire Wadi Abyad transect, with the same fabric observed within all individual samples at each site (i.e. over approximately a 1m$^3$ area) and between sites distributed across the 50 m length of the transect. This demonstrates a very focused, moderately plunging cluster of $K_{\text{max}}$ axes that plots within the foliation plane, with $K_{\text{min}}$ axes.
close to the pole to the magmatic foliation (Figure 5.32a). Thin section analyses show that the consistency in $K_{\text{max}}$ orientations reflects a consistent linear preferred orientation of silicate crystals.

The Khafifah transect, however, exhibits a distinctly different fabric, with $K_{\text{max}}$ and $K_{\text{int}}$ axes from individual samples forming clusters at the site-level, but with data from all sites collectively defining a girdle distribution across the 70 m transect that parallels the observed foliation, again with tightly clustered $K_{\text{min}}$ axes that lie close to the pole to the magmatic foliation (Figure 5.32b). Thin section analyses demonstrate that $K_{\text{max}}$ axes in individual Khafifah samples average out significant variability in the orientation of silicate crystals at the microscopic scale, with the presence of an anastomosing texture that at a sample-level mimics the girdle distribution seen in the overall AMS results. In

Figure 5.34: Variation in inferred melt transportation styles through the foliated gabbros sections beneath the melt lenses in (a) Wadi Abyad and (b) Wadi Khaffah.
addition, the application of a tectonic tilt correction to the AMS data from both transects that restores the Moho to the horizontal results in an overall steeping of the magnetic fabric (Figure 5.33). The $K_{\text{max}}$ axes from the Wadi Abyad transect become sub-vertical, while the magnetic foliation in the Khafifah transect also becomes steeper.

The consistency of the magnetic fabric and petrofabric within the Wadi Abyad foliated gabbros is taken to represent a zone of focused vertical transportation of melt upwards to feed the high-level melt lens (Figure 5.34). However, a similar fabric could potentially be produced by downward transport.
subsidence of a crystalline mush from the AMC, as in the gabbro glacier model. The more varied fabric observed in the foliated gabbros from Wadi Khafifah, however, is interpreted to represent transportation of melt upwards in discrete anastomosing channels through a semi-solid crystalline mush (Figure 5.34). Flow upwards of melt, in either a focused zone or within discrete channels, is considered more likely in both cases given the pervasive magmatic nature of the foliated gabbros, the total lack of crystal plastic deformation and, in the case of Wadi Khafifah, the fact that gravity driven downward flow would be expected to generate a more focused fabric. These finding are, therefore, taken to imply that the foliated gabbros also formed in situ during the ascent of melt from the base of the crust towards the AMC (Figure 5.35), as proposed by the melt migration model of MacLeod & Yaouancq (2000).
Chapter 6: Palaeomagnetic analyses

6.1 Introduction

The systematic sampling undertaken for magnetic fabric analyses also provided an opportunity to undertake a full palaeomagnetic study of the southern massifs of the Semail ophiolite, with the aim of determining the timing of remanence acquisition and whether relative rotations of crustal blocks has occurred, either during seafloor spreading or during later emplacement of the ophiolite. In particular, systematic sampling through crustal sections in Wadi Abyad and Wadi Khafifah (plus less extensive sampling at Somrah and in Wadi Nassif) allows the extent and timing of the previously reported remagnetization (Feinberg et al., 1999) to be constrained, thereby allowing previously reported palaeomagnetic data to be interpreted with increased confidence. The results of these palaeomagnetic analysis are presented in this chapter.

6.2 NRM intensities

Prior to demagnetization, the intensity of the natural remanent magnetization (NRM) was measured for all specimens (Figure 6.1).

The corresponding geometric mean NRM intensities (taking anti-logs) and ranges are:

- Layered gabbros: mean = 0.53 Am\(^{-1}\); min = 0.01 Am\(^{-1}\); max = 4.83 Am\(^{-1}\)
- Foliated gabbros: mean = 1.66 Am\(^{-1}\); min = 0.05 Am\(^{-1}\); max = 29.0 Am\(^{-1}\)
- Varitextured gabbros: mean = 1.09 Am\(^{-1}\); min = 0.42 Am\(^{-1}\); max = 1.9 Am\(^{-1}\)
- Dykes: mean = 0.06 Am\(^{-1}\); min = 0.001 Am\(^{-1}\); max = 0.80 Am\(^{-1}\)

NRM intensities recorded in this investigation and from previous palaeomagnetic studies on the gabbros and dykes from the Semail ophiolite are
broadly comparable. These studies gave NRM intensities for the gabbros that range from 0.03 to 6 A/m, with a mean of 1-2 A/m (e.g. Luyendyk and Day, 1982; Feinberg et al., 1999; Weiler, 2000; Gee & Meurer, 2002), whereas dykes gave a mean intensity of ~0.6 A/m (Luyendyk et al., 1982). Luyendyk and Day (1982) also observed the same difference between mean NRM intensity values of the lower (i.e. layered) and upper (i.e. foliated) gabbros, with higher NRMs recorded in the foliated gabbros. A similar difference between foliated and layered gabbros was also seen by Swift & Johnson (1982) in the Bay of Islands ophiolite and has been identified by Dunlop & Prévot (1982) from their work on Deep Sea Drilling Project (DSDP) samples.

Figure 6.1: Histograms of NRM intensities for each pseudostratigraphic level sampled in the Semail ophiolite.
6.3 Demagnetization characteristics and selection of magnetization components

Demagnetization gradually destroys the magnetization of a sample to reveal the characteristic remanent magnetization (ChRM) component. The ChRM can, however, sometimes be masked by secondary (younger) magnetizations that must first be eliminated by stepwise demagnetization using progressively increasing levels of either thermal energy or alternating fields. The characteristic responses of individual specimens from Wadi Abyad, Wadi Khaffifah, Somrah and Wadi Nassif to both thermal and AF demagnetization techniques are illustrated in the Zijderveld diagrams of Figures 6.1 to 6.4. These data are presented in tilt corrected coordinates (using the local orientation of the Moho as a palaeohorizontal marker).

Demagnetization results from the Wadi Abyad layered and foliated gabbros (Figure 6.1) generally display very clean single-component decay to the origin with high coercivity/high unblocking temperatures (e.g. WA1201HA, WA0802HA and WA3001Aii), as do samples from a discrete dyke sampled at site WA38 (see WA3801Aii). A low-stability component (which is better defined in the AF demagnetization data) can, however, be observed in about 20% of the specimens (e.g. WA04B07 and WA2002B) and is removed in AF fields of approximately 15 mT or by temperatures of about 400°C. In the layered and lower foliated gabbros this low-stability overprint has a N-directed magnetization (e.g. WA0501). In the higher foliated gabbros (i.e. from the Wadi Abyad foliated gabbro transect) the secondary component is much more randomly orientated (e.g. WA2002B and WA3401Aii) and also displays some anomalous reversely magnetized remanences; these results have, therefore, been omitted from subsequent analyses. These data, predominantly, give N/NW-directed ChRM
with shallow to moderate inclinations and only show negligible directional differences between AF and thermally demagnetized samples. Results from the northern part of the Wadi Abyad crustal section, however, display significantly different remanence directions (e.g. WA1302 and WA1504A). All varitextured
gabbro specimens, those from a discrete dyke intruding into the varitextured gabbros and dyke-rooting zone specimens have ChRMs with moderate inclinations and NE declinations and display clean, linear decay to the origin. About half of these specimens, predominately from dyke sites WA16 and WA17
but some from the varitextured gabbro sites WA13 and WA15, display a secondary component that is only revealed during thermal demagnetization (e.g. WA1504B and WA1704Aii). When present, this secondary component displays consistently more northerly directed remanences than the NE-directed ChRM and is destroyed by approximately 400°C. A number of specimens from site WA16 also display three component magnetizations (e.g. WA1601Aiii) that show a progression from a NW direction (NRM to 350°C), through a NNE-directed magnetization (350-550°C), culminating in a ChRM with a NE-directed remanence (550-590°C).

Demagnetization characteristics of specimens from the Wadi Khafifah area are generally similar to those seen in the Wadi Abyad layered and foliated gabbros (Figure 6.2). Dyke site KF09 has, however, been excluded from further analysis as results showed poor correlation between individual specimens giving very scattered site-level directions. Predominately single-component decay of the magnetization to the origin with high coercivity/high unblocking temperatures is seen and nearly all specimens (with the exception of six) are normally magnetized after a tectonic tilt correction to restore the Moho (which dips 29/173) to horizontal is applied (e.g. KF0406B, KF2303B and KFS0401Aii). Approximately 25% of the specimens (from a total of 200) also display a secondary magnetization component that commonly gives a N-directed remanence (e.g. KF0310H, KF1001Aii and KFS0101Bi). Other samples showed more scattered, secondary magnetizations (e.g. KF1502B, KF1702B and KF1801B), but given their wide scatter these components have been omitted from subsequent analyses. The secondary components are generally removed by AF fields of 15 mT or by temperatures between 350-400°C. The ChRM directions are comparable with results from the Wadi Abyad layered and
Figure 6.3: Characteristic Zijderveld plots of thermal (Th) and alternating field (AF) demagnetization data for specimens from the Wadi Khaffifah crustal section. LG = layered gabbros; FG = foliated gabbros; VG = varitextured gabbros; D = dykes.
foliated gabbros and display a range of moderately inclined, N/NW-directed magnetizations. The more northerly remanences are generally from the layered gabbros (and dyke site KF07), whilst the foliated gabbros display magnetizations that are more NW-directed.

Representative Zijderveld plots of the tilt corrected demagnetization data from Somrah are presented in Figure 6.3. The majority of the layered gabbro specimens display linear, single-component decay to the origin with high coercivity/high unblocking temperatures and show consistently clean decay using both demagnetization techniques (e.g. SR0211HA and SR0506). Six specimens (out of a total of 52) display a stable secondary component that has a NNW-directed magnetization and is destroyed by AF fields of approximately 15 mT and temperatures of 350°C (e.g. SR0209HA). The ChRMs of the layered gabbro specimens are consistently directed towards the NW, with results from thermal demagnetization and AF demagnetization giving similar results.

Specimens from the two dyke sites (SR06 and SR07) generally display a two-
component magnetization (e.g. SR0701Bii) with a N/NNW-directed overprint comparable to the secondary component in the host layered gabbros. The demagnetization is generally clean and linear but AF demagnetization produced slightly noisier results when used on specimens from site SR06 (e.g. SR0602Aii) compared to the thermal demagnetization data. An identical ChRM to the
layered gabbros was also picked for the dyke sites with a NW-directed characteristic remanence.

Demagnetization results from sites WN02, WN03 and WN04 in Wadi Nassif (site WN05 has been disregarded) display single-component decay to the origin with high coercivity/high unblocking temperatures (e.g. WN0307H and WN0401H; Figure 6.4). Thermal demagnetization generally gave cleaner results compared to AF demagnetization (e.g. WN0210H), but all results are of good quality and permitted a stable magnetization component to be picked. The ChRM of specimens from these three sites gave magnetizations with NW/NNW declinations and shallow inclinations. Specimens from site WN01 (from within the MTZ) could only be thermally demagnetized due to their size and gave slightly noisy results but two magnetization components could be picked from the six individual specimens (e.g. WN0103H). The secondary overprint component has a N-directed magnetization with a very steep inclination (70-80°) and is destroyed by temperatures of approximately 350°C. The ChRM of specimens from site WN01 are, however, similar but not identical to those

---

1 The tumbling mechanism in the AF demagnetizer requires near perfectly sized cores.
recorded in the overlying layered gabbro sequence with specimens displaying a shallowly inclined, N-directed magnetization.

6.4 Site-level characteristic remanent magnetizations

Principal component analysis (PCA), using least square best-fit lines (Kirschvink, 1980), was used to identify the individual specimen magnetization components (both ChRM and any secondary components). All picked directions had maximum angular deviations <10°, demonstrating good linearity of the picked components. These were then combined to calculate site mean directions and associated Fisher (1953) statistics.

Table 6.1 lists site mean magnetization directions in geographic and tilt corrected coordinates from each sampling locality in pseudostratigraphic order up section. Apart from the large number of specimens demagnetized from the foliated gabbro transects in Wadi Abyad and Wadi Khafifah (37 and 72 specimens, respectively), an average of 8 specimens was used to calculate each site mean magnetization direction. Of the 47 sampling sites (the two foliated gabbro transects are considered collectively as one site each), all except five sites (WA09, WA11, WA15, KF01, KF02) have α95 values < 10° (with an average value of 5.5°) and Fisher precision parameters (k) > 10, with approximately 75% of sites having k values greater than 100 (average value of 352). These low α95 values and high k values demonstrate the high quality of the demagnetization data obtained in this investigation and the accuracy of the calculated site means. Figures 6.5 to 6.8 show the site mean magnetization directions in geographic and tilt corrected coordinates from each sampling locality on equal-area stereographic projections.

As identified during PCA, a difference in the magnetization directions of the layered and foliated gabbros compared to the upper part of the plutonic
sequence (i.e. the varitextured gabbros and dyke-rooting zone) is clearly apparent in the Wadi Abyad sequence in both geographic and tilt corrected coordinates (Figure 6.5). Site mean magnetizations in geographic coordinates are normally magnetized with moderate to steep inclinations. Application of a tilt correction to remove the dip of the local Moho (dip and dip direction = 32°03′) causes the inclination of the site means to become shallower but they retain
positive inclinations, indicating a normal magnetization (Table 6.1). The foliated gabbros and layered gabbros have N/NW-directed remanences both before and after tilt correction, whereas the varitextured gabbros (WA13 and WA15), discrete dykes sampled from within the varitextured gabbros (WA14 and WA16), and the dyke-rooting zone (WA17) display NE-directed directions with moderately steep inclinations in geographic coordinates that become shallower after tilt correction. The mean directions and Fisher statistics for the Wadi Abyad section are listed in Table 6.2.

These distinct magnetization directions recorded by different pseudo-stratigraphic levels of the lower crust in Wadi Abyad suggest fundamental differences in both the timing and acquisition process of magnetization.

In geographic coordinates the majority of site mean directions from Wadi Khafifah have shallow negative inclinations (Table 6.1) and display either N or
NW-directed declinations (Figure 6.6). Correcting for the dip of the local Moho (dip and dip direction = 29/173) results in a shift to moderate positive inclinations (suggesting normal polarity magnetization) with N or NW-directed remanences (Figure 6.6). In both geographic and tilt corrected coordinates the layered gabbros (sites KF04-06 and KF08, plus dyke site KF07) give northerly-directed remanences whilst the foliated gabbro sites (plus site KF03 from the upper layered gabbros) display magnetizations more towards the NW. The overall locality mean directions and Fisher statistics are listed in Table 6.2.

Table 6.2: Locality mean magnetization directions

<table>
<thead>
<tr>
<th>Locality</th>
<th>Unit</th>
<th>N</th>
<th>Geog Dec/Inc</th>
<th>Tilt corr. Dec/Inc</th>
<th>k</th>
<th>α95</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wadi Abyad</td>
<td>Layered &amp; foliated gabbros</td>
<td>17</td>
<td>318/44</td>
<td>340/29</td>
<td>20.3</td>
<td>8.1</td>
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<td>Wadi Abyad</td>
<td>Varitextured gabbros &amp; DRZ</td>
<td>5</td>
<td>061/54</td>
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<td>8.9</td>
</tr>
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<td>Layered gabbros</td>
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<td>318/20</td>
<td>319/21</td>
<td>138.7</td>
<td>5.7</td>
</tr>
<tr>
<td>Wadi Nassif</td>
<td>MTZ and layered gabbros</td>
<td>4</td>
<td>336/-23</td>
<td>338/07</td>
<td>28.7</td>
<td>17.4</td>
</tr>
</tbody>
</table>

N = number of sites; Dec = declination; Inc = inclination; Geog = geographic (in situ); Tilt corr = tilt corrected; k = Fisher precision parameter; α95 = 95% cone of confidence

Figure 6.7: Equal area stereographic projections in geographic (left) and tilt corrected (right) coordinates showing the distribution of site mean magnetization directions from Wadi Khaffah. Layered gabbros = blue symbols; foliated gabbros = red; dykes = green. The large light purple symbol on each plot denotes the locality mean magnetization direction. Solid/open symbols = lower/upper hemisphere projections; ellipses = projection of α95 cones of confidence around site means.
Individual site mean magnetization directions from the Somrah layered gabbro (SR01-05) and dyke (SR06+07) sites are very well clustered and show shallowly inclined, NW-directed remanences in both geographic and tilt corrected coordinates. The tilt correction at this locality is defined by the orientation of the modal layering (dip and dip direction = 06/098) (Table 6.1 and Figure 6.7). Given the similar site mean directions recorded at Somrah in both the gabbro and the dykes, it can be surmised that both units acquired their magnetization at the same time. The locality mean directions and Fisher statistics for Somrah are listed in Table 6.2.

Site mean magnetizations from Wadi Nassif have negative inclinations in geographic coordinates, and all except site WN01 display a NW-directed remanence; WN01 shows a N-directed magnetization (Table 6.1 and Figure 6.8). Correcting for the dip of the Moho (dip and dip direction = 32/180) results in site mean directions with shallow inclinations (~6°) that suggest a normal polarity magnetization, while maintaining their N (WN01) and NW (WN02-04)

Figure 6.8: Equal area stereographic projections in geographic (left) and tilt corrected (right) coordinates showing the distribution of site mean magnetization directions from Somrah. Layered gabbros = blue symbols; dykes = green. The large light purple symbol on each plot denotes the locality mean magnetization direction. Solid/open symbols = lower/upper hemisphere projections; ellipses = projection of α95 cones of confidence around site means.
declinations (Table 6.1 and Figure 6.8). The more northerly-directed magnetization recorded by site WN01 could be a result of its location within the MTZ and not the "proper" layered gabbro sequence sampled at sites WN02-WN04. The locality mean magnetization direction for Wadi Nassif MTZ and lower layered gabbro sites combined is listed in Table 6.2.

As described previously (Chapter 3), the Semail ophiolite is divided into about a dozen tectonic blocks that are loosely grouped together as the northern, central and southern massifs. Sampling efforts for this study were primarily focused within the southern massifs (Rustaq, Samail and Samad-Wadi Tayin) where the best exposures of lower oceanic crustal rocks are found. Previous palaeomagnetic studies within the southern massifs by Luyendyk & Day (1982), Luyendyk et al. (1982), Shelton (1984), Thomas (1988), Feinberg et al. (1999), Perrin et al. (2000) and Weiler (2000) (see Chapter 3, section 3.6 for full details) consistently reported N/NW-directed magnetizations with moderate to shallow
positive inclinations (Figure 6.9). The majority of data obtained within this study are therefore wholly comparable with previous results and suggest that a predominantly N/NW-directed remanence is preserved within the gabbroic lower oceanic crust of the southern massifs. The NE-directed magnetizations observed in the pseudostratigraphically highest outcrops of the Wadi Abyad crustal section, however, have not previously been reported and are significantly different to the predominant N/NW-directed magnetizations observed within the southern massifs. The importance and origin of the NE-directed magnetization and how it relates to the more commonly detected N/NW-directed magnetization will be discussed in detail in section 6.7.

Figure 6.10: Summary of published palaeomagnetic data (black outlined clock diagrams) from the southern massifs of the Semail ophiolite, combined with results from this study (red outlined clock diagrams) (modified from Weiler, 2000; base map modified from e.g. Lippard et al. 1986; Nicolas & Boudier, 2011). Arrows = mean tilt corrected remanence directions. Sources: L = Luyendyk & Day (1982); S = Shelton (1984); T = Thomas (1988); F = Feinberg et al. (1999); W = Weiler (2000); M = Meyer (this study). Blue stars: WA = Wadi Abyad; SR = Somrah; WN = Wadi Nassif; KF = Wadi Khaffifah). Note the consistency between the N/NW-directed magnetizations reported in this study to the published data. Note also the anomalous NE-directed magnetization recorded in the highest-levels of the Wadi Abyad crustal section.
6.5 Secondary magnetization components

6.5.1 Secondary components from the layered and foliated gabbros

As discussed above, a number of specimens displayed more than one component when demagnetized. Removal of secondary components took place in AFs of approximately 15 mT and between 350-400°C when thermal demagnetization was used. The secondary magnetization components from samples collected within the varitextured gabbros, dykes and dyke-rooting zone site (i.e. sites WA13-17 with NE-directed site mean ChRM) from the top of the Wadi Abyad plutonic sequence are considered separately in section 6.5.2 and are, therefore, not included in the following evaluation. For the layered and foliated gabbros (and discrete dykes within these units), site-level mean secondary magnetization directions (in geographic and tilt corrected coordinates) for sites where sufficient data were obtained are listed in Table 6.3.

Of these sites, three had only one specimen showing a secondary component (e.g. WA39, KF02 and KFS02) and a further five sites (e.g. WA07, KF07, KF01 and both foliated gabbro transects) had unacceptably high \( \alpha_{95} \) and low k values (Table 6.3) and have, therefore, been omitted from further consideration. The remaining 10 statistically viable site mean secondary magnetizations (three from Wadi Abyad, four from Khafifah, two from Somrah and one from Wadi Nassif) are presented on the stereonets of Figure 6.10 in both geographic and tilt corrected coordinates.

In geographic coordinates, these secondary magnetizations are well-grouped, with an overall mean direction of Dec/Inc = 351/34, N = 10, k = 18.4, \( \alpha_{95} = 11.6^\circ \). The application of locality-specific tectonic corrections results in a significant increase in dispersion (Dec/Inc = 354/39, N = 10, k = 6.7, \( \alpha_{95} = 20.2^\circ \)), and the data fail a fold test (Tauxe, 2010). Acquisition of this secondary magnetization must have, therefore, taken place after tectonic disturbance of
the Moho (in Wadi Abyad, Khafifah and Wadi Nassif) and gabbro layering (at Somrah). Their mean direction in geographic coordinates is statistically indistinguishable from the present day axial geocentric field direction of Dec/Inc = 360/41 for sites in the southern part of the Al Hajar Mountains (see red star on Figure 6.10). Combined with their low coercivities and relatively low unblocking temperatures, this indicates that they represent recent viscous remanent magnetizations (VRMs).

Table 6.3: Secondary magnetization component data from the Semail ophiolite, excluding results from the top of the Wadi Abyad crustal section. Secondary (i.e. low coercivity/unblocking temperature) component site mean magnetization directions from sites with N/NW-directed ChRMs in both geographic and tilt corrected coordinates. N = number of specimens; Dec. = declination; Inc. = Inclination; k = Fisher precision parameter; α₉⁵ = semi-angle of 95% cone of confidence; LG = layered gabbros; FG = foliated gabbros; D = dykes.

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<th></th>
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<th>α₉⁵</th>
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<td>WA07</td>
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<td>LG</td>
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<td>356 / 45</td>
<td>347 / 77</td>
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</tr>
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</table>
6.5.2 Secondary magnetization components from the varitextured gabbro and dyke-rooting zone horizons in Wadi Abyad

The secondary magnetizations identified in specimens from the upper part of the Wadi Abyad crustal section (Table 6.4 and Figure 6.11) are noticeably different to those seen in sites with N/NW-directed ChRM (compare with Figure 6.10). Their mean direction is Dec/Inc = 021/63, \(N = 3, k = 67.8, \alpha_{95} = 15.1\) in geographic coordinates and Dec/Inc = 026/31 after tilt correction. Site WA16 also displayed a third (younger?) magnetization component that is directed approximately towards the NW with inclinations of 54° and 44° in geographic and corrected coordinates, respectively (Table 6.4 and Figure 6.11, labelled as WA16t).
When plotted along with the ChRM site means from the entire Wadi Abyad crustal section (Figure 6.12), the secondary (sites WA15, WA16s and WA17) and tertiary (site WA16t) magnetization components from sites WA15-17 form an arc (along a great circle path) between the NE-directed ChRM site
means and the N/NW-directed ChRM site means. The angular differences between the ChRM site means and the secondary magnetization component are approximately 21°, 17° and 32° (in both geographic and tilt corrected coordinates) respectively for sites WA15, WA16 and WA17, and ~35° between WA16s and WA16t.

Figure 6.13: Equal area stereographic projections in geographic (left) and tilt corrected (right) coordinates showing the distribution of secondary (green ellipses) and tertiary (blue ellipses) site mean magnetizations from the upper part of the Wadi Abyad section relative to the site mean ChRMs from the whole of Wadi Abyad. ChRMs from the upper Wadi Abyad units have red ellipses. Black symbols = site mean magnetizations from the layered and foliated gabbros. Note how the secondary and tertiary magnetization components of the varitextured gabbros (WA15), intruding dykes (WA16) and the DRZ (WA17) form an arc (along a great circle) between the NE-directed ChRMs and the N/NW-directed remanences in the underling layered and foliated gabbros. Solid/open symbols = lower/upper hemisphere projections; ellipses = projection of $\alpha_{95}$ cones of confidence around site means.

The overprint magnetization components identified in the varitextured gabbros and DRZ sites reveal a progression from a high stability NE-directed remanence to a lower stability N/NW-directed remanence, with the NW-directed overprint (in specimens from site WA16) being similar to ChRM directions recorded in the lower units (i.e. foliated and layered gabbros) of the Wadi Abyad crustal section (as well as in Wadi Khaffifah, Somrah and Wadi Nassif). This trend, from more stable NE-directed to less stable N/NW-directed magnetization
is also clearly highlighted when complete thermal demagnetization data from representative specimens of the varitextured gabbros and DRZ are plotted (Figure 6.13). Layered (and foliated) gabbro specimens predominately display single component magnetizations, whereas the magnetization of the varitextured gabbros and the DRZ specimens migrates from N-directed NRM s to NE/ENE directions at high temperatures as the low temperature overprints are removed during demagnetization.

Figure 6.14: Equal area stereographic projection in tilt corrected coordinates showing thermal demagnetization data from representative individual samples. Note how layered (and foliated) gabbros typically have single component magnetizations, whereas varitextured gabbro and dyke-rooting zone samples migrate from N to NE/ENE directions as low temperature overprints are removed. NRM = natural remanent magnetization.
The correlation between the single component, N/NW-directed, magnetizations observed within the layered and foliated gabbros and the overprint magnetizations seen in many of the overlying varitextured gabbro and DRZ samples is evidence that the layered and foliated gabbros have been completely remagnetized (consistent with the findings of Feinberg et al., 1999). This also suggests that the varitextured gabbros and DRZ rocks from Wadi Abyad retain an older, possibly original, remanence that is only isolated at high demagnetization temperatures.

6.6 Nature and timing of remanence acquisition

The majority of the calculated ChRMs, with the exception of the NE-directed magnetizations from Wadi Abyad, display a single component, N/NW-directed, remanent magnetization with discrete high coercivity/high unblocking temperatures (see section 6.3). Furthermore, rock magnetic experiments (see Chapter 4, section 4.2) have identified fine-grained SD to PSD secondary magnetite/titanomagnetite (formed by the serpentinization of olivine/hydrothermal alteration of clinopyroxene) as the dominant remanence-carrying minerals within these rocks. Consequently, these results, along with the occurrence of a comparable overprint magnetization in the uppermost gabbros and DRZ of Wadi Abyad, are consistent with the N/NW-directed ChRMs, recorded in the lower and middle gabbros, being a chemical remanent magnetizations formed by the alteration of iron-bearing minerals (i.e. olivine and clinopyroxene) to magnetite/titanomagnetite. This N/NW-directed magnetization is, therefore, a secondary magnetization/remagnetization acquired sometime after initial formation of the rock. A palaeomagnetic fold test can be used to determine the timing of remanence acquisition relative to the tectonic disruption of the Moho during obduction and emplacement of the ophiolite onto the
Arabian passive margin. This regional-scale fold test involves comparing site mean magnetization directions before and after tectonic tilt correction to determine if clustering and precision of mean magnetization directions increase or become more dispersed (see Chapter 2, section 2.4).

Tectonic tilt corrections for each locality have been defined by assuming that the local orientation of the Moho represents a palaeohorizontal surface prior to tectonic disruption (Wadi Abyad, Wadi Khafifah and Wadi Nassif). One exception is at Somrah, where a tilt correction based on the orientation of the laterally continuous, planar layering in the gabbros has been used. Figure 6.14 shows all site mean directions before and after application of these tilt corrections. Excluding data from the top of the plutonic sequence in the Wadi Abyad section that have NE-directed remanences (sites WA13-WA17), sites with N/NW-directed magnetizations have a mean magnetization of Dec/Inc =

Figure 6.15: Stereographic projections in geographic (left) and tilt corrected (right) coordinates showing the distribution of site mean magnetizations for all sites (plotted by lithology) from all localities. Layered gabbros = blue symbols; foliated gabbros = red; varitextured gabbros = orange; dyke-rooting zone = dark purple; dykes = green. The large light purple symbol on each plot denotes the combined mean magnetization direction for the southern massifs of the Semail ophiolite. Solid/open symbols = lower/upper hemisphere projections; ellipses = projection of $\alpha_{95}$ cones of confidence around site means.
330/19, $k = 7.35$, $\alpha_{95} = 8.7$ (N = 37) in geographic coordinates and Dec/Inc = 337/26, $k = 18.57$, $\alpha_{95} = 5.3$ in tilt corrected coordinates. The improvement in Fisher statistics after tilt correction suggests that the N/NW-directed magnetizations were acquired prior to tectonic disruption of the sampling localities. A more rigorous test of this hypothesis is provided by applying a statistical fold test to these data. Here the test of Tauxe & Watson (1994) is used, which combines eigen analysis and parameter estimation (bootstrap) techniques and gives confidence limits on the degree of unfolding required to

![Diagram](image)

Figure 6.16: Fold test (Tauxe and Watson, 1994) results for (a) the layered and foliated gabbros from Wadi Abyad, Wadi Khafifah, Somrah and Wadi Nassif and (b) the layered and foliated gabbros from Wadi Abyad, Wadi Khafifah and Somrah (excluding Wadi Nassif).
produce the tightest grouping of data.

Results of this analysis (performed using the PmagPy program foldtest.py\(^1\)) are shown in Figure 6.15 using data from Wadi Abyad, Wadi Khafifah, Somrah and Wadi Nassif and also recalculated excluding data from Wadi Nassif (which are less well defined at the locality level; see Table 6.2). The dashed red lines in Figure 6.15 are representative plots of the maximum eigenvalue \(\tau_1\) as a function of untilting, and the solid green lines are the cumulative distributions of the percentage untilting required to maximise \(\tau_1\) for all the bootstrapped data sets. The dashed vertical lines are 95% confidence bounds on the percentage of untilting that yields the most clustered result (maximum \(\tau_1\)). If these confidence bounds include 0% then a post-tilt magnetization is indicated, and if they include 100% then a pre-tilt magnetization is indicated (Tauxe & Watson, 1994). In both runs of the test, the confidence bounds bracket 100% untilting, demonstrating that the N/NW-directed remagnetization documented in these rocks was acquired prior to tectonic disruption of the Moho. This indicates that this magnetization, although secondary and not related to the formation of the crust at an oceanic spreading centre, was acquired soon after ophiolite formation and before Campanian structural disruption of the Moho and the overlying ophiolitic nappe (Lippard et al., 1986). This break-up related to development of a major structural culmination (Jebel Akhdar) that resulted in uplift and slippage of the ophiolite and Oman melange off of the developing structural high (Cawood et al., 1990).

The overall mean direction of this N/NW, pre-tilting remagnetization is Dec/Inc = 337.1/24.4, N = 37, k = 19.3, \(\alpha_{95} = 5.4\) (tilt corrected). To use this direction in tectonic analysis it is necessary to demonstrate that the

\(^1\) http://earthref.org/PmagPy/cookbook/
remagnetization event took place over a sufficiently long time period to have adequately averaged out palaeosecular variation (PSV) of the geomagnetic field. This can be tested statistically using the methodology of Deenen et al., (2011), who performed a bootstrap analysis of samples with different N values drawn from PSV models in order to provide upper and lower 95% confidence bounds on $A_{95}$ values of virtual geomagnetic pole (VGP) distributions that adequately average PSV. VGPs calculated from the remagnetized sites have an overall $A_{95}$ of 5.1°. This falls between the critical values for $N = 37$ of $A_{95\text{max}} = 8.4°$ and $A_{95\text{min}} = 4.0°$ (Deenen et al., 2011), demonstrating that these data account for expected PSV and may therefore be employed in tectonic analysis (only if the PSV model of Deenen et al., (2011) holds true for the Cretaceous Normal Superchron (see below) can the data account for the expected PSV).

It is not possible to perform a fold test on data from the single site in the dyke-rooting zone that has the most E-directed remanence of the upper plutonic rocks of Wadi Abyad. However, this must either represent an older remagnetization or a primary thermoremanent magnetization (TRM) dating from

![Figure 6.17: Declination, inclination and palaeolatitude calculated from the African apparent polar wander path of Torsvik et al. (2012) for a site at 23.44°N, 57.62°E (Wadi Abyad). Red symbol = palaeolatitude calculated from the tilt corrected site mean direction at site WA17 (dyke-rooting zone in Wadi Abyad).](image-url)
the time of ophiolite formation in the Late Cretaceous. The positive inclination at this site is consistent with magnetization in the northern hemisphere during a normal polarity interval (as noted in other Oman studies, e.g. Luyendyk and Day, 1982; Thomas, 1988). Remanence was most likely acquired during the Cretaceous Normal Superchron (chron C34N: 114-83 Ma) (Cande & Kent, 1992; 1995). The tilt corrected mean inclination at this site (13°; Table 6.1) indicates a mean palaeolatitude of 6.6°N. Comparing this to palaeolatitudes determined for a site in Wadi Abyad from the African apparent polar wander path (Torsvik et al., 2012; Figure 6.16) suggests that these rocks were magnetized at a position some 400-500 km off the Arabian continental margin if remanence was acquired at the time of formation of the ophiolite. This is consistent with estimates for the southwards displacement of the ophiolite during detachment and subsequent emplacement (Lippard et al., 1986; Searle & Cox, 1999).

6.7 Tectonic interpretation: rotation and remagnetization of the Semail ophiolite

6.7.1 Constraints from palaeomagnetic results from this study

The palaeomagnetic results demonstrate the importance and advantages of sampling complete crustal transects in the Semail ophiolite in order to understand the nature and variability of magnetization directions. By systematically sampling the whole lower crustal sequence exposed in Wadi Abyad, a pattern of remagnetized lowermost gabbros and retention of earlier magnetization by the overlying dyke-rooting zone (DRZ) has been revealed. These data can be used to place constraints on the rotation and remagnetization history of the ophiolite, with the important caveat that original magnetizations have only been recovered at one high-level site (mainly because the sampling was primarily intended to allow study of fabric
development in the lower crustal gabbroic sections). Since the remagnetized sites acquired remanences prior to structural disruption of the Moho, it is appropriate to consider all data in tilt corrected coordinates in order to discuss tectonic rotation of the ophiolite. To recap, the relevant tilt corrected data are:

- **DRZ (single site):** Dec/Inc = 065/13, k = 171, $\alpha_{95} = 4.2$, n = 8 specimens
- **Layered and foliated gabbros:** Dec/Inc = 337.1/24.4, k = 19.3, $\alpha_{95} = 5.4$, N = 37 sites

The tilt corrected inclination of 24° from the remagnetized layered and foliated gabbros exceeds that of any Late Cretaceous reference direction derived from the APWP of Torsvik *et al.* (2012) (see Figure 6.16). The positive fold test indicates that these rocks acquired their remanence prior to structural break-up of the ophiolite while the Moho was consistently oriented across the different localities, but not necessarily still horizontal. Hence, the inclination data suggest that some significant tilting of the ophiolite had occurred prior to remagnetization. According to the model of Feinberg *et al.* (1999), remagnetization took place during or shortly after emplacement of the ophiolite onto the continental margin, which was complete by the end of the Campanian at around 74 Ma (Lippard *et al.*, 1986; Searle & Cox, 1999; Shervais, 2001).

The 70 Ma reference direction derived from the Torsvik *et al.* (2012) African APWP (Figure 6.16) has an inclination of 6-16° (taking into account statistical uncertainties), implying a minimum Moho tilt of 8° at the time of remagnetization.

The comparison of the mean declination of the remagnetized sites (337°±5°) to the 70 Ma reference direction (Figure 6.16) suggests that the southern massifs of the ophiolite have experienced 20-35° of anticlockwise rotation relative to stable Africa since remagnetization (taking statistical uncertainties into account). The mean declination of 065°±4° of the site in the DRZ implies a clockwise rotation relative to stable Africa since formation of the
ophiolite (assuming this direction represents a primary remanence) of 59-71°, but taking into account the post-remanetization anticlockwise rotation the data require a 78-107° clockwise total early rotation. To summarise, the palaeomagnetic data suggest the following sequence of events in the evolution of the sampled parts of the ophiolite:

1. Acquisition of a stable magnetization during or shortly after crustal accretion at c. 90-95 Ma. In the samples collected for this study, this magnetization is currently only retained by rocks in the dyke-rooting zone.
2. Clockwise rotation by 78-107° and minor tilting of the ophiolite, retaining the structural integrity of the Moho across the sampled massifs.
3. Remagnetization of the layered and foliated gabbros from below mediated by fluid flow and alteration, supporting the model of Feinberg et al. (1999).

**6.7.2 Comparison with published palaeomagnetic data and interpretations**

Previous palaeomagnetic research conducted on both intrusive and extrusive rocks in the Semail ophiolite (e.g. Thomas et al., 1988) have shown a large declination anomaly between the northern and southern massifs of the ophiolite (see Chapter 3, section 3.6 for full details). Typically, results from the northern massifs give E/SE-directed magnetizations, whereas remanence directions recorded in southern massifs display N/NW-directed magnetizations (e.g. Perrin et al., 2000; Weiler, 2000) (Figure 6.17). In order to account for this apparent disparity (an angular difference of approximately 130°), a number of tectonic rotation models have been developed (e.g. Thomas et al., 1988; Perrin...
et al., 1994; Feinberg, 1999; Perrin et al., 2000; Weiler, 2000). Each model
differs significantly in their interpretation of this declination anomaly. Thomas et
al. (1988) and Perrin et al. (1994, 2000) favour a large differential rotation
between the northern and southern domains. Weiler (2000) suggested that a
large rotation of the northern massifs took place relative to a fairly static,
coherent grouping of the southern massifs as a result of ridge-related tectonism.
Additionally, palaeomagnetic results from lavas in the northern massifs (e.g. Perrin et al., 1994, 2000; Weiler, 2000; Godard et al., 2003) have demonstrated a complex history of differential tectonic rotations between individual massifs within the northern domain, related to the break-up of the lithosphere above the subduction zone. These rotations are demonstrably of intraoceanic (rather than emplacement-related) origin as analyses of samples from different magmatic phases (V1 and V2) within each block show evidence for rotation during accretion. The interpretation of relative rotations between northern and southern blocks, however, is made more complex by a late-stage, emplacement-related, pervasive remagnetization event that appears to have affected only the southern massifs (Feinberg et al., 1994), with the northern massifs retaining E/SE-directed magnetizations that are presumed to represent primary remanences. The timing of the N/NW-directed remagnetization of the southern massifs and how it relates to the E/SE-directed magnetization of the northern massifs, therefore, has fundamental implications for the rotation history of the Semail ophiolite.

The N/NW-directed remagnetized remanences with moderate to low inclinations recorded in this study from Wadi Khafifa, Somrah, Wadi Nassif and the foliated and layered gabbros in Wadi Abyad are consistent with those used by Weiler (2000) to argue for only limited rotation of the southern domain and for relative rotation of the northern massifs to be internal to the ophiolite, rather than an ophiolite-wide event. However, the discovery in this study of ENE-directed remanences in the uppermost levels of the Wadi Abyad section implies that originally ~E-directed remanences in the southern blocks have been obscured by pervasive remagnetization of the lower crust. The large, pre-remagnetization clockwise rotation of 78-107° inferred from the Wadi Abyad
data is within the range of rotations implied by data from the lavas in the northern blocks (Perrin et al., 1994, 2000; Weiler, 2000; Godard et al., 2003). Together these data are consistent with a regional-scale, bulk rotation of the ophiolite of c. 120-130° (but with superimposed localised variability that would require additional research to fully document), prior to the remagnetization of the southern blocks.

Feinberg et al. (1999) attributed the N/NW-directed remagnetization of the southern massifs to an obduction-related serpentinization event triggered by a hydrothermal wave of high-temperature fluid expelled from beneath the ophiolite during final emplacement at about 75 Ma (i.e. after the Cretaceous Normal Superchron of 114-83 Ma, to account for several antipodal magnetizations they identified), remagnetizing the ophiolite from the base up. This is consistent with the constraint on timing of remagnetization provided by the positive fold test observed in this study, that demonstrates remagnetization prior to Campanian disruption of the Moho (Lippard et al., 1986; Cawood et al., 1990). However, the new data presented here precisely locate the upper limit of remagnetization associated with the hydrothermal wave, which in Wadi Abyad appears to only have extended up to the top of the foliated gabbros, thus preserving an older, potentially primary magnetization in the overlying varitextured gabbros and DRZ. Hence, these results show that the extent of this major remagnetization event can only be unravelled by systematic crustal sampling from the upper crust down into the underlying gabbros. Without such constraints the possibility of different remanence directions being acquired at different times being interpreted erroneously in terms of relative tectonic rotation is a distinct risk.
To conclude, the palaeomagnetic data from this study, along with previously published work, suggest the following sequence of events occurred during the evolution of the Semail ophiolite (Figure 6.19):

1. Formation of the Semail ophiolite at c. 90-95 Ma along a SW-NE trending palaeo-ridge system (Figure 6.19a), along with the acquisition of the now E/SE-directed magnetizations recorded in the northern massifs and the ENE-directed remanences found in the uppermost levels of the Wadi Abyad section.

2. Initiation of a large, ophiolite-wide clockwise rotation of approximately 120-130°, probably driven by impingement of Arabia with a (previously intraoceanic) subduction zone to the south of the future ophiolite, with rotation accommodated by subduction roll-back (Figure 6.19b).

3. Eventual obduction and emplacement of the ophiolite on to the Arabian margin (Figure 6.19c). During this event, the southern massifs impinged upon the Jebel Akhdar structural high, triggering a wave of high-temperature orogenic fluids resulting in the near wholesale remagnetization of the southern massifs from the base up (in support of the model of Feinberg et al. (1999), see Chapter 3, section 3.6). Further (post-remagnetization) development of the Jebel Akhdar culmination then drove anticlockwise back-rotation of the southern blocks (by 20-35°) (Figure 6.19c).

4. Subsequent structural break-up of the ophiolite nappe-stack related to isostatic extensional collapse (Cawood et al., 1990), resulting in the present-day configuration of the Semail ophiolite.

However, it must be stressed that the sequence of events is based on the limited palaeomagnetic data currently available from the ophiolite, and relies particularly on interpretation of the ENE-directed magnetization observed within
one high-level site in Wadi Abyad. Clearly there is scope for more extensive sampling across the critical dyke rooting zone interval in order to confirm the model described above.

Figure 6.19: Schematic model for rotation of the Semail ophiolite: (a) formation by spreading along a NE-SW oriented axis, and acquisition of original remanences; (b) impingement of the Arabian margin on a previously intraoceanic subduction zone, leading to roll-back and clockwise rotation of the future ophiolite; and (c) eventual emplacement of the ophiolite on the Arabian margin, with remagnetization and anticlockwise rotation of the southern blocks related to the influence of the Jebel Akhdar structural high (culmination).
Chapter 7: Conclusions

7.1 Summary

The principal aim of this study was to apply magnetic fabric and palaeomagnetic analyses to a suite of samples collected from the gabbroic sequence of the Semail ophiolite (Oman) in order to investigate the processes by which the lower oceanic forms and deforms. Little attempt had been made previously to use magnetic fabric analyses to quantify the variation in crystalline fabrics of lower oceanic crustal rocks in Oman, even though current models for accretion of the lower oceanic crust predict the development of very different petrofabrics in various parts of the gabbro section. Systematic sampling presented here documents in great detail for the first time the variation of crystalline fabrics through the entire lower oceanic crust of the Semail ophiolite, providing the following key insights:

- AMS principal axes are controlled by the distribution anisotropy of magnetite (plus more minor contributions from paramagnetic phases in samples containing less magnetite), but despite a dominance of secondary magnetite it has been shown that AMS fabrics are coaxial with the silicate fabrics in these rocks. AMS may therefore provide information on magmatic and deformation fabrics in lower crustal rocks.

- Layered gabbros of the southern massifs of the Semail ophiolite have AMS fabrics that are layer-parallel but also have a regional-scale consistency of the orientation of $K_{\text{max}}$ axes. This consistency across sites separated by up to 100 km indicates large-scale controls on fabric development and may be due to consistent magmatic flow associated with the spreading system or
the influence of plate-scale motions on deformation of crystal mushes emplaced in the lower crust.

- Detailed sampling through an individual layer in the gabbros exposed at the classic Somrah locality reveals variations in AMS parameters through the layer that give insights into its mode of formation. Fabrics are consistent with either magmatic flow during emplacement of a melt layer into a lower crustal sill complex, or traction/drag along the base of such a layer in response to regional-scale stresses (e.g. mantle drag). These fabrics, and those observed throughout the layered gabbro sections, are not compatible with the gabbro glacier model, but are consistent with the alternative multiple-sills model for lower crustal accretion.

- AMS in oceanic gabbros could potentially provide information on the variation of strain through the lower crust, which is predicted to increase exponentially with depth in the gabbro glacier model. However, the mineralogical controls on corrected anisotropy degrees in these rocks preclude the use of AMS as a strain marker, unless sampling targets rocks with very similar mineralogical compositions (e.g. exclusively sampling plagioclase-rich, olivine-poor lithologies to avoid variations in anisotropy due to varying percentages of secondary magnetite).

- Magnetic fabric analysis of lower crustal foliated gabbros along two detailed transects reveals distinct differences in fabric style at the same pseudostratigraphic level. These fabrics are consistent with either focused or anastomosing magmatic flow upwards through this layer, supporting the melt migration model for crustal accretion of MacLeod and Yaouancq (2000).
In addition, palaeomagnetic analyses of the same suite of samples have documented two distinct magnetization directions in the lower crustal rocks of the southern Oman ophiolite for the first time:

- N/NNW-directed remanences in the layered and foliated gabbros that are demonstrably produced by a widespread remagnetization event. These pass a regional-scale fold test, and therefore remagnetization occurred prior to structural disruption of the ophiolite in the Campanian. In agreement with previous models, this remagnetization event is inferred to result from a hydrothermal wave from the base of the ophiolite upwards, resulting in production of new magnetite and alteration that obliterated the primary magnetization in these rocks.

- ENE-directed remanences in the dyke-rooting zone of Wadi Abyad, which are partially overprinted by N/NNW-directed components. These are likely to represent remanences that pre-date the remagnetization event, and hence this level represents the upper stratigraphic limit of the remagnetizing fluids.

- The recovery of ENE-directed magnetizations in the southern massifs of the ophiolite for the first time reconciles previously contradictory results that have been variously interpreted in terms of relative rotation. Overall, the new data suggest large (c. 100°) clockwise rotation of the ophiolite, followed by remagnetization of much of the lower crust and a subsequent anticlockwise rotation of 20-35°.

### 7.2 Suggestions for further work

A number of aspects of this study deserve further attention or could trigger related analyses. The most significant suggestions are:
• Further detailed, high-resolution sampling across individual layers within the layered gabbro sequence in order to detect microfabrics related to magmatic emplacement or shear deformation. This should include detailed analysis of zones of oblique fabrics at the margins of layers, and should combine AMS measurements with SEM electron backscatter diffraction analyses of silicate and oxide fabrics.

• Further sampling within the upper foliated gabbros exposed in additional crustal sections in the ophiolite in order to more fully quantify variations in fabric styles at this critical level. This would provide improved insights into melt migration across this zone.

• AMS analysis of fabrics across the transition from the foliated gabbros into the varitextured gabbros to determine the relationship between petrofabrics in the inferred fossil axial magma chamber and those in the underlying feeder zone.

• SEM analysis to determine the crystallographic distribution of magnetite in relation to the silicate framework in more detail in order to more fully understand the source of the AMS signal in lower crustal rocks of Oman.

• Further palaeomagnetic sampling across the gabbro-dyke transition zone along the southern massifs to detect remanences that pre-date the widespread remagnetization event in order to provide robust constraints on the rotation history in relation to that of the northern blocks.
Appendix A: Magnetic fabric results from discrete dykes

A.1 Introduction to sampling of discrete dykes within the layered and foliated gabbros

Discrete dykes were sampled in Wadi Abyad (WA38 from the foliated gabbros; see Figure 3.27 for location), Wadi Khafifah (KF07 and KF09 from the layered and foliated gabbros, respectively; see Figure 2.28) and Somrah (SR06 and SR07 from the layered gabbros; see Figure 3.29) (Figure A.1).

Several AMS studies have used the orientation of the $K_{\text{max}}$ axes within dykes to infer the direction of magma flow related to dyke emplacement, although to conclusively determine the magma flow direction requires specific targeted sample collection from opposing dykes margins that was not undertaken in this study (Staudigel et al., 1992; Tauxe et al., 1998).

A.2 Results of magnetic fabric analysis

Table A.1 presents AMS data from discrete dykes sampled within the layered and foliated gabbros in this study. These data are plotted, along with corresponding structural data (i.e. dyke margin orientation) and mean tensor (Jelínek, 1978) principal axes (with resultant confidence ellipses), on the stereographic projections presented in Figure A.2.

AMS results from the Khafifah (KF07 and KF09) and Somrah dykes (SR06 and SR07) are generally consistent and show a strong magnetic fabric that correlates fairly well with measured dyke plane. Although the $K_{\text{max}}$ axes at both Somrah dyke sites do slightly fall off the plane of the dyke margin. This is also true for the Wadi Abyad dyke (WA38), here the $K_{\text{max}}$ axes plot well off the plane of the dyke plane and there is also a substantial difference between the
azimuth of $K_{\text{max}}$ and the mineral lineation. In conclusion, the consistency of the $K_{\text{max}}$ axes at a site-level in the majority of the dykes are indicative of a flow fabric formed as a result of alignment of crystals parallel to the flow vector.

Figure A.1: Discrete dykes sampled within the Somrah layered gabbros; (a) Dyke site SR06 cutting gabbro layering at a high angle (photo credit: M. Anderson); (b) Detail of interfingering dyke material intruding into gabbros (photo credit: M. Anderson); (c) Dyke site SR07 cutting gabbro layering at a high angle (photo credit: M. Anderson).
Figure A.2: Equal area stereographic projections in geographic coordinates showing the results of magnetic fabric analysis of discrete dykes sampled within the layered and foliated gabbros. WA38 from Wadi Abad; KF07 and KF09 from Wadi Khafifah; SR06 and SR07 from Somrah.

Table A.1: AMS results from discrete dykes sampled within the layered and foliated gabbros of the Semail ophiolite. N = number of specimens; Dec. = declination; Inc. = inclination; $P_J$ = corrected anisotropy degree; $T$ = shape parameter.
Appendix B: Palaeomagnetic and magnetic fabric results from Tuf

B.1 Introduction to the Tuf area
The Tuf area (within the western part of the Samail massif) (Figure B.1) was the focus of research by Korenaga & Kelemen, (1997), where they investigated gabbroic sills within the MTZ in an effort to assess magma transportation processes within the lower oceanic crust. They concluded that the gabbroic sills within the MTZ most likely formed by \textit{in situ} fractional crystallization along the edges of small, open-system melt lenses (Korenaga & Kelemen, 1997). Additionally, this study found these sills to be compositionally and texturally similar to the overlying layered gabbros, suggesting they share a common petrogenesis, as well as providing further evidence supporting \textit{in situ} crystallization within the lowermost crust (see also Kelemen et al., 1997).
In an attempt, therefore, to evaluate the magnetic fabrics produced by the *in situ* emplacement of material, sampling of a 20 m thick sill near to the village of Tuf was conducted. The sill was similar in appearance to layered gabbros within the lower oceanic crust and displayed multiple cycles of compositional layering (Figure B.2), suggesting continuous replenishment of the sills during crystallization (Korenaga & Kelemen, 1997).

Figure B.1: Geological map of the Tuf area (north of Maqsad) with location of sampling sites (TU01-03). Inset map shows location of Tuf within the Samail block in the southern part of the Semail ophiolite (modified from Korenaga & Keleman, 1997).
Figure B.2: Sampled gabbroic sill near Tuf: (a) View of sill in relation to surrounding mantle rocks (photo credit: M. Maffione); (b) Basal contact of sill, site TU01 (source: M. Maffione); (c) Middle of sill, site TU02 (photo credit: M. Maffione); (d) Top of sill, site TU03. Note the small-scale compositional layering (photo credit: M. Maffione).
B.2 Results of magnetic fabric analysis

Table B.1 presents AMS data from Tuf. Sites TU01 and TU03 sampled the base and top of the gabbroic sill (respectively), while samples collected approximately every 2 m through the sill are combined in TU02. These data are plotted, along with corresponding structural data (i.e. the layering of the gabbros within the sill) and mean tensor (Jelinek, 1978) principal axes (with resultant confidence ellipses), on the stereographic projections presented in Figure B.2.

<table>
<thead>
<tr>
<th>Site</th>
<th>N</th>
<th>Km</th>
<th>$K_{\text{max}}$</th>
<th>$K_{\text{int}}$</th>
<th>$K_{\text{min}}$</th>
<th>$P_J$</th>
<th>$T$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Dec. / Inc.</td>
<td>Confidence angles</td>
<td>Dec. / Inc.</td>
<td>Confidence angles</td>
<td></td>
</tr>
<tr>
<td>Tuf</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>TU01</td>
<td>16</td>
<td>3.41E-04</td>
<td>16 / 12 42.1/32.1</td>
<td>110 / 19 44/20.3</td>
<td>257 / 67 29.4/16.8</td>
<td>1.053</td>
<td>0.03</td>
</tr>
<tr>
<td>TU02</td>
<td>14</td>
<td>3.62E-03</td>
<td>240 / 3 36.8/22</td>
<td>345 / 80 55/34.9</td>
<td>150 / 10 55.3/14.6</td>
<td>1.102</td>
<td>0.201</td>
</tr>
<tr>
<td>TU03</td>
<td>13</td>
<td>3.58E-03</td>
<td>24 / 48 25.8/14</td>
<td>190 / 41 29.1/20.3</td>
<td>289 / 7 27.6/17.1</td>
<td>1.103</td>
<td>-0.165</td>
</tr>
</tbody>
</table>

Table B.1: AMS results from a gabbroic sill within the MTZ sampled near the village of Tuf in the Semail ophiolite. N = number of specimens; Dec. = declination; Inc. = inclination; $P_J$ = corrected anisotropy degree; $T$ = shape parameter.

Figure B.3: Equal area stereographic projections in geographic coordinates showing the results of magnetic fabric analysis of samples collected from a gabbroic sill within the MTZ near the village of Tuf.

AMS results from sites TU01 and TU03 are fairly consistent, while TU02 displays a significant scattering of anisotropy axes leading to a poorly defined magnetic fabric. In generally, however a correlation between the magnetic...
fabric and the measured petrofabric (here, magmatic layering) can be recognised with $K_{\text{max}}$ axes typically plotting on or near to the plane of layering.

**B.3 Palaeomagnetic results**

Demagnetization results from Tuf are presented in Figure B.4. Tuf sites TU01 and TU02 (site TU03 has been disregarded due to very poor results) generally display a ChRM directed towards either the NW (TU01) or the NNW (TU02) with linear, if a little noisy, decay to the origin. Both thermal and AF demagnetization techniques gave comparable results, whereas thermal demagnetization also occasionally revealed a secondary magnetization that is removed by approximately 350°C.

![Figure B.4: Characteristic Z-plots from the thermal demagnetization of specimens from the gabbroic sill sampled near the village of Tuf. LG = layered gabbros.](image)

Figure B.4: Characteristic Z-plots from the thermal demagnetization of specimens from the gabbroic sill sampled near the village of Tuf. LG = layered gabbros.
Site mean magnetization directions calculated for Tuf (TU01 and TU02) are presented in Figure B.5. However, given the limited sampling and the small geographic distance between individual samples at Tuf, both sampling sites will be considered collectively so as to attain a more robust locality mean magnetization direction. In geographic coordinates a locality mean of Dec/Inc = 334/13, k = 14.51, $\alpha_{95} = 8$, n = 24 is calculated. Correcting for the dip of the internal layering within the sill (dip and dip direction = 13/019) results in a slight shallowing of the magnetization direction to Dec/Inc = 335/04. The NW-directed magnetization obtained from Tuf is wholly comparable with results from Wadi Abyad (layered and foliated gabbros), Wadi Khafifah, Somrah and Wadi Nassif, however, given the low number of samples collected these results are not as well constrained.

Figure B.5: Equal area stereographic projections in geographic (left) and tilt corrected (right) coordinates showing the distribution of site mean magnetization directions from Tuf. The large light purple symbol on each plot denotes the locality mean magnetization direction. Solid/open symbols = lower/upper hemisphere projections; ellipses = projection of $\alpha_{95}$ cones of confidence around site means.
Appendix C: Relationship between the upper and lower crust in the Troodos ophiolite

C.1 Introduction

In contrast to the evidence for steady-state magmatism at fast-spreading ridges, seismic data at slow-spreading ridges (e.g. Sinha et al., 1998) indicate presence of only intermittent, small magma chambers, raising interesting questions regarding the relationship between the lower and upper crust. In the Troodos ophiolite (Cyprus), which is believed to have formed at a slow-spreading rate, a non-orthogonal relationship between the orientation of the sheeted dykes and layering in the underlying gabbros is observed (Agar & Klitgord, 1995; Abelson et al., 2001). This may reflect decoupling and relative rotation of the upper and lower crust during periods of amagmatic spreading, or a primary relationship resulting from formation of layering in the gabbro section at an initially non-horizontal orientation. A pilot investigation was, therefore, undertaken to examine this geometrical relationship using palaeomagnetic techniques in order to reconstruct the original geometry of the dykes and the gabbros at the time of magnetic remanence acquisition.

C.2 Relationship between upper and lower crust in the Troodos ophiolite

The general architecture of the Troodos ophiolite is now well-constrained (e.g. Allerton & Vine, 1991; MacLeod et al., 1990). It is widely accepted that the Solea Graben in the central Troodos ophiolite represents a palaeo-spreading ridge, although debate on the precise location of the axis set out by MacLeod et al. (1990) is still on-going (e.g. Varga, 2003; Abelson et al., 2002; Nuriel et al.,
The main finding of MacLeod et al. (1990) was actually the identification that not only was a palaeo-spreading ridge preserved in the Troodos ophiolite but a palaeo ridge-transform intersection was also present. This therefore meant that the Troodos ophiolite preserves both the inside and outside corner of an idealized ridge-transform intersection.

Improvements in the imagining of in situ oceanic crust during the 1990s gave rise to new models on the formation and architecture of the oceanic lithosphere as well as how differences in spreading rate affect the evolution of the crust away from the spreading axis. Agar & Klitgord (1995) describe how images of oceanic crust showed what appeared to be faults and detachments zones within the oceanic crust. These faults were attributed to amagmatic processes that take place at slow-spreading ridges like the Mid Atlantic Ridge when magma supply is low. The Troodos ophiolite has since been used to test models of the oceanic crustal architecture of slow-spreading axes, as it is believed to have formed at a ridge system very similar in terms of dynamics as the present-day Mid Atlantic Ridge (Agar & Klitgord, 1995; Abelson et al., 2001).

Faults and detachments similar to those seen in the acoustic imagining of oceanic crust were found within the Troodos ophiolite, and were studied by Hurst et al. (1994) and Agar & Klitgord (1995) who came to the conclusion that the sheeted dyke-gabbro contact was in fact a detachment that formed and propagated as the crust moved away from the spreading axis. In the Agar & Klitgord (1995) model, as the newly formed crust moved away from the spreading axis, cooling and a lack of magma supply in the sheeted dyke complex caused extensional domino style faulting dipping towards the spreading axis. This generated a detachment zone between the sheeted dykes and gabbros below. As movement away from the axis continued the plutonic
complex cooled allowing faults in the overlying sheeted dyke sequence to propagate into the gabbros. The decoupling of the sheeted dykes from the gabbros then caused the generation of antithetic faults within the gabbros, rotating the whole gabbro sequence in the opposite direction to the sheeted dyke complex (Agar & Klitgord, 1995). In the Hurst et al. (1994) model, rotation and antithetic faulting do not occur in the lower crust and, therefore, remain undeformed.

Granot et al. (2006) and Nuriel et al. (2009) recently reinterpreted the nature of the detachment of the sheeted dyke complex from the underlying plutonics. Nuriel et al. (2009) state that the detachment zone on the outside corner (the western side) of the palaeo-spreading ridge formed an oceanic core complex (OCC). This detachment zone propagated along the sheeted dyke-gabbro contact and caused the domino style faulting as the OCC fault flattened out. This model has its advantages, but Nuriel et al. (2009) however fail to state how the detachment on the inside corner (the eastern side) of the Solea Graben palaeo-spreading ridge studied by Agar & Klitgord, (1995) is formed. Granot et al. (2006) however conducted a transect across the palaeo-spreading centre near the palaeo ridge-transform intersection. Their study found that the sheeted dykes and the gabbros in the outside corner underwent a similar horizontal style and sense of deformation and that the detachment zone between the sheeted dyke complex and the gabbros only generated a little extra rotation in the sheeted dykes compared to the gabbros; they therefore speculated that a deeper detachment fault was the cause of the similar rotation (Granot et al., 2006). Granot et al. (2006) believe that this deeper detachment fault formed at the spreading axis and subsequently deepened as the oceanic crust moved away from the ridge axis and cooled. The detachment fault between the
sheeted dyke complex and the gabbros in the inside corner however plays a more important role in the accommodation of extension than in the outside corner (Granot et al., 2006). The decoupling created greater degrees of rotation in the sheeted dykes than the gabbro, however the antithetic sense of faulting in the gabbros seen by Agar & Klitgord (1995) was not reported by Granot et al. (2006). The sheeted dyke complex and the gabbros from the inside corner do however have the same sense of rotation and deformation around a vertical axis, which increase in amount of rotation with increasing distance from the spreading axis. This however could be an artefact of movement along the transform fault as the Granot et al. (2006) transect was conducted near to it (Granot et al., 2006). Clearly no real consensus in the nature of the detachment zone and the sense of rotation has been reached and it is obvious that multiple factors were interacting within both the sheeted dyke complex and the gabbro sequence. Varying styles and senses of deformation have been seen in the area around the Solea Graben palaeo-spreading axis and complexities yet to be identified and studied remain.

C.3 The Troodos ophiolite

The Troodos ophiolite, Cyprus, exposes a well-preserved sequence of oceanic crust and related upper mantle material (e.g. Granot et al., 2006). The formation of the oceanic crust has been dated to around 93 Ma during the Turonian in the Late Cretaceous with uplift during the Late Tertiary (Agar & Klitgord, 1995; Granot et al., 2006). The Troodos ophiolite was not affected greatly by emplacement related tectonics and metamorphism, consequently a complete oceanic crustal stratigraphy can be seen from the extrusive pillow lava, through the sheeted dykes passing into plutonic rocks and finally crossing the
petrological Moho into serpentinized peridotites (Dilek & Flower, 2003; Granot et al., 2006).

Sampling took place around the town of Chandria to the northwest of Agros in the eastern part of the Troodos ophiolite. The Agros area itself is termed the Agros window by Agar & Kiltgord (1995); they describe it as a vertical section through the sheeted dyke complex down into the gabbro sequence below. In the Chandria area the same vertical section can be studied where a road has been cut in the mountainside (known in this study as the Madari Ridge road section) to the north of Chandria exposing the transition from plutonic rocks up into the sheeted dyke complex. Agar & Kiltgord (1995) noted that in the Agros area dykes are found within the plutonic sequence and that they increase in number up through the section gradually, however, the contact between plutonics and sheeted dykes is very sharp. The same was found to be true in this study around Chandria, where the change from 100% sheeted dykes (sites AG01 – AG02) to predominantly (>95%) gabbro (site AG04) occurred over an 800 m section along the road. Sites AG05 – AG08 also confirmed the Agar & Kiltgord (1995) statement that dykes are found within the plutonic sequence, here dolerite, plagiogranite and gabbroic dykes were sampled along with the host (normally layered) gabbro. Hand samples of an ultramafic sill and dyke were also collected at site AG08. Table C.1 gives a brief description and the location of each site sampled during fieldwork in Cyprus with the prefix AG for Agros, the nearest large town to the sampling sites.
Figure C.1 shows a geological map of the central part of the Troodos ophiolite, including a detailed map of the study area near the town of Agros (see insert). The study area included the sheeted dyke-gabbro boundary and a suite of samples was collected from the sheeted-dyke complex down through the boundary and into the gabbro sequence. A total of six sites were sampled as described above in Table C.1, two sites within the sheeted-dyke complex (AG01 and AG02), two sites within the transitional boundary zone (AG03 and AG04) and two sites within the gabbro (AG05-07 and AG08) chosen for their interesting igneous relationships.

<table>
<thead>
<tr>
<th>Site</th>
<th>Coordinates</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>AG01</td>
<td>34°57’5”N, 32°59’49”E</td>
<td>Sheeted dykes on the Madari Ridge road section.</td>
</tr>
<tr>
<td>AG02</td>
<td>34°56’55”N, 32°59’58”E</td>
<td>Sheeted dykes on the Madari Ridge road section 500 m down from AG01.</td>
</tr>
<tr>
<td>AG03</td>
<td>34°56’31”N, 33°00’13”E</td>
<td>Dyke-gabbro transition zone on the Madari Ridge road section approximately 1 km from AG02.</td>
</tr>
<tr>
<td>AG04</td>
<td>34°56’25”N, 33°00’18”E</td>
<td>Gabbro on the Madari Ridge road section and Polystypos crossroads.</td>
</tr>
<tr>
<td>AG05, AG06 and AG07</td>
<td>34°56’39”N, 32°59’46”E</td>
<td>Dyke cut gabbro road section outside of Chandria.</td>
</tr>
<tr>
<td>AG08</td>
<td>34°56’3”N, 32°59’46”E</td>
<td>Layered gabbro cut by dykes and sills on the Chandria to Agros road.</td>
</tr>
</tbody>
</table>

Table C.1: Description and location details of sampling sites around the town of Agros (AG) in the Troodos ophiolite.
C.4 Results and discussion

Almost all specimens from the sheeted dyke localities (sites AG01 and AG02) show two stable components of magnetization. However, the low temperature component is not geologically significant and is not discussed further. Typical examples of demagnetization behavior are shown in Figure C.2. Results suggest that the characteristic remanence is carried by titanium-poor titatomagnetite or magnetite with maximum unblocking temperatures of 550-
560°C, although some samples show significant unblocking at around 300°C, consistent with some remanence being carried by pyrrhotite.

Figure C.2: Examples of demagnetization data from sheeted dykes at sites AG01 and AG02 (in geographic coordinates).

Figure C.3 shows the results of demagnetization of gabbro samples from AG03 and AG04. After removal of a very low temperature/low coercivity component a single magnetization component is observed. This high temperature/high coercivity component is seen throughout the specimens from both sites and produces a tight clustering of magnetization directions. Thermal demagnetization of specimen AG0308A showed intensity dropping off sharply before the magnetite Curie temperature. AG0403B however was demagnetized using the AF method and so the intensity graph is related to the coercivity of the sample. In this case, specimen AG0403B was demagnetized below 10% of its original intensity at 50 mT. The results of demagnetization produced a tightly clustered group of similar direction for both AG03 and AG04 with a mean direction of 47/299. This is different to the Troodos mean magnetization.
direction and therefore it can be concluded that rotation of these packages of rocks (probably as a group) has occurred after acquisition of remanence.

Data from the Chandria road section (sites AG05 and AG07) is presented in Figure C.4 in geographic coordinates. Both example specimens show a stable magnetization component directed to the west with a moderate inclination. Similar to previous results the magnetization of the gabbro was stable until temperatures >500°C when a sharp decrease in intensity is seen before reaching zero at the magnetite Curie temperature (see AG0502 intensity graph). A more distributed unblocking of remanence is seen in sample AG0701A, with an inflection at ~350°C that may suggest that some remanence is carried by pyrrhotite.

Figure C.4: Examples of demagnetization data from gabbros at sites AG03 and AG04 (in geographic coordinates).
At site AG08 three separate lithologies were sampled, an ultramafic layer that intruded into a layered gabbro host rock and a gabbroic dyke that crosscuts the layering of the layered gabbro (Figure C.5). The intensity decrease pattern seen in these units is similar to trends seen previously in this study with gabbro and ultramafic specimens unblocking sharply just before the magnetite Curie temperature. The gabbroic dyke sampled at site AG08 however has a more distributed unblocking, again a hint of a two-step unblocking distribution (like in previous dyke units) at around 300-350°C and 550°C. As Figure 4.15 shows all three units records a similar ancient magnetization direction. The gabbroic dyke however also records a lower unblocking temperature component that is destroyed between 300 and 400°C. The direction of this secondary component is similar in direction to other younger components seen within sampled dyke units in this study and is not geologically significant. The highest stability component seen at AG08 is coincident in all the lithologies and gives a westward trending magnetization direction.

Figure C.4: Examples of demagnetization data from sites AG05 and AG07 (in geographic coordinates).
Site mean directions of magnetization at all sites are given in Table C.2, together with associated Fisher statistics and formation mean directions calculated from the site means. The data are also presented on equal area stereographic projections in Figure C.6, in both geographic (in situ) and stratigraphic (tilt corrected) coordinate systems.

Figure C.5: Examples of demagnetization data from site AG08 (in geographic coordinates).
Table C.2: Palaeomagnetic data from the Troodos ophiolite. Site mean magnetization directions in both geographic and tilt corrected coordinates. N = number of specimens; Dec. = declination; Inc. = Inclination; k = Fisher precision parameter; α95 = semi-angle of 95% cone of confidence; SD = sheeted dykes; G = gabbros; D = dykes.

<table>
<thead>
<tr>
<th>Site</th>
<th>Lithology</th>
<th>N</th>
<th>Geographic coordinates</th>
<th>k</th>
<th>α95</th>
<th>Tilt corrected</th>
<th>k</th>
<th>α95</th>
</tr>
</thead>
<tbody>
<tr>
<td>AG01</td>
<td>SD</td>
<td>9</td>
<td>245.8</td>
<td>38.9</td>
<td>77.7</td>
<td>1.9</td>
<td>244.1</td>
<td>32.3</td>
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<tr>
<td>AG02</td>
<td>SD</td>
<td>8</td>
<td>250.6</td>
<td>47.9</td>
<td>34.7</td>
<td>0.5</td>
<td>253.9</td>
<td>40.4</td>
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<tr>
<td>AG03</td>
<td>G/D</td>
<td>12</td>
<td>328.1</td>
<td>46.9</td>
<td>92.7</td>
<td>4.5</td>
<td>365.5</td>
<td>46.9</td>
</tr>
<tr>
<td>AG04</td>
<td>G</td>
<td>10</td>
<td>297.7</td>
<td>45.3</td>
<td>216.6</td>
<td>3.3</td>
<td>338.3</td>
<td>70.9</td>
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<tr>
<td>AG05</td>
<td>G</td>
<td>6</td>
<td>277.4</td>
<td>35.7</td>
<td>342.3</td>
<td>3.6</td>
<td>277.1</td>
<td>34.7</td>
</tr>
<tr>
<td>AG06: Late dyke</td>
<td>D</td>
<td>5</td>
<td>295.7</td>
<td>60.7</td>
<td>93.4</td>
<td>8</td>
<td>295.4</td>
<td>64.7</td>
</tr>
<tr>
<td>AG08: Host gabbro</td>
<td>G</td>
<td>17</td>
<td>272.5</td>
<td>66.1</td>
<td>130</td>
<td>3.1</td>
<td>280.6</td>
<td>57.7</td>
</tr>
<tr>
<td>Dykes</td>
<td>AG01 and AG02</td>
<td>17</td>
<td>250.4</td>
<td>43.2</td>
<td>40.51</td>
<td>6.7</td>
<td>248.4</td>
<td>36.2</td>
</tr>
<tr>
<td>Gabbro</td>
<td>AG03, AG04, AG05 and AG08</td>
<td>52</td>
<td>285.4</td>
<td>50.8</td>
<td>16.64</td>
<td>5</td>
<td>288.2</td>
<td>32</td>
</tr>
</tbody>
</table>

Figure C.6: Site mean directions of magnetization for the gabbro sites and sheeted dyke sites sampled in Cyprus in geographic and tilt corrected coordinates for comparison to the Troodos mean direction of magnetization. A. Mean directions of magnetization taken from Table C.2 in geographic coordinates for sheeted dyke and gabbro sites with Troodos mean direction shown. The late dyke sampled at AG08 is shown in black to show is magnetization affinity with the gabbro suite compared to the sheeted dyke complex. B. Mean directions of magnetization in tilt corrected coordinates for the sheeted dyke and gabbro sites, once again the Troodos mean direction is shown. C. Formation mean directions of magnetization for the gabbro suite and sheeted dyke complex with the Troodos mean direction of magnetization shown.
Comparison of palaeomagnetic data with those of Abelson et al. (2002) and Granot et al. (2006) reveal some similar results. While this study sampled both the dykes and the gabbros to determine the relationship between the upper and lower crust, peculiarly Abelson et al. (2002) and Granot et al. (2006) did not directly sample the dykes so comparisons cannot be made. However, a similar direction for the gabbros in geographic coordinates is seen in the data presented by Abelson et al. (2001) while the shallow inclination of the magnetization direction seen when a tectonic tilt correction is applied is also seen in the data of Granot et al. (2006).

The following key points are evident from the distributions of site and formation mean directions:

1) Tilt correction of magnetizations at gabbro sites (rotating observed layering in the gabbros to horizontal around the present day line of strike) results in dispersion of data that are clustered in geographic coordinates (mean: Dec = 285.4°, Inc = 50.8°, α95 = 5°, K = 16.64, n = 5). This suggests observed differences in the orientation of layering between sites either: (a) result from deformation prior to remanence acquisition; or (b) represent primary differences in orientation, i.e. that layering does not always represent a palaeohorizontal surface in gabbroic magma chambers.

2) Tilt correction of magnetizations from sheeted dyke sites (rotating dyke orientations to vertical around the present day line of strike of dyke margins) results in a mean direction of Dec = 248.4°, Inc = 36.2° (n = 2). When compared to the Troodos Magnetization Vector reference direction (Dec = 274°, Inc = 36°; Morris et al., 1998) this suggests a 25° anticlockwise rotation of the sheeted dykes exposed along Madari Ridge, and intrusion of dykes in an initially vertical orientation.
3) The mean directions of sheeted dyke and gabbro sites in geographic coordinates are significantly different, with an angular difference of 26.8°. This may be explained by relative tectonic rotation of these pseudostratigraphic levels after acquisition of remanence, and provides evidence of structural decoupling of these levels during seafloor spreading.

C.5 Conclusions

Decoupling of the sheeted-dyke complex from the gabbro suite in Agros may have occurred; the non-orthogonal relationship seen in the field can be identified from the site mean magnetization directions. The significant difference seen in the mean directions of sheeted dyke and gabbro sites in geographic coordinates may be explained by relative tectonic rotation of these pseudostratigraphic levels after acquisition of remanence, and therefore provides evidence of structural decoupling of these levels during seafloor spreading.
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