The petrology, geochemistry and structure of the plutonic rocks of the Oman ophiolite

Thesis

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The petrology, geochemistry, and structure
of the plutonic rocks of the
Oman Ophiolite

A thesis presented for the degree of
Doctor of Philosophy
by
Paul Browning

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The Open University
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Variable print quality
Frontispiece. Landsat image (scale 1:1 000 000) of the Haylayn and Rustaq Blocks, with sketch map showing the corresponding regional geology.
The Babel fish is small, yellow and leech-like, and probably the oddest thing in the Universe. It feeds on brainwave energy received not from its own carrier but from those around it. It absorbs all unconscious mental frequencies from this brainwave energy to nourish itself with. It then excretes into the mind of its carrier a telepathic matrix formed by combining the conscious thought frequencies with nerve signals picked up from the speech centres of the brain. The practical upshot of all this is that if you stick a Babel fish in your ear you can instantly understand anything said to you in any form of language. The speech patterns you actually hear decode the brainwave matrix which has been fed into your mind by your Babel fish.

Now it is such a bizarrely improbable coincidence that anything so mindbogglingly useful could have evolved purely by chance that some thinkers have chosen to see it as a final and clinching proof of the non-existence of God.

The argument goes something like this: "I refuse to prove that I exist," says God, "for proof denies faith, and without faith I am nothing."

"But," says Man, "the Babel fish is a dead giveaway isn't it? It could not have evolved by chance. It proves you exist, and so therefore, by your own arguments, you don't. QED."

"Oh dear," says God, "I hadn't thought of that," and promptly vanishes in a puff of logic.

The Hitch Hiker's Guide to the Galaxy

(The same thing could be argued about double diffusive convection.)
INTERNAL MEMORANDUM

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Abstract

The Oman Ophiolite represents a fragment of oceanic crust and upper mantle that was emplaced onto the Arabian continental margin during the late Cretaceous. An area of 3000km², comprising the Haylayn and Rustaq Blocks, has been mapped and specific localities were investigated in order to examine the petrological and structural processes that operate during and subsequent to the formation of oceanic lithosphere at a spreading axis.

Four successive plutonic suites have been recognized:
(i) the Mantle Sequence, Cumulate Sequence and High Level Intrusives that form the lower part of the classic Penrose ophiolite stratigraphy, which has been intruded by three Later Plutonic Suites,
(ii) the Late Trondhjemite-Gabbro Complexes,
(iii) the Late Peridotite-Gabbro Complexes, and
(iv) the Late Biotite Granites.

The tectonite harzburgite of the Mantle Sequence represents the residuum formed during 17% partial melting of upward flowing fertile spinel lherzolite mantle at pressures of 20kB or more. The primary magmas generated (tholeiitic picrites, minimum MgO 14%) supplied a sub-oceanic spreading ridge magma chamber in which formed the layered mafic and ultramafic rocks of the Cumulate Sequence and the isotropic gabbros, diorites and trondhjemites of the High Level Intrusives. Structures and petrofabrics developed in the Mantle Sequence preserve lines and planes of mantle flow beneath the oceanic crust. Layering in the Cumulate Sequence formed in situ within a long-lived, dynamic, compositionally stratified, open system magma chamber.

The Later Plutonic Suites record the post-spreading igneous history of the Oman Ophiolite. The Late Trondhjemite-Gabbro Complexes and the Late Peridotite-Gabbro Complexes reflect incipient arc formation and subsequent arc-rifting on and in young oceanic crust. The Late Biotite Granites are considered to be syn-emplacement plutons derived from the underthrust Arabian continental margin.

A back-arc (marginal) basin rather than a major ocean basin is considered to be the likely tectonic setting in which these successive magmatic episodes occurred.
Acknowledgements

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Chapter 1

Introduction

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Chapter 1
Introduction

1.1 Ophiolites

It is instructive to consider briefly the historical development of ideas on the origins of ophiolites, as it provides a useful platform from which to view the main areas of continuing debate. Recent reviews of the subject are provided by Coleman (1977) and Gass (1980a,b).

1.1.1 Definition

The term "ophiolite" (derived from the Greek root ophi, meaning snake or serpent) was first used by Brongniart (1827) to describe serpentinites, their shiny green appearance being reminiscent of reptiles. It was not until the early twentieth century that Steinmann (1927) noted the important assemblage of serpentinites, pillow lavas, and radiolarian cherts (the so-called "Steinmann Trinity") within the Mediterranean mountain chains, and used the word "ophiolite" to describe a rock association rather than a rock type.

For almost fifty years "ophiolite" was used and abused until, under the unifying paradigm of plate tectonics, the participants of the GSA Penrose Conference provided the following definition (Anonymous 1972):

"Ophiolite refers to a distinctive assemblage of mafic to ultramafic rocks. It should not be used as a rock name or as a lithologic unit in mapping. In a completely developed ophiolite the rock types occur in the following sequence, starting from the bottom and working up (see Figure 1.1):

1. Ultramafic complex, consisting of variable proportions of harzburgite, lherzolite, and dunite, usually with a metamorphic tectonite fabric (more or less serpentinised);
2. Gabbroic complex, ordinarily with cumulus textures commonly containing cumulus peridotites and pyroxenites and usually less deformed than the ultramafic complex;
3. Mafic sheeted dyke complex;

Associated rock types may include (i) an overlying sedimentary section typically including ribbon cherts, thin shale interbeds,
and minor limestones; (ii) podiform bodies of chromite generally associated with dunite; and (iii) sodic felsic intrusive and extrusive rocks.

Faulted contacts between mappable units are common. Whole sections may be missing. An ophiolite may be incomplete, dismembered, or metamorphosed ophiolite. Although ophiolite generally is interpreted to be oceanic crust and upper mantle, the use of the term should be independent of its supposed origin."

Figure 1.1. Idealised ophiolite stratigraphy after Penrose Conference definition (Anonymous 1972), showing the likely corresponding oceanic seismic layers.

The final sentence of this consensus definition reflects the three main areas of continuing debate; are ophiolites to be equated with oceanic crust and upper mantle, and, if so, what are the processes involved in their formation and the mechanisms of their emplacement onto continental margins?

It is somewhat artificial to consider the development of ideas in these three fields separately; inevitably the growth of a new concept in one area led to a simultaneous advance in understanding in another. Nonetheless the following sections attempt to outline the important contributions that led first to the identification of ophiolite complexes as fragments of oceanic lithosphere formed at constructive plate margins, and
second to an understanding of the processes behind the formation of such oceanic lithosphere. A review of the models for emplacement of ophiolites onto continental margins is beyond the scope of this study.

1.1.2 Ancient oceanic lithosphere?

Steinmann's (1927) recognition of the frequent association of serpentinites and pillow lavas with radiolarian cherts placed ophiolites firmly within a distal oceanic regime. From then until the mid-sixties however, the ophiolitic assemblage was viewed (particularly by European geologists) as being autochthonous, that is intrusive into, or extrusive onto eugeosynclinal sedimentary sequences (e.g. Dubertret 1952, Brunn 1960, 61, Aubouin 1965).

Following his proposal (Hess 1955) of a peridotite-serpentinite model for the oceanic crust, Hess (1960) suggested that the serpentinites of Puerto Rico represented uplifted oceanic material. Shortly afterwards the concept of sea-floor spreading was first aired by Hess (1962), Dietz (1961), and Vine and Matthews (1963). They proposed that new oceanic crust is created at mid-ocean ridges and spreads laterally, being supplied and driven by convection currents that rise and diverge within the oceanic mantle below.

The Troodos Massif, Cyprus, then became central to the development of ideas on the origin of ophiolites and the generation of new crust at oceanic ridges. Bishopp (1952) first suggested that the Troodos Massif represented a volcanic pile developed in an oceanic region that once lay between Eurasia and Africa. Gass and Masson-Smith (1963) concurred with this view and emphasized the intrusive nature of the remarkable Sheeted Intrusive Complex. In particular they tentatively suggested that

"the exposed rocks of the Troodos Plutonic Complex might in fact represent the sub-Mohorovicic peridotitic material which had been partially fused to provide the volcanic material of the Sheeted Intrusive Complex and the Pillow Lava Series."

Elaborating on the earlier work Gass (1967) notes that

"the structure .... of the Sheeted Intrusive Complex, is the type that would be formed on the crest of a mid ocean ridge where
diverging cells in the mantle would produce crustal tension. So, it is possible that the Troodos Massif represents a volcanic edifice formed on the median ridge of Tethys ...."

clearly the link between ophiolite complexes and mid oceanic ridges had now been made.

Moores and Vine (1971) highlighted the importance of the Sheeted Intrusive Complex:

".... within the Sheeted Complex there are 100% dykes implying therefore, 100km of extension in 100km of exposure. The only possible mechanism which has been proposed to account for such extension is sea-floor spreading."

These authors also noted the preponderance of one-sided chilled margins to the dykes of the Sheeted Complex which testified to the constructional nature of the spreading centre environment in which they formed.

Combined with the increasing body of evidence resulting from the expansion in oceanic drilling, dredging and seismic studies, and the growing appreciation of their allochthonous nature, ophiolites were rapidly integrated with the "new global tectonics" (e.g. Coleman 1971, Dewey and Bird 1971, Bailey et al. 1970). By the mid-seventies ophiolite complexes were widely accepted as fragments of oceanic lithosphere that formed at constructive plate margins and were subsequently emplaced onto continental margins.

1.1.3 Processes of formation?

For the first half of the twentieth century the dichotomy in thinking between European and American geologists stifled the unravelling of the petrogenesis of ophiolites. On one hand geologists of the European school (Dubertret 1952, Brunn 1961, and Aubouin 1965) followed the original hypothesis of Steinmann (1927) in regarding ophiolites as colossal eruptions from fractures within developing geosynclines, which differentiated after emplacement to form the apparent stratigraphic sequence from peridotite upwards into gabbros and finally basalts. On the other hand American geologists sought to account for "alpine type peridotites" (Benson 1926), considered to be plutonic in nature and
intrusive into folded geosynclinal sediments. These were explained in terms
of an olivine crystal mush (Bowen 1927), or a wet peridotitic magma (Hess
1938). The problem lay in that the occurrences and associations of the
peridotites in America and Europe were quite different, which led to
European geologists emphasising the close association of peridotite, gabbro
and pillow lava, whereas their American counterparts tended to consider the
peridotites as separate from associated mafic rocks (Coleman 1977).

The publication by Wyllie (1967) of a collection of papers by 33
authors on ultramafic and related rocks is regarded by Coleman (1977) as
being instrumental in focusing the divergent views of the American and
European geologists on the ophiolite problem. In the Wyllie volume Thayer
(1967) emphasised the cogenetic relationship between the peridotites and
the gabbros and diabase, suggesting they could be derived from a single

Figure 1.2. Consensus oceanic spreading model (from Gass and

P Browning 6 Chapter 1.1
primary magma, at last reconciling the American and European positions. (With hindsight it is ironic to note that the paper by Gass (1967) in the same Wyllie volume was grouped under "concentrically zoned ultramafic complexes" with the Duke Island Complex, Alaska!)

As the Troodos Massif was central to providing the link between ophiolites and oceanic lithosphere, it was also pivotal in the development of understanding of the processes occurring at constructive plate margins. The work of Gass (1968), Greenbaum (1972), Gass and Smewing (1973) and particularly Allen (1975), complemented by the models developed in the light of studies on modern oceanic ridges (Talwani et al. 1971, Aumento et al. 1971, Cann 1974) provided a consensus oceanic spreading centre model (Figure 1.2).

On this model, melts derived from the partial fusion of an aluminous peridotite upper mantle rise to feed a sub-axial crustal magma chamber. Tectonite harzburgite represents the residuum formed after partial melting; the pods of dunite represent fractionates from the rising partial melts. In the sub-axial magma chamber mafic and ultramafic cumulates form on the walls and floor, whilst isotropic gabbros underplate the roof. Episodic injection at the spreading ridge crest results in the generation of a sheeted dyke complex, and extrusion of pillow lavas.

In this way a dynamic, long-lived, open system magma chamber which generates the characteristic ophiolite crustal stratigraphy is envisaged. Magma supply from below, and extrusion above are balanced by the rate of crystallization. The continual lateral spreading of the magma chamber flanks maintains a constant volume to the magma chamber.
Figure 1.3. Sketch map showing the distribution of Tethyan ophiolites, and the location of the Oman Ophiolite (from Searle 1980).
1.2 The Oman Ophiolite

The Oman (or Semail) Ophiolite is the largest and best preserved member of a chain of ophiolite complexes extending from the Western Alps through the northern Appennines, the Dinaric Alps, the Taurus Mountains, the Zagros and into the Himalayas (Figure 1.3). This line is regarded by many authors (e.g. Gansser 1964, Dewey and Bird 1970, Smith 1971, Dewey et al. 1973) as marking the locus of Tethyan oceanic plate subduction. In areal extent (about 30 000 km²) the Oman Ophiolite must rank as one of the largest masses of mafic and ultramafic rocks in the world. For comparison, the relative sizes of the Skaergaard, Stillwater and Bushveld Intrusions and the Troodos Ophiolite are shown in Figure 1.5.

1.2.1 Regional geology

Occupying the south-eastern corner of the Arabian Peninsula, the Sultanate of Oman is dominated by the 700 km long arcuate chain of the Oman Mountains. Rising to heights of over 3000m the mountains are bounded to the north-east by the gravel plain of the Batinah Coast, and to the south-west by the deserts of the "Empty Quarter", the Rub al Khali (Figure 1.4).
Figure 1.4. Geography of the Sultanate of Oman, showing the outcrop area of the Oman Mountains (from Graham 1980b).
Figure 1.5. Simplified geological map of the Oman Mountains (from Rothery 1982). The location of the study area (the Haylayn and Rustaq Blocks) is indicated. The areal extents of the Skaergaard, Stillwater and Bushveld intrusions, and the Troodos Ophiolite, are shown at the same scale for comparison.
Understanding of the stratigraphy and structure of the Oman Mountains is based largely on the classic work of Glennie et al. (1973,74) and later refinements by Gealey (1977), Graham (1980b), Searle (1980), and Woodcock and Robertson (1982). A simplified map of the geology is shown in Figure 1.5, and the structural and stratigraphic relationships are summarized in Figure 1.6. The lithostratigraphic units recognized in the Oman Mountains comprise:

1. Crystalline basement rocks: gneisses, amphibolites and schists intruded by granites (860 Ma).
2. Infra-Cambrian to pre-Mid Permian; succession of quartzites, shales, limestones and dolomites deposited under dominantly continental conditions.
3. Hajar Super-Group; Mid-Permian to Upper Cretaceous shallow marine sediments, largely carbonates, lying unconformably on the pre-Middle Permian basement.
4. Sumeini Group; Mid-Permian to Upper Cretaceous shelf edge and slope carbonates.
5. The Hawasina Complex; Permian to mid-Cretaceous basinal facies, mainly redeposited limestones, turbidites and cherts.
6. The Haybi Complex; a complex assemblage of Hawasina sediments, volcanics, metamorphics and serpentinite.
7. The Semail Ophiolite; Upper Cretaceous oceanic crust and mantle.
8. The Batinah Complex; melange and Hawasina-type sediments.
9. Maastrichtian and Tertiary limestones; shallow marine carbonates, overlying unconformably all older units in Oman.

The Sumeini Group, the Hawasina Complex, the Haybi Complex and the Semail Ophiolite were all shown to be allochthonous. Successively higher thrust sheets are progressively more distal in facies, demonstrating the "telescoping" of the continental shelf edge, slope and rise deposits, and ultimately the ocean floor, that occurred during thrusting. Emplacement was directed south-westward onto the Arabian continental margin during closure of the Hawasina (Tethyan) ocean in the Late Cretaceous.
1.2.2 Work up to 1974 on the Oman Ophiolite

The first systematic exploration of the Oman Mountains was that attempted by Pilgrim (1908, 24) who referred to the ophiolitic rocks as firstly the "Basic Igneous Series of Oman" and then later as the "Semail Intrusive Series". In a pioneering work that expressed ideas that bear a remarkable resemblance to views not widely accepted until over fifty years later, Lees (1928) recognized the allochthonous nature of his "Hawasina Series" and the renamed "Semail Igneous Series". With respect to the latter, Lees commented on its both intrusive and extrusive nature and noted that of the rock types present "the commonest is probably serpentinite, then gabbro and diorite." Both Pilgrim and Lees "were greatly handicapped by the hostile attitude of the tribes and by the unhealthy climate."

autochthonous viewpoint attributed the ophiolitic rocks to "submarine extrusion of serpentinite 'nappes' on an immense scale and probably in a viscous state" (Morton 1959).

Reinhardt (1969 and in Glennie et al. 1973,74) recognized the following well-ordered magmatic stratigraphy (from bottom to top) in the "Semail Ophiolite Nappe" (Figure 1.7):

1. Ultra-basic to basic rocks, mainly peridotites (P).
2. Complex transition zone between P and G (PG).
3. Coarsely granular basic rocks, mainly gabbros (G).
4. Complex zone of hypabyssal gabbroid rocks. Coarse, medium and fine grained basic rocks with prevalent ophitic textures (HG).
5. Dyke swarms of hyabbyssal basic rocks, mainly diabase (D).
6. Extrusive rocks, mainly spilites and basalts associated with minor amounts of pelagic sediments (E).

Figure 1.7. Ophiolite stratigraphy recognized by Reinhardt (1969) and in Glennie et al. (1973,74).
Observations central to the model proposed by Reinhardt for the genesis of the ophiolite are paraphrased as follows:

(i) The entire unit of the Semail Ophiolite forms a co-genetic group of rocks structurally, chemically and mineralogically.

(ii) Primary structures (i.e. cross-cutting and host-xenolith relationships) suggest that the various rock members are not contemporaneous. In particular a hiatus can be inferred between the main part of the peridotites (P) and the younger gabbros (G).

(iii) The Semail Ophiolite is thought to have been generated in a system having a wide temperature and pressure range. Progressive differentiation was coupled with decreasing temperature and pressure.

(iv) The peridotites (P) are better considered a metamorphic than an igneous complex; they record equilibrium conditions of about 7kB and over 1000°C.

(v) A planar symmetry is typical of the layering developed in the peridotites and gabbros (PG and G). A "fan structure" is often developed in the area between inclined gabbro layers and the subvertical diabase dyke swarms (see Figure 1.7).

Reinhardt identified a spreading oceanic ridge as the site of formation of the ophiolite (Figure 1.8). The high temperatures and pressures of equilibration, the observed hiatus between the peridotites and gabbros, and the chemically "burned-out" nature of the main mass of peridotites (P) prompted him to suggest a refractory residue model for these rocks. Partial melting at depth provided the magmas that subsequently crystallized to form the peridotites and gabbros (PG and G), the hypabyssal gabbroid rocks (HG), and the diabase and extrusives (D and E).

The lack of unilateral grading structures from the layered peridotites and gabbros (PG and G) led Reinhardt to suggest that they formed in a vertical feeder zone, and were later rotated to the horizontal by a "conveyor belt" mechanism. In addition:

"the fan structure could be original and reflect the tilting of the plutonic strata, while the diabase dykes were generated in a
Figure 1.8. Model developed by Reinhardt (in Glennie et al. 1974) for ophiolite genesis at spreading ocean rise. Note the vertical feeder zone in which the layered peridotites and gabbros (PG and G) are formed, before being rotated to the horizontal by a "conveyor belt" mechanism.

vertical sense... The shallowest part of the feeder system was subject to tensile stress as a result of the drift-apart movement of the underlying uplifted rock masses. The volcanic crust was therefore intermittently pulled apart and the cracks
so formed were filled with upstreaming basaltic magma. The resulting rock units are the vertically arranged diabase dyke swarms of formation (D). They formed the feeder for the extrusive formation (E)."

Reinhardt (in Glennie et al. 1974) noted structures that cut across the axial trend of the ophiolite, and suggested an analogy with transform faults. Regarding the nature of the magma type forming the diabase swarm and the pillowed extrusives, Reinhardt (1969) follows Amstutz (1968) in attributing their spilitic character to "advanced differentiation rather than the result of subaquatic alteration."

Allemann and Peters (1972) concurred with Reinhardt (1969) in assigning pressures and temperatures of equilibration of about 5kbar and 1200°C respectively to the peridotites. They also supported the concept of a basaltic magma being derived by partial melting of an aluminous lherzolite mantle, and leaving a refractory harzburgite residue. Mica-bearing quartz diorites and granite dykes were reported cutting the ophiolite suite, and were considered co-genetic. Excepting the granitic rocks, Allemann and Peters (1972) comment on the limited degree of fractionation compared to Skaergaard and Hawaiian trends, concluding that:

"the ophiolite series did not develop from a single magma reservoir, but from a basaltic magma generated more or less continuously from the mantle."

1.2.3 The Open University and U.S.G.S. projects

Two studies were initiated in the mid 1970's specifically to investigate further the origin and nature of the Oman Ophiolite. The Open University Oman Ophiolite Project led by Prof. I.G. Gass, embarked on a 1:100,000 mapping programme in the north and central Oman mountains (Smewing 1979, Lippard 1980, Lippard and Rothery 1981, Browning and Lippard 1982), whilst a number of American workers co-ordinated by Dr R.G. Coleman of the United States Geological Survey, commenced mapping a 30km wide strip from Muscat to Ibra in the south-east Oman mountains (Hopson et al. 1981, Bailey 1981). The generalized stratigraphy of Oman Ophiolite was found to correspond in all respects to the idealized Penrose stratigraphy (Anonymous
The nomenclature of the ophiolite stratigraphy adopted by the Open University workers is shown in Figure 1.9.


That the Oman Ophiolite formed at an oceanic spreading axis, and that the broad features of the consensus ophiolite petrogenetic model apply, are not now in dispute. The main areas of investigation have focused on how petrogenetic and structural processes operate at a spreading axis, and what is an appropriate plate tectonic setting for such a spreading axis (i.e. major ocean basin or smaller marginal sea).

The occurrence of copper deposits within the pillow lavas of the Oman ophiolite prompted Bailey and Coleman (1975) to draw an analogy with the massive sulphides that are present in the Troodos Massif, Cyprus. Smewing et al. (1977), describing a mineralized fault zone which they attributed to sub-sea floor hydrothermal processes, "imported" the stratigraphic nomenclature developed for the Troodos complex (Gass and Smewing 1973, Allen 1975 and see Figure 1.9). Thus the lavas of the Oman Ophiolite were divided into the earlier "Axis Sequence" and the later "Upper Pillow Lavas" on the basis of morphology, metamorphic facies and geochemistry. The trondhjemites occurring in the High Level Intrusives were investigated by Aldiss (1978).

Further subdivisions of the lava stratigraphy were made by Alabaster et al. (1980), from bottom to top:

(i) The Geotimes Unit (equivalent to the Axis Sequence of Smewing et al. 1977) dominantly basaltic pillow lavas directly overlying the Sheeted Dyke Complex.

(ii) The Lasail Unit comprising basalts through andesites to rhyolites, associated with inclined intrusive felsitic sheets centred on high-
level trondhjemitic plugs.

(iii) The Alley Unit a second basalt-rhyolite sequence, overlying either the Lasail Unit or the Geotimes Unit, and associated with graben structures.

On the basis of trace element geochemistry Pearce et al. (1981) and Alabaster et al. (in press) extended this work and suggested that:

(i) The Geotimes Unit was erupted during back-arc spreading in a marginal basin.

(ii) The Lasail Unit formed as discrete volcanic seamounts founded on
the previously formed back-arc oceanic crust, during a period of island arc magmatism.

(iii) The Alley Unit volcanism occurred during a later rifting phase which caused graben development.

(iv) The additional Salahi Unit (massive, columnar basalt flows of restricted development) is due to magmatism associated with continent-arc collision.

Likely intrusive equivalents of these volcanic suites have been identified respectively as (Alabaster et al. 1980, Smewing 1980b, Browning and Smewing 1981, Pearce et al. 1981):

(i) the layered gabbros and peridotites of the Cumulate Sequence,
(ii) widely distributed 5km sized gabbro, diorite and trondhjemite intrusions that cut all the crustal levels of the ophiolite,
(iii) strongly fault controlled 1km sized predominantly wehrlitic intrusions, and
(iv) potassic granites in dykes of up to 100m wide.

Variations in the regional orientation of the Sheeted Dyke Complex were interpreted by Smewing (1980a) as being due to generation in a "leaky" transform zone. Pearce et al. (1981) argue that a significant proportion of the sheeted dykes were injected during "arc" magmatism rather than the "axis" (i.e. back-arc spreading) episode. A spreading ridge lying to the south west of the outcrop area was indicated by a study by Pallister (1981) of the chilling directions in the Sheeted Dyke Complex near Ibra; however this was in conflict with other indications of ophiolite symmetry, such as the inclination of cumulate layering relative to the Petrological Moho (see Figure 1.9). Dykes considered to have been originally sub-vertical (i.e. related to the overlying Sheeted Dyke Complex) were found by Rothery (1982) in the Maydan syncline to cut the layering in gabbros at small angles; the steep attitude of the gabbro layering was considered to be primary.

The layered mafic and ultramafic rocks of the Cumulate Sequence have been investigated by Smewing (1980b, 81), Hopson and Pallister (1980), Pallister and Hopson (1980, 81) and Browning and Smewing (1981).
substantial, dynamic, long-lived, open system magma chamber is favoured, in which mantle-derived melts fractionate to form the Cumulate Sequence, and feed the Sheeted Dyke Complex and the pillow lavas above. The type and composition of the phase assemblages developed at the base of the Cumulate Sequence indicates that (Smewing 1980b, 81, and Browning and Smewing 1981):

(i) the degree of fractionation of the melts entering the base of the magma chamber was variable,
(ii) the melts were sufficiently diverse in composition to produce contrasting crystallization orders, and
(iii) the bimodal character of some cyclic units reflects the temporary ponding of primitive magma inputs at the base of the chamber before mixing with the bulk of the overlying, more evolved magma body.

Neary and Brown (1979) and Brown (1980) consider the massive chromitites occurring within dunitic pods in the Mantle Sequence to represent the first formed fractionates from partial melts rising to the overlying crustal magma chamber. Increases in Fe and Al with respect to Cr in the chromitites with decreasing depth within the Mantle Sequence reflect a fractionation trend.

From their study of the Mantle Sequence in the south-east Oman Mountains, Boudier and Coleman (1981) suggest that the dykes and pods within the tectonite harzburgite record two phases of magmatism; dunites represent olivine accumulates from a picritic tholeiite rising to the spreading axis, whereas websteritic and gabbroic dykes were formed from later off-axis olivine-poor tholeiite magmas. In addition they propose that the harzburgite does not represent the contemporaneous co-genetic residue from which the picritic tholeiite was derived, and that the olivine-poor tholeiite was the product of partial melting of fertile mantle at greater depths. Boudier and Coleman (1981) observe clear olivine crystallographic maxima related to the regional foliation and spinel lineation within the tectonite peridotite, which are attributed to mantle flow. Mylonitic textures, a result of ophiolite detachment and emplacement, are observed in narrow bands at the base of the sequence and grade upward into
porphyroblastic textures.

After taking extensive fracturing and grain boundary porosity into account as important in lowering seismic velocities, Christensen and Smewing (1981) conclude that the seismic structure of the Oman Ophiolite is remarkably similar to that of the oceanic crust and mantle.

The 95Ma U-Pb isotopic age determined by Tilton et al. (1981) on zircons from trondhjemites agrees well with the early Cenomanian to early Turonian biostratigraphic ages of sediments that are intercalated within the pillow lavas (Tippit et al. 1981). From a study of lead isotopic composition and of U-Th-Pb systematics, Chen and Pallister (1981) conclude that the magmas of the Oman Ophiolite were derived from an oceanic mantle and not an island arc source.

Combined neodymium, strontium and oxygen investigations (McCulloch et al. 1980, 1981 and Gregory and Taylor 1981) emphasise the oceanic affinity of the Oman Ophiolite, and the effects of seawater interaction. Nd/Sm systematics were shown to be undisturbed with Nd$^{143}/144$ initial ratios being typical of oceanic mantle. (Internal isochrons of 100Ma and 130 Ma were determined). The Rb/Sr and $O^{18}/16$ systems have been disturbed by exchange with seawater. Gregory and Taylor (1981) have attempted to model the hydrothermal circulation system in the light of $O^{18}/16$ studies. Lanphere et al. (1981) interpret variable Sr$^{87}/86$ data as indicating melting of a somewhat heterogeneous source region, but cannot totally exclude seawater exchange as a cause of the variation.

In summary, the structure and composition of the Oman Ophiolite have been found to have strong similarities with modern oceanic crust and upper mantle. Its magmatic history would appear to preclude formation exclusively at a spreading axis; at least three later igneous episodes have been identified. The tectonic setting of the major spreading phase remains a matter of debate; the American group favour a typical major ocean basin setting, whilst the Open University workers would appeal to a marginal basin.
setting, with subsequent island arc development.

1.3 The present study

The purpose of this study was two-fold; first, to map the geology of the Haylayn and Rustaq Blocks (Figure 1.5) in order to complete the regional mapping programme started by the Open University in 1975; second, to investigate the petrology, geochemistry and structure of the plutonic rocks of the ophiolite in order to complement the work presented for Ph.D. by Alabaster (in prep.) on the volcanics and associated mineralization, Brown (in prep.) on the chromitite deposits of the Mantle Sequence, Graham (1980b) on the structure and sedimentology of the Hawasina window, Rothery (1982) on the application of remote sensing techniques to mapping, Searle (1980) on the metamorphic and volcanic rocks underlying the ophiolite, and Shelton (in prep.) on geophysical investigations.

1.3.1 The geology of the Haylayn and Rustaq Blocks

The adjacent Haylayn and Rustaq ophiolite blocks are bounded by the Jebel Akhdar anticline (exposing autochthonous pre-Permian basement and Permian to Cretaceous shelf carbonates) and the Hawasina window (the allochthonous slope facies sediments) on their south western margin, and by the recent outwash gravels of the Batinah Coast to the north-east (Figures 1.5 and 1.10).
Figure 1.10. Geological sketch map of the Haylayn and Rustaq Blocks. For regional setting see Figure 1.5.
Only the lower stratigraphic levels of the Oman Ophiolite (see Figure 1.9) are extensively exposed in the Haylayn and Rustaq Blocks. Tectonite ultramafic rocks (the Mantle Sequence), are successively overlain by layered gabbros and peridotites (the Cumulate Sequence), isotropic gabbros, diorites and trondhjemites (the High Level Intrusives), and a Sheeted Dyke Complex. Pillowed volcanics of the Extrusive Sequence are present only as scattered outcrops on the northeast margin of the Haylayn Block.

The relatively simple structural picture of a north or northeast dipping ophiolite slab is complicated by open folds with NW-SE trending axes which result in outcrop repetition in the west of the Rustaq Block and the southeast of the Haylayn Block. Numerous dip faults displace the ophiolite stratigraphy in both blocks. Within the Haylayn Block (Figure 1.10) strike faults oriented WNW-ESE control the distribution of the Later Plutonic Suites (see below). Numerous dip faults, varying in orientation from NW-SE to NE-SW, displace the ophiolite stratigraphy in both blocks. The precise structural history of these faults is not known; many are considered likely to be pre-emplacement structures that underwent rejuvenation during ophiolite obduction.

The classic "Penrose" ophiolite stratigraphy has been intruded by three later plutonic episodes, for which a time sequence has been established from cross-cutting relationships mapped in the Haylayn Block (Figure 1.11). In time sequence:

1. Late Trondhjemite - Gabbro Complexes, represented by widely distributed, up to 5km sized plutons consisting of isotropic gabbro, diorite and trondhjemite. A ubiquitous feature of these plutons is the abundance of xenoliths of dolerite, isotropic gabbro, and layered gabbro and peridotite.

2. Late Peridotite - Gabbro Complexes, represented by elongate intrusions up to 1km in length, which are strongly fracture-controlled. Generally at least half the pluton is made up of poikilitic wehrfite. The margins are typically gabbroic, and in some examples the ultramafic
rocks are overlain successively by layered gabbro, diorite and trondhjemite.

3. **Late Biotite Granites**, in dykes of up to 100m in width, also fracture controlled.

![Diagram](image)

**Figure 1.11.** Diagram to show the cutting relationships of the later plutonic suites that cut the classic "Penrose" ophiolite stratigraphy of the ophiolite in the Haylayn Block.
1.3.2 Logistics and aims

Two successive winter field seasons were spent on geological investigations on the Rustaq Block (December-March 1978/79) and the Haylayn Block (November-April 1979/80). Working from a base in Sohar (see Figure 1.5), field excursions of up to ten days in duration were made by Land Rover into the Oman Mountains. Access to the less rugged, higher stratigraphic levels of the ophiolite may be made by graded roads. Only one motorable track completely traverses the ophiolite, along Wadi Bani Ghafir in the Haylayn Block. Travel along the northern margin of the Rustaq Block is impeded by sand dunes. Except in Wadi Bani Kharus, the Petrological Moho (the boundary between the Mantle Sequence and the Cumulate Sequence, see Figure 1.9) within the Rustaq Block can be reached only by foot.

Photogeology at scales between 1:20 000 and 1:60 000 was followed by regional mapping on 1:100 000 topographic base maps, areas of particular interest and importance being identified for later, more detailed study. In all an area of 3000km$^2$ was mapped, the results being included on the Wadi Hawasina - Rustaq Oman Geological Map Sheet 4/5 (Browning and Lippard 1982, see enclosure 1).

The nature of the Petrological Moho within the Rustaq Block, and the character of the rock types developed above and below it, was the focus of study in the first field season. The character of the Cumulate Sequence-High Level Intrusives-Sheeted Dyke Complex transition, and elucidation of the field relations of the later plutonic bodies that cut the ophiolite stratigraphy in the Haylayn Block were the objectives of the second field season. The petrography and field relations of the Mantle Sequence, the Cumulate Sequence, and the High Level Intrusives, the nature of the Petrological Moho and the Cumulate Sequence-High Level Intrusives-Sheeted Dyke Complex transition are the subject of Chapter 2. The petrography and field relations of the Later Plutonic-Suites are described in Chapter 3. Reference will be made to other areas of the Oman Ophiolite that were visited, and to samples collected by other Open University workers.

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Later chapters document the results of analytical work by electron microprobe on the chemistry of the phase assemblages developed in the Cumulate Sequence and Mantle Sequence. A comparison of the geochemical variation at three contrasting scales within the Cumulate Sequence is provided by studies of a 1m thick rhythmic layer, of a 600m section of rhythmically layered olivine gabbros, and of the lateral variation along the 50km strike length of the Rustaq Block, which are described in Chapters 4, 5, and 6 respectively. Attention turns to the phase chemistry of the Mantle Sequence in Chapter 7. Pétrofabric data provided by N.I. Christensen is used to augment the petrological studies in Chapter 8. In Chapter 9 an estimate is made of the Oman primary magma composition, and taking into account recent advances in the understanding of basaltic magma chambers, a petrogenetic model is proposed.

The overall aims of this work were to gain a better understanding of:

(i) magma chamber processes in general, particularly those mechanisms responsible for the formation of layered igneous rocks,
(ii) the generation and subsequent evolution of magmas at constructive plate margins,
(iii) the nature of constructional processes at spreading axes, and
(iv) to identify post-spreading axis magmatic episodes and their likely tectonic setting.

1.3.3 Rock classification and description

Classification schemes based on modal mineralogy used in this study are adapted from Streckeisen (1976), Coleman and Peterman (1975), and Cox et al. (1979). Figure 1.12 shows categories suitable for ultramafic, gabbroic and leucocratic rocks occurring in the ophiolite.
Figure 1.12. Classification schemes used in this study, adapted from Streckeisen (1976), Coleman and Peterman (1975), and Cox et al. (1979). Gabbroic rocks containing more than 5% modal amphibole are prefixed by "Hornblende".

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In recent years several authors (Campbell 1978, Mc Birney and Noyes 1979, and Rice 1981) have questioned the origin of layered igneous rocks by a mechanism of crystal settling. However, the nomenclature established by Wager, Brown and Wadsworth (1960) to describe layered igneous rocks was developed with crystal settling mechanisms and subsequent crystallization processes firmly in mind. Although such terms as adcumulate, orthocumulate and mesocumulate have strong genetic connotations, they remain the only ones available for describing textures in layered igneous rocks. As such the now widely accepted "cumulate glossary" (Wager et al. 1960, Jackson 1971, Cox et al. 1979) will be utilised, but, unless specifically stated, no genetic connotation is implied other than "cumulus" inferring early crystallizing and "intercumulus" inferring later crystallizing.

The stratigraphic term "Cumulate Sequence" (Figure 1.9), used by Open University workers to refer to the layered gabbros and peridotites that display cumulate textures, has been retained; rocks that show such textures remain "cumulates" irrespective of their modes of formation.
Chapter 2

The Mantle Sequence, the Cumulate Sequence, and the High Level Intrusives

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  2.1.1 Tectonite peridotite
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Chapter 2

The Mantle Sequence, the Cumulate Sequence, and the High Level Intrusives

The first three sections of the present chapter form a general description of the petrography and field relations of the Mantle Sequence, the Cumulate Sequence, and the High Level Intrusives occurring within the Haylayn and Rustaq Blocks (Figure 2.1 and Enclosure 1). The two final sections describe, by reference to specific localities, firstly the nature of the Petrological Moho (the boundary between the Mantle Sequence and the Cumulate Sequence) and the variation in the overlying Cumulate Sequence stratigraphy, and secondly the character of the transition from the Cumulate Sequence through the High Level Intrusives to the Sheeted Dyke Complex.

2.1 Mantle Sequence

Two components form the Mantle Sequence; first, and by far the most important volumetrically, are pervasively foliated tectonite peridotites; second, a wide variety of gabbroic and ultramafic pods, dykes, pegmatite veins and massive chromitites which are enclosed by, or cut, the tectonite peridotite. The base of the Mantle Sequence is defined by the Semail Thrust, and the top by the Petrological Moho which also marks the base of the overlying Cumulate Sequence. Typically the Mantle Sequence forms a rugged, brown weathering, steeply dissected terrain, with peaks rising to 1000m (Plate 2.1), on which a dendritic drainage pattern is well developed. A traverse through the Mantle Sequence within the Rustaq Block was completed along Wadi Bani Kharus (Figures 2.1 and 2.2) in order to examine the petrology and structure of the tectonite peridotite, and the character and distribution of the enclosed or cross-cutting gabbroic and ultramafic bodies.
Figure 2.1. Geological sketch map of the Haylayn and Rustaq Blocks. The locations of the sections described in this chapter are shown. (See Figure 2.5 for other localities in the Rustaq Block.)
Plate 2.1. View looking west from Tawiyah Moho showing characteristic rugged, steeply dissected Mantle Sequence topography.

2.1.1 Tectonite Peridotite

The bulk (estimated as greater than 99%) of the Mantle Sequence is formed by tectonized harzburgite, dunite and lherzolite. If the degree of serpentinisation is moderate (less than 50%) their colour may be a deep red-brown, but as serpentinisation becomes more intense they blacken. When freshly broken, a green-black granular surface is typical. On weathered surfaces areas of olivine are smooth and brown, orthopyroxene grains stand out and are distinctly yellow-brown, chromite grains have a black splendent quality, and clinopyroxene, if present, is green and vitreous.

A rhythmic compositional layering (or more precisely, a ratio layering defined by variations in the respective proportions of olivine and orthopyroxene) is usually developed in the peridotites, though massive varieties do occur. Laterally impersistent on an outcrop scale, the layering shows bilateral symmetry (i.e. there is no preferential direction in which mineral grading occurs), and ratio contacts are not strongly
developed. Typically this compositional layering is parallel to the foliation defined by the preferred orientation and/or flattening of orthopyroxene grains. In addition a lineation defined by elongation of spinel grains is found to lie within the plane of orthopyroxene foliation.

The bilateral symmetry of the layering and the lack of strong ratio contacts that characterize the compositional layering, combined with its close geometrical relationship to the orthopyroxene foliation and the spinel lineation, suggest that it should be best regarded as a metamorphic segregation (or gneissose) layering. In Chapter 8 a fuller account of the structural and petrofabric data from the Mantle Sequence in Wadi Bani Kharus is presented; in particular it is shown that the mean orientation of the orthopyroxene foliation (and the attitude of the Petrological Moho) lie normal to the plane containing the dykes of the Sheeted Complex. It is therefore argued that the mean orthopyroxene foliation may represent a palaeohorizontal and may be used to estimate a stratigraphic thickness of the Mantle Sequence. For a mean orthopyroxene foliation dipping at 25°N, a thickness of 6.3 km is indicated for Wadi Bani Kharus (Figure 2.2).

As a result of the continual grading of harzburgite into dunite by the decreasing modal proportion of orthopyroxene, or more rarely the harzburgite grading into lherzolite by increasing amounts of clinopyroxene, assessing the exact proportions of the differing lithologies is not straightforward. From field observation it may be concluded that harzburgite predominates, dunite is subordinate, and lherzolite is rare.

In thin section the harzburgites appear mosaic-porphyroclastic in texture (following the terminology of Harte (1976)). Up to 10 mm sized porphyroclasts of orthopyroxene, often strained and kinked, and showing very fine exsolution of clinopyroxene parallel to 100, lie in a finer (up to 5mm sized) mosaic of equant, polygonal olivine grains (Plate 2.2). Two types of chrome spinel grains are present; small (less than 0.05 mm), scattered grains sometimes showing good crystal form, and larger (up to 2mm) grains, typically displaying cusparate boundaries.
Plate 2.2. Photomicrograph of tectonite harzburgite. Large (3mm) orthopyroxene porphyroblast (in extinction) in fine groundmass of recrystallized olivine. Note undulose extinction revealing glide planes in olivine. This sample (1499) is unusual in being free of serpentinization. Field of view 8mm.

In the more strongly deformed specimens, restricted to the top quarter of the section and exemplified by sample 2307, disrupted-mosaic-porphyroclastic textures are present. The shape of the orthopyroxene porphyroclasts becomes markedly tabular, often occurring as aggregates of grains rather than single grains, the olivine mosaic becomes finer grained (less than 2mm), and the spinel grains are disrupted being present as discontinuous stringers of relatively anhedral small grains.

Some of the samples from the lower half of the traverse (e.g. 2315, 2328, 2333 and 2339, see Figure 2.2) show comparatively moderate deformation, the orthopyroxenes being less severely strained and of similar size to the olivine grains. A possibly more lherzolitic nature to the Mantle Sequence is indicated mid-section at the level of samples 2315 and 2317, in which lenticles of 3mm sized clinopyroxene grains occur in association with the cuspatelike variety of spinel grain. An average mode of the harzburgites
of the Mantle Sequence has been given by Brown (in prep.) as 0.800 Opx0.9 Chr0.9 Cpx0.2. The bottom two kilometres of the Wadi Bani Kharus section are made up of heavily serpentised, often mylonitised, dunite and harzburgite interlayered on a 1-10 metre scale. This is equivalent to the "Banded Unit" of Searle (1980), and its formation is attributed to low temperature deformation during detachment and thrusting of the ophiolite slab (Searle 1980, Boudier and Coleman 1981).

2.1.2 Gabbroic and ultramafic pods and dykes

A variety of mafic and ultramafic dykes and pods are seen to cut or be enclosed by the tectonite harzburgite (Figure 2.2). These are divided into two categories on the basis of their field relations:

(i) Pods of dunite and olivine gabbro that, when not highly deformed, form irregular, sometimes anastomosing bodies 1-10m in size with diffuse margins that appear to cut pre-existing Mantle Sequence structures, but, when strongly deformed lie within the Mantle Sequence foliation, and show well defined margins.

(ii) Dykes of gabbro, olivine gabbro, gabbronorite and pyroxenite, which are undeformed, sharp sided, discordant, often tabular and between 1cm and 1.5m in width, that cut the compositional layering and foliation of the Mantle Sequence.

In the lower third of the Wadi Bani Kharus section (Figure 2.2) irregular pods of dunite of the order of ten metres in size are seen to cut the tectonite harzburgite (Plate 2.3). Their margins are diffuse, appearing to "interfingeer" along the northward dipping plane of compositional layering and foliation in the harzburgite. Metre sized bodies of harzburgite float in the dunite pod. In the top of the Wadi Bani Kharus traverse (Figure 2.2) irregular bodies of pegmatitic olivine gabbro, up to two metres in size, again with diffuse margins, are seen to cut the tectonite harzburgite (Plate 2.4).
Figure 2.2. Section through the Mantle Sequence of Wadi Bani Kharus. Distribution of various pod and dyke types is shown.

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Immediately below the Petrological Moho a number of gabbroic bodies of contrasting textural styles are found deformed along the plane of compositional layering and foliation in the tectonite harzburgite (Plate 2.5). The most deformed example of harzburgite, sample 2307, also came from this locality. Three types of gabbroic body were identified:

(i) Fine grained (less than 2mm), sharp sided bodies of olivine gabbro of up to 10cm in thickness (Plate 2.6).

(ii) Coarse grained, sharp sided bodies of "flasered" olivine gabbro
Plate 2.4. Detail of diffuse margin of irregular pod of olivine gabbro in tectonite harzburgite from top of Wadi Bani Kharus section.

containing "augen" of clinopyroxene up to 3cm in size (Plate 2.7).

(iii) Diffuse, elongate, centimetre sized blebs of clinopyroxene and plagioclase in a harzburgite host (Plate 2.8), which may be compared with the plagioclase lherzolites of the Troodos massif (Menzies and Allen 1974).

The large clinopyroxene augen are not considered to be relict grains; in thin section they are unstrained and show vermicular "structural unmixing" of clinopyroxene blebs in a clinopyroxene host (see Chapter 4), and therefore would appear to be neoblastic. All three gabbroic types are considered equivalent to the relatively undeformed olivine gabbro pegmatite pods that occur at slightly lower levels in the Mantle Sequence; their constrasting textural styles reflect differing stages in a deformational history. It is suggested that type (i) evolves to type (ii) on extensive recrystallization of the large clinopyroxene augen; continued deformation leads to boudinaging of type (ii) to yield type (iii) (Figure 2.3). In this context it is pertinent to note that metamorphic segregation of
Plate 2.5. Three types of contrasting gabbroic bodies that occur below the Petrological Moho in Wadi Bani Kharus. From top of picture downwards, diffuse-bleb type, flasered sharp-sided type and fine sharp-sided type. See Plates 2.6, 2.7 and 2.8 for closer scale views.

Orthopyroxene need not be the only mechanism of formation of the harzburgite-dunite compositional layering that characterizes the tectonite peridotite of the Mantle Sequence; such layering could also form as a result of the extensive deformation of initially irregular dunite pods (see Figure 2.3), like those described from the lower half of the Wadi Bani Kharus section (as in Plate 2.3). Late aphyric basic dykes, part of the Later Plutonic Suites (Chapter 3), cut all three types of gabbro body and the
Plate 2.6. Fine grained, sharp-sided body of olivine gabbro concordant with foliation in tectonite harzburgite, Wadi Bani Kharus Moho.

Petrological Moho.

Of the sharp sided, discordant, tabular bodies that cut the Mantle Sequence, three categories have been distinguished (Figure 2.2):

(i) **pyroxenitic types** restricted to the lower half of the section,
(ii) **gabbroic types** that occur in the upper half of the traverse, and
(iii) **gabbronoritic types** that occur at all levels.

With respect to the pyroxenites a further point about their distribution may be made; orthopyroxenites occur in the lower part, whilst clinopyroxenites are present in the upper part of the lowermost Mantle Sequence. The orthopyroxenite dykes are seen to cut the relatively undeformed, irregular dunite pods in the lower Mantle Sequence. They are pegmatitic, with grain sizes of up to 3cm common, and in thin section show sutured grain boundaries with associated fine recrystallization (Plate 2.9). Grains may be strained or mechanically broken, and clinopyroxene may be present as exsolved blebs or discrete grains along with chrome spinel. The
Plate 2.7. Coarse, flasered, sharp sided body of olivine gabbro concordant with foliation in tectonite harzburgite, Wadi Bani Kharus Moho.

Clinopyroxenites are texturally similar to the orthopyroxenites, except that grains of orthopyroxene are interstitial; sample 2326 shows evidence of cataclasis during brittle fracture. Some Cu-Ni-Fe sulphides are present as accessory grains.

Gabbronoritic dyke types are seen to cut the clinopyroxenite dykes, and show a variety of phase assemblages and textural types; in particular some composite dyke types are observed. Sample 2298 is a fine grained (less than 1.5mm), granular textured olivine gabbronorite that shows no evidence of chilling against the harzburgite host. Sample 2321 is a composite gabbronorite dyke which has coarse (up to 3mm) margins, with a fine (less than 0.5mm) recrystallized centre (Plate 2.10). Sample 2322 is also composite; a fine (less than 2mm) recrystallized centre of gabbronorite is bordered by pegmatite (greater than 6mm grain size) gabbro. Sample 2325 is a pegmatitic olivine gabbronorite showing a large range of grain size from 0.5 to 7mm. (For all samples locations see Figure 2.2.)
Plate 2.8. Diffuse, elongate blebs of gabbro in tectonite harzburgite immediately below Petrological Moho in Wadi Bani Kharus. The bulk composition of the rock is plagioclase lherzolite.

The gabbro and olivine gabbro dykes restricted to the upper half of the Mantle Sequence are pegmatitic, and also show sutured grain boundaries and concomitant recrystallization of the constituent phases. Chrome spinel and Ni-Fe sulphides are present as accessory phases.

The orientation of dykes within any one category would appear to be fairly consistent (Figure 2.4); the pyroxenites vary in strike between N-S and NW-SE, and dip between north and east; the gabbronorites strike dominantly NE-SW and dip northwest; and the gabbro dykes strike between NW-SE and N-S, and dip between south and west.

On a 10m outcrop scale the Mantle Sequence may appear heterogeneous but over the 700km length of the northern Oman Mountains they are, as noted by Allen (1975) with respect to the same lithologies on the Troodos massif, "impressive in their monotony."
Figure 2.3. Schematic diagram to illustrate formation of contrasting gabbro bodies at Wadi Bani Kharus Moho by progressive deformation of irregular gabbroic pods with diffuse margins that are seen slightly lower in the Mantle Sequence. Also shown are two mechanisms for the production of compositional layering in harzburgite: deformation of irregular dunite pods or metamorphic segregation of orthopyroxene.

2.2 The Cumulate Sequence

The Cumulate Sequence consists of layered gabbroic and ultramafic rocks showing cumulate textures. The stratigraphy of the various mafic and ultramafic rock types comprising the Cumulate Sequence is not uniform throughout the Oman Ophiolite. The only generalization that may be made is that a preponderance of ultramafic types characterizes the lower half of the Cumulate Sequence, whilst mafic types are more abundant in the upper half. An account of the vertical and lateral variation in the assemblages developed within the Cumulate Sequence of the Rustaq Block is given in
section 2.4. The base of the Cumulate Sequence is named the "Petrological Moho", and is defined as the point of upward disappearance of tectonite harzburgite. The upper limit of the Cumulate Sequence is taken as the level at which regular layering disappears from the plutonics; this marks the lower limit of the High Level Intrusives. The Cumulate Sequence forms a less intensely dissected terrain than the Mantle Sequence, but it is nonetheless rugged (Plate 2.11), with mountains of up to 1000m in height. Characteristically it weathers to lighter shades of brown than the Mantle Sequence, depending on the relative proportions of gabbros and ultramafics present.

In Chapter 8 the attitudes of compositional layering within the Cumulate Sequence are compared with the orientation of structures in the Mantle Sequence and the Sheeted Dyke Complex; it is concluded that the compositional layering may form at all angles, with moderate inclinations (c.30°) typical of the lower third of the Cumulate Sequence, and steeper and more irregular attitudes occurring in the higher levels. As such
stratigraphic thicknesses measured normal to the plane of compositional layering are not meaningful. Structural thicknesses for non-faulted sections in the Rustaq Block, based on the assumption that the orientation of the Petrological Moho and the base of the High Level Intrusives are similar, indicate thicknesses of between 700m, at Wukabah and 3000m at Abu Dhabu (Figure 2.6).

Differing phase assemblages are developed in the Cumulate Sequence depending on the crystallization order displayed. For the lower levels of the Cumulate Sequence in the Haylayn and Rustaq Blocks the most common crystallization order seen is olivine (ol) (+ chrome-spinel (sp)) → clinopyroxene (cpx) → plagioclase (plag), giving rise to dunite (ol), wehlite (ol + cpx) and olivine gabbro (ol + cpx + plag) assemblages. However, in the eastern part of Rustaq Block the crystallization order ol (+sp) → plag → cpx is displayed, leading to the development of troctolite (ol + plag) assemblages. (From the Wadi Ragmi area in the Fizh Block, Smewing (1981) has reported crystallization orders involving orthopyroxene
such as ol (+sp) -> cpx -> orthopyroxene (opx) -> plag which give websterite (cpx + opx) and gabbronorite (N) and gabbro (G) types appear to be consistent within any one category.

Figure 2.4. Lower hemisphere, equal area projection showing poles to discordant dykes in Wadi Bani Kharus Mantle Sequence traverse. Orientations of pyroxenite (Ox and Cx), gabbronorite (N) and gabbro (G) types appear to be consistent within any one category.

Cumulate textures are ubiquitous, though in the lower third of the Cumulate Sequence the overprinting of structures related to those
developed in the underlying Mantle Sequence (see Chapter 8) and the concomitant recrystallization may subsequently modify the original textural character of the rock types present. For example, an adcumulate may suffer recrystallization to resemble more an orthocumulate (Plate 2.12). Grain sizes vary between 1mm and 20mm, with medium grain sizes being the most common. Igneous lamination, especially of plagioclase grains is common. Meso- and adcumulates predominate, but heteradcumulates commonly occur in wehrlitic assemblages. Crescumulate features have not been observed.

Rhythmic layering is very well and widely developed in the Cumulate Sequence. "Average rocks" (Wager and Brown 1968) showing no mineral grading for thicknesses greater than a few metres are rare, and if present are found towards the top of the Cumulate Sequence. Layers are predominantly mineral graded (Plate 2.13), though isomodal layers do occur (Plate 2.14). Present on a 10cm to 2m scale, the layers are defined by ratio contacts. Size grading and form contacts have not been observed. On an outcrop scale it is not unusual for layers to overstep and truncate...
Plate 2.12. Photomicrographs showing a typical adcumulate (2400) from Wadi Bani Kharus before (above) and after (below) recrystallization following deformation (2396). After deformation the adcumulate resembles more an orthocumulate in texture.

other layers (Plate 2.15), or for them to vary in thickness, often pinching out. With this in mind it can be said that a "layer group" may be
persistent over a distance of the order of several kilometres. Better control on the dimensions of layers requires more detailed mapping.


Plate 2.15. Layer truncation in olivine gabbro.

Small scale structures (possibly analogous to sedimentary structures) are occasionally exhibited in or between layers. Possible channel scour and fill (Plate 2.16), load cast and flame structures (Plate 2.17), and slump folding (Plate 2.18) are illustrated. In one locality a "dropstone" of olivine melagabbro deforms layers below it, and subsequent layers are draped over its top (Plate 2.19). Locally alignment of acicular grains (usually pyroxene) defines a magmatic lineation.

The layering structures occurring on a metre scale described above are superimposed upon phase layering at a much larger 5-100m scale. On the basis of the large scale phase layering cyclic units have been defined within the Cumulate Sequence (which are described in detail in section 2.4). If the prevailing crystallization order is for example ol -> cpx -> plag, complete cyclic units comprising dunite, wehrlite and olivine gabbro phase layers may be distinguished. If locally only alternating phase layers of dunite and wehrlite are present, they are referred to as "beheaded" cyclic units (Jackson 1970).
No chilled margins, border zones, or sudden discordances in layering orientation have been seen within the Cumulate Sequence. Apart from the Later Plutonic Suites (Chapter 3) and their associated intrusives, other
Plate 2.18. Slump folding.

Plate 2.19. "Dropstone" disrupting previously formed layers.

cross-cutting bodies are present; fine aphyric basaltic dykes considered equivalent to those of the overlying Sheeted Dyke Complex and pegmatites as
dykes or apparently layers (that do show transgressive relationships). The pegmatite mineralogy is often that of its host (or its hydrous equivalent), although discordant pyroxenite assemblages similar to those described in the Mantle Sequence do occur. Fibrous actinolite, epidote, quartz and disseminated sulphides may occur along fracture and joint surfaces.

The layered gabbroic and ultramafic rocks of Cumulate Sequence, in contrast to the tectonite harzburgite of the Mantle Sequence, are markedly more diverse in terms of the mineral assemblages, textural types and layering structures they display.

2.3 The High Level Intrusives

The High Level Intrusives comprise a laterally impersistent, heterogeneous assemblage of mafic to felsic plutonic rocks characterized by an absence of regular layered structures. The High Level Intrusives overlie the layered rocks of the Cumulate Sequence, the contact being transitional. The top of the High Level Intrusives may be gradational, and is defined as the level at which the abundance of dykes related to the overlying Sheeted Dyke Complex becomes greater than 30%. Highly variable in thickness, the High Level Intrusives may range from being absent to up to 700m. Except where white weathering felsic rock types are abundant, the High Level Intrusives form brown weathering hills similar in colour to those of the Cumulate Sequence, but which are of a much lower, subdued topography with heights rarely exceeding 300m.

Massive, varitextured, and xenolithic gabbros, diorites, quartz diorites and trondhjemites, often heterogeneous on a local scale, characterize the High Level Intrusives. Grain size is highly variable especially where pegmatitic facies are present, but the rocks are typically medium grained (3mm). More mafic varieties are usually sub-ophitic and more felsic types subhedral granular.
The original igneous mineralogies are often replaced as the High Level Intrusives show widespread evidence of greenschist facies hydrothermal metamorphism. The more mafic rock types were originally gabbro with or without olivine and/or primary amphibole and/or magnetite. Quartz bearing assemblages, often showing myrmekitic intergrowths, are diorites, quartz diorites, and trondhjemites. Secondary hydrothermal minerals such as actinolite (replacing clinopyroxene and brown hornblende), epidote (replacing plagioclase) and chlorite are widespread.

Aphyric dolerite dykes related to the overlying Sheeted Dyke Complex are common. Small bodies of trondhjemite often pervade the more mafic types of the High Level Intrusives, as do veins and pods of pegmatite (plagioclase, amphibole, quartz and rarely garnet assemblages). Xenoliths of dolerite in varying states of resorption are widespread, being derived from the overlying Sheeted Dyke Complex.

2.4 Cumulate Sequence Stratigraphy and the Petrological Moho

The Petrological Moho, marking the boundary between the Mantle Sequence and the Cumulate Sequence, was defined in section 2.2 as the point of upward disappearance of tectonite harzburgite. In terms of the consensus ophiolite petrogenetic model outlined in Chapter 1, the Petrological Moho marks a significant boundary in terms of the histories of the rocks above and below it; above (within the Cumulate Sequence) all rocks have a cumulate origin, even though tectonite fabrics may now be present, below (excluding the dykes and pods of the Mantle Sequence) rocks show no vestige of a cumulate character.

For the regional mapping of the Cumulate Sequence stratigraphy within the Rustaq Block three mapping units were adopted; Cg, predominantly gabbroic cumulates, Cp, mainly peridotitic cumulates, and Cpg, mixed cumulate peridotites and gabbros. The distribution of the various cumulate types within the Rustaq Block is shown in Figure 2.5 (see also Enclosure 1).
In the following sections a number of traverses from the Rustaq Block are described which were undertaken firstly to assess the vertical variation in rock types through the entire Cumulate Sequence, secondly to determine the degree of lateral variation in phase assemblages developed at the base of the Cumulate Sequence along the Petrological Moho, and thirdly to examine if there is any relation between the rock types occurring in the pods within the Mantle Sequence underlying the Petrological Moho and those rock types present in the Cumulate Sequence above it.
Figure 2.5. Geological sketch map showing the Cumulate Sequence stratigraphy developed in the Rustaq Block. The locations of traverses discussed in section 2.4 are shown.
Figure 2.6. Sections showing the variation in rock types vertically through the Cumulate Sequence. (a) Al Maydah, (b) Wukabah, (c) Wadi Bani Kharus, and (d) Abu Dhabu. Location of sections is shown on Figure 2.5.
2.4.1 Vertical variation in Cumulate Sequence stratigraphy

The vertical variation in terms of rock type and thickness within the Cumulate Sequence is described with reference, from west to east, to four contrasting traverses made within the Rustaq Block, at Al Maydah, Wukabah, Wadi Bani Kharus and Abu Dhabu. The location of the traverses is shown in Figure 2.5.

The character of the Cumulate Sequence along the Al Maydah road section is summarized in Figure 2.6a. The total thickness of layered rocks exceeds 3km. The first 1000m is made up almost entirely of ultramafic cumulates; chromitites, dunites and wehrlites. The uppermost 1000m is nearly all olivine gabbro. Phase layering defined by ol (sp), ol (sp) + cpx, ol + cpx + plag sequences, can be distinguished on scales from 10 to 800m. This type of stratigraphy characterizes the Cumulate Sequence of the Rustaq Block in the area to the west of the NW-SE fault along one tributary of Wadi Bani Suq (Figure 2.5).

Figure 2.6b displays the salient features of the Cumulate Sequence west of Wukabah. Compared to the usual thicknesses developed within the Rustaq Block of up to 3km, this 750m thickness of dunite is far from typical. Phase layering is very limited and is restricted to the top of the section where olivine gabbro layers occur. This type of stratigraphy characterizes the Cumulate Sequence between the NW-SE fault through Wadi Bani Suq, and the NNW-SSE fault through Wukabah (Figure 2.3).

In contrast to the Cumulate Sequence typified by the traverses at Al Maydah and Wukabah, where ultramafic cumulates form a significant proportion of the thickness developed, ultramafic assemblages are absent from the Cumulate Sequence in Wadi Bani Kharus (Figure 2.6c). About 1.5km of olivine gabbros showing only rhythmic layering, rest directly on the harzburgite of the Mantle Sequence. An absence of ultramafic rock types from the Cumulate Sequence typifies the area east of the NNW-SSE Wukabah fault, and to the west of the NNW-SSE Sulawah fault (Figure 2.5).
A return to the type of stratigraphy developed in the western area of the Rustaq Block is shown by the Cumulate Sequence of the Abu Dhabi traverse (Figure 2.6d). With a total thickness of about 3000m, ultramafic rock types once more characterize the lower half of the section. Phase layering is defined by ol (±sp), ol + cpx + plag sequences on a 100-600m scale. Only olivine gabbros are present in the top half of the section. This type of cumulate stratigraphy is typical of the area to the east of the NNW-SSE Sulawah fault (Figure 2.5).

2.4.2 Lateral variation in Cumulate Sequence stratigraphy

Traverses through the entire Cumulate Sequence suggest that significant lateral variation in stratigraphy is present. A readily identifiable datum along which to assess the degree of lateral variation in the Cumulate Sequence is provided by the Petrological Moho. The present section describes, from west to east, more detailed traverses made at six localities through the lowest sections of the Cumulate Sequence, specifically investigating the crystallization order displayed at different positions above the Petrological Moho. The location of the detailed traverses is shown in Figure 2.5.

The character of the base of the Cumulate Sequence at West Moho III is shown in Figure 2.7a. Dunite rests on harzburgite of the Mantle Sequence. Four cyclic units, defined by phase layering, have been distinguished on a 100m scale. In the lower units dunite -> wehrlite -> olivine gabbro sequences are observed. Higher, plagioclase and clinopyroxene simultaneously follow olivine, giving rise to dunite -> olivine gabbro sequences. At the top of cycle III, orthopyroxene follows clinopyroxene and plagioclase, giving a gabbronorite assemblage; olivine is absent. The nature of the cyclic units and textural relationships indicate that the most complete crystallization order in this traverse is ol (±sp) -> cpx -> plag -> opx.

The West Moho II traverse is shown in Figure 2.7b. The dunite at the base of the Cumulate Sequence is considerably thicker than at West Moho III; over 100m are present with thin (less than 5cm) chromitite layers.
occurring at the base. As at West Moho III dunite - wehrlite - olivine gabbro sequences predominate, three cycles being distinguished. At the top of cycle III orthopyroxene is present with olivine, clinopyroxene and plagioclase. Again, on the basis of the nature of the cyclic units and petrography the most complete crystallization order is ol (+ sp) -> cpx -> plag -> opx.

Some eight cycles, on a 50m scale, were distinguished in the traverse at West Moho I (Figure 2.7c). Orthopyroxene was not observed. Dunite - wehrlite - olivine gabbro assemblages predominate, with olivine-free gabbros forming the top of cycles VI and VII. As in West Moho II, thin (less than 5cm) developments of chromitite are present in the lowermost dunite. The most complete crystallization order exhibited in this traverse is ol (+sp) -> cpx -> plag. Pyroxenite dykes are seen to cut and be cut by olivine gabbro pegmatites.

In marked contrast to West Mohos III, II and I, the West Wadi Bani Kharus traverse (Figure 2.8a) shows only a very minor development of ultramafic assemblages. Olivine gabbro rests directly on harzburgite tectonite of the Mantle Sequence. A thin (less than 2m), layer of dunite is succeeded entirely by olivine gabbros. (Reference to Figure 2.6 reveals that olivine gabbros characterize almost the whole of the Wadi Bani Kharus Cumulate Sequence.) Orthopyroxene is absent from this traverse. The crystallization order of clinopyroxene and plagioclase cannot be determined from the character of the layering. Textural evidence from olivine gabbro assemblages indicates an ol -> plag -> cpx order.

Again a minimal development of ultramafic assemblages characterizes the base of the Cumulate Sequence at the East Wadi Bani Kharus traverse (Figure 2.8b). The lowest 4m shows dunite - troctolite - olivine gabbro phase layering on a c.1m scale, allowing confirmation of the ol -> plag -> cpx crystallization order suggested from textural evidence at the West Wadi Bani Kharus Moho.

As with the West Mohos III, II and I, and in contrast to the West and East Wadi Bani Kharus Mohos, the Tawiyah Moho traverse exhibits a
considerable development of ultramafic assemblages in the base of the Cumulate Sequence (Figure 2.8c). About 150m of dunite rests on the harzburgite of the Mantle Sequence; some thin (less than 5 cm) chromitite layers are present at the base, and a relatively well-ordered "matrix layering" on a 30m scale is present as shown by the modal variation of intercumulus plagioclase and clinopyroxene. Both troctolite and olivine gabbro assemblages directly follow dunite. Intercumulus orthopyroxene is present at the top of the lowest dunitic unit. An acceptable generalized crystallization order seems to be ol (sp) -> cpx + plag -> opx, with more or less simultaneous precipitation of clinopyroxene and plagioclase following olivine.

2.4.3 Pod types occurring below the Petrological Moho

The types of pods occurring immediately below the Petrological Moho at the sites of the detailed traverses are shown in Figures 2.6 and 7. Discordant, sharp sided, often pegmatitic gabbroic, gabbronoritic and pyroxenitic dykes and sheets (which are also observed cutting the lower levels of the overlying Cumulate Sequence) are excluded from this discussion as they are considered to be later features, possibly cogenetic with the Later Plutonic Suites (Chapter 3). The characterization of Cumulate Sequence stratigraphy above the Petrological Moho made in section 2.4.2 extends to the distribution of pod types in the upper 1km of the Mantle Sequence. Namely, where ultramafic assemblages are present on the Petrological Moho ultramafic pods (dominantly dunite, sometimes wehrlite) are seen below; where gabbroic assemblages rest on the Petrological Moho, gabbroic pods are present below. In the localities examined within the Rustaq Block the various pod types, whether mafic or ultramafic, were seen to be concordant or nearly concordant with the Mantle Sequence foliation, and to be prevasively foliated by this fabric, as described in section 2.1.2 for the Petrological Moho at Wadi Bani Kharus (Plate 2.7).
Figure 2.7. Lateral variation in Cumulate Sequence stratigraphy as shown by sections through the Petrological Moho at (a) West Moho III, (b) West Moho II, (c) West Moho I. Location of the traverses is shown on Figure 2.5.
Figure 2.8. Lateral variation in Cumulate Sequence stratigraphy as shown by sections through the Petrological Moho at (a) West Wadi Bani Kharus Moho, (b) East Wadi Bani Kharus Moho, and (c) Tawiyah Moho. Location of the traverses is shown on Figure 2.5.
From the foregoing account of traverses typical of the Cumulate Sequence in the Rustaq Block, a number of generalizations can be made. First, the degree of evolution of magmas responsible for the crystallization of the Cumulate Sequence, as reflected by the presence or absence of ultramafic assemblages within any one traverse, is variable. Second, the crystallization order shown by the phase assemblages of the Cumulate Sequence is also variable; within the Rustaq Block both ol -> cpx -> plag and ol -> plag -> cpx sequences are observed. Third, the type of cumulate stratigraphy developed above the Petrological Moho is mirrored in the nature of pod types in the underlying uppermost Mantle Sequence; these pods are nearly always deformed parallel to the harzburgite foliation. Fourth, areas of differing character (in terms of cumulate assemblage evolution and/or crystallization order) are separated by major faults.

2.5 The Cumulate Sequence - High Level Intrusives - Sheeted Dyke Complex transition

The characteristic features of the High Level Intrusives were documented in section 2.3; they are a lack of regular layering structures, heterogeneity in terms of rock type on a small scale, and their thickness is laterally impersistent on a regional scale. The nature of the lower and upper contacts of the High Level Intrusives with the Cumulate Sequence and the Sheeted Dyke Complex respectively is the subject of the present section. Four contrasting traverses (the first from the Rustaq Block, the remaining examples from the Haylayn Block) through these contacts will be described in an effort to emphasize the variability of the Cumulate Sequence - High Level Intrusives - Sheeted Dyke Complex transition.
Figure 2.9. Traverses through the Cumulate Sequence - High Level Intrusives - Sheeted Dyke Complex transition. (a) Wadi Bani Kharus, (b) Musayfiyah, (c) Mashin, and (d) Wadiyah. The location of the sections is shown Figure 2.1.
The location of the Wadi Bani Kharus traverse is shown in Figure 2.1, and the field relations and petrography observed are summarized in Figure 2.9a. In the lower half of the traverse the east to northeast dipping layering in the olivine gabbros of the underlying Cumulate Sequence is seen to vary between 20° to over 65°. Dykes in the overlying Sheeted Complex strike NW-SE and dip at 70° to the southwest. At two points in the lower part of the section the succession of layered gabbros is interrupted by massive isotropic gabbros. No clear cross cutting relationships were seen, and gradational contacts are inferred. However a body of gabbro showing igneous lamination is seen to cut layered gabbro at about 100m from the base of the traverse. Pegmatitic gabbroic pods may show clear intrusive relations with their host. The upper half of the traverse comprises massive, isotropic gabbros and diorites. The contact with the base of the Sheeted Dyke Complex is not seen. A rapid transition, in less than 30m, may be inferred from gabbros containing less than 30% dykes to Sheeted Dykes with less than 30% gabbro screens. Small irregular dykes of trondhjemite cut the isotropic gabbros and the base of the Sheeted Dyke Complex. The mineralogy of the host rocks of the transition is apparently well ordered. Olivine gabbros are present only in the lower half, magnetite bearing gabbros are confined to the upper half. The only quartz or primary amphibole assemblage is that of the small trondhjemite dykes. The degree of hydrothermal alteration increases up section, as reflected by the reaction of clinopyroxene to actinolite, the clouding of plagioclase and the presence of secondary opaque minerals.

The location of the Wadi Musayfiyah traverse is shown in Figure 2.1. The field relations and petrography are summarized in Figure 2.9b. As with the Wadi Bani Kharus transition, the attitude of layering steepens up section in the Cumulate Sequence, and the lower half of the traverse is punctuated by the presence of massive, isotropic gabbro. The attitudes of compositional layering vary between 35-85°, dipping to the north or northwest, whilst the dykes of the Sheeted Complex strike NW-SE, dipping at 60° to the southwest. In contrast to the Wadi Bani Kharus traverse however, the upper half of the section is characterized by the occurrence
of dolerite xenoliths and the development of a significant volume of trondhjemite. The topmost trondhjemite is clearly intrusive into the base of the overlying Sheeted Dyke Complex, and the xenoliths of dolerite would appear to have been derived by stoping. The mineralogy of the transition is more diverse than at Wadi Bani Kharus. Olivine gabbros are succeeded by magnetite bearing gabbros and then a quartz bearing trondhjemite assemblage. Primary amphibole is present in the upper half of the transition. Again, the degree of hydrothermal alteration of primary igneous phases increases up section.

The location of the Mashin traverse is shown in Figure 2.1, and the petrography and field relations observed is depicted in Figure 2.9c. Unlike the Wadi Bani Kharus and the Wadi Musayfiyah traverses where the Cumulate Sequence to High Level Intrusives transition occurs over 700m and 175m respectively, the transition at Mashin is very rapid, occurring within 20m. Nevertheless a similar crude stratigraphic order is present in that steeply dipping layering of the Cumulate Sequence gives way to massive gabbros which become xenolithic with height. Layers of the Cumulate Sequence dip north at 65°, whilst the dykes of the Sheeted Complex strike NW-SE and dip south at 70°. Here the upper contact with the base of the Sheeted Dyke Complex can conclusively be demonstrated to be transitional. Dyke abundance increase from less than 10% to 100% in less than 10m. The degree of assimilation of xenoliths increases with depth, and in contrast with the Wadi Musayfiyah traverse, gabbroic and ultramafic assemblages are present as well as doleritic fragments. Again olivine gabbros occupy the top of the Cumulate Sequence, and the isotropic gabbro and diorite are magnetite bearing. Primary amphibole is present as an intercumulus phase in the top of the Cumulate Sequence, and in the more evolved isotropic diorites.

The location of the Wadiyah traverse is shown in Figure 2.1, and the field relations are shown in Figure 2.9d. Once more layering in the top of the Cumulate Sequence gives way to massive diorite and trondhjemite. However, in this example the attitude of the layering is near 30° to the northeast, rather than angles of nearer 60° that characterize the other
sections described. The dykes of the Sheeted Complex are vertical and strike NW-SE. As with the Mashin traverse the complete transition from Cumulate Sequence to Sheeted Dyke Complex is rapid, occurring in about 30m, the contact at the base of the Sheeted Dyke Complex being unequivocally transitional. The massive diorite and trondhjemite although unlayered, cannot be described as "isotropic". On a metre scale they are very heterogeneous (Plate 2.20), with much rapid variation in grain size and in the ratio of mafic to felsic constituents ("varitextured"). Xenoliths of dolerite again are present in the upper portion of the diorite and trondhjemite. They show varying degrees of assimilation, and later dykes of dolerite can be seen (Plate 2.21) cutting xenoliths of partially resorbed roof rocks, which witness the continual history of dyke injection, stoping, assimilation and reinjection that occurred within the transition. The base of the transition is cut by a later trondhjemite pluton, part of a Late Trondhjemite - Gabbro Complex (see Chapter 3).

Plate 2.20. Varitextured diorite-gabbro typical of the High Level Intrusives. From the Wadiyah traverse.
Plate 2.21. Xenoliths showing varying degrees of assimilation in a trondhjemite-diorite host. Dykes related to the overlying Sheeted Complex cut both xenolith and host. From the High Level Intrusives, Wadiyah traverse.

The foregoing accounts of the field relations and petrography of the Cumulate Sequence-High Level Intrusives-Sheeted Dyke Complex transition underline its heterogeneous nature, both on a local and regional scale. It may vary in thickness between 10m and 700m, in the occurrence and composition of xenoliths, in the presence or absence of primary amphibole, in the presence or absence of significant volumes of trondhjemite, and in the intrusive or transitional nature of the contact at the base of the Sheeted Dyke Complex. Generalisations that may be made are that the attitude of the layering in the top of the Cumulate Sequence is often steep (60° or more) and that this passes upwards into massive, usually isotropic gabbro and diorite. Olivine gabbro and magnetite bearing gabbro and diorite are the assemblages that predominate. The degree of hydrothermal alteration increases upwards through the transition.
Chapter 3
The Later Plutonic Suites

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Chapter 3
The Later Plutonic Suites

3.1 Introduction

Within the Haylayn Block three successively later suites of plutonic rocks cut the classic Penrose ophiolite stratigraphy (see Figures 3.1 and .2, and Enclosure 1). In chronological order they are named:

(i) the Late Trondhjemite-Gabbro Complexes,
(ii) the Late Peridotite-Gabbro Complexes, and
(iii) the Late Biotite Granites.

The present chapter describes the petrography and field relations of each intrusive suite by reference to specific examples studied within the Haylayn Block, and discusses their likely tectonic settings.

3.2 Late Trondhjemite-Gabbro Complexes

The distribution of Late Trondhjemite-Gabbro Complexes within the Haylayn Block is shown in Figure 3.2. Ranging up to 6km in extent, they are seen to intrude the higher stratigraphic levels of the ophiolite (the Cumulate Sequence, the High Level Intrusives and the Sheeted Dyke Complex). Gabbros, diorites and trondhjemites are the dominant component rock types, often containing xenoliths derived from the ophiolite stratigraphy that the complexes intrude. The complexes form white weathering areas of low relief (Plate 3.1).
Figure 3.1. Diagram to show the cutting relationships of the successive later plutonic suites.
Plate 3.1. The Budit Late Trondhjemite-Gabbro Complex, looking northwest from the southeast end of the intrusion. The white low weathering ground in the foreground and middle distance is trondhjemite-gabbro of the late complex; the roof contact can be seen in the flanks of the ridge of Cumulate Sequence that forms the background.

The gabbros may be varitextured or more uniform in grain size; the latter are medium grained (up to 3mm) and show no layering structures. They are ophitic or sub-ophitic in thin section, with randomly oriented laths of plagioclase enclosed by clinopyroxene or a primary actinolitic amphibole. Amphibole may occur as rims to clinopyroxene, or may be present as interstitial subhedra. Typical modes vary between 8-21% clinopyroxene, 18-34% amphibole, about 60% plagioclase, with accessory oxide phases.

The diorites are similar in grain size and texture to the gabbros. Clinopyroxene is however absent, and interstitial anhedra of quartz and Fe-Ti oxide become prominent. Characteristic modes are amphibole 38%, plagioclase 53%, quartz 6% and oxide 3%.
Figure 3.2. Geological sketch map to show the distribution of the various late plutonic bodies that intrude the ophiolite in the Haylayn and Rustaq Blocks.
Plate 3.2. Photomicrograph of typical sub-ophitic gabbro from the Wadiyah Late Trondhjemite-Gabbro Complex. Sample 2636, field of view 8mm.

The trondhjemites are finer grained than the gabbros and diorites, not exceeding 2mm in grain size. Myrmekitic intergrowths of quartz and plagioclase are common, as are acicular actinolitic amphibole and zoned plagioclase laths. Zircon or sphene may be present as accessory phases. Typical modes vary between 3-12% amphibole, 41-53% plagioclase, 39-44% quartz, and 3-4% oxides.

All the Late Trondhjemite-Gabbro Complexes have suffered hydrothermal alteration; the trondhjemite members, in particular, can be severely altered. Ragged, fibrous secondary actinoLite replaces both clinopyroxene and primary actinolitic amphibole; chlorite and secondary oxide also develop. Breakdown of plagioclase to albite, epidote and calcite leaves only a quartz-rich residue to the original mineralogy of a trondhjemite.

The similarity of the rock types occurring in the Late Trondhjemite-Gabbro Complexes with those typical of the High Level
Intrusives makes certain field discrimination difficult when they are juxtaposed. The two largest Late Trondhjemite-Gabbro Complexes occurring in the Haylayn Block (at Wadiyah and Budit, see Figure 3.2), having been intruded at the level of the Cumulate Sequence-High Level Intrusives-Sheeted Dyke Complex transition, and are described in the sections below, highlight this problem.

3.2.1 Wadiyah Late Trondhjemite-Gabbro Complex

The geological map of the Wadiyah Late Trondhjemite-Gabbro Complex is shown in Figure 3.3. Lying along the WNW-ESE regional strike of the ophiolite, the complex is 4.5km long, and is bounded by the Cumulate Sequence to the southwest and the Sheeted Dyke Complex to the northeast. Over two-thirds of the complex consists of trondhjemite that intrudes an earlier, but associated, massive gabbro. Two facies of trondhjemite occur within the major body; a finer more felsic type and a slightly darker, coarser variety with higher modal amphibole. Small (less than 250m) massive wehrlite plugs,
part of the Late Peridotite-Gabbro Complex suite, cut both the trondhjemite and gabbro (see Plate 3.4), and the host ophiolite stratigraphy.

Plate 3.4. A massive, dark weathering plug of peridotite, part of the Late Peridotite-Gabbro Complex suite, cutting the contact between trondhjemite and gabbro of the Wadiyah Late Trondhjemite-Gabbro Complex. The intrusive contact of the white weathering trondhjemite into the massive gabbro can clearly be seen.

Smaller trondhjemite, gabbro and peridotite bodies lying to the east of the main Wadiyah Complex (Figure 3.3) are fault controlled; whether the W-E trending structure has influenced the emplacement of the main complex is not clear.
Figure 3.3. Geological sketch map of the Wadiyah Late Trondhjemite-Gabbro Complex.
The field relations observed on the northern margin of the southeastern end of the intrusion are critical in demonstrating that the trondhjemite and gabbro of the Wadiyah complex are not part of the High Level Intrusives of the normal ophiolite stratigraphy; they are shown schematically, in cross-section, in Figure 3.4. A typical gradational, but rapid, transition from Sheeted Dyke Complex through High Level Intrusives to Cumulate Sequence is developed (see Chapter 2). Within the zone of dykes with gabbro screens (D1) local intrusive contacts are seen; in particular small bodies of trondhjemite and varitextured gabbro carrying assimilated xenoliths of dolerite, are seen to cut, and are cut by dykes related to the overlying sheeted complex. The dykes within the High Level Intrusives are truncated by the later massive gabbro body (itself very similar to the isotropic gabbros of the High Level Intrusives), which is in turn cut by the large trondhjemite body. The Late Trondhjemite-Gabbro Complex is devoid of dykes related to the Sheeted Dyke Complex and is therefore distinctly later than the High Level Intrusives.

The field relations of the southern margin of the Wadiyah Late Trondhjemite-Gabbro Complex are considerably more complex and ambiguous than on the northern margin. Dykes related to the overlying Sheeted Dyke Complex, which provided critical time markers, are absent from the host rock. Much of the outcrop area comprises fine grained, varitextured or pegmatitic gabbro and diorite, carrying up to metre sized xenoliths of sometimes massive, sometimes layered, mafic and ultramafic cumulates, and smaller, more assimilated fragments of dolerite (Plate 3.5). The form of these xenolithic gabbros and diorites is highly irregular; they are considered to represent the pervasive intrusive equivalents of the massive gabbro body that occurs on the northern margin of the complex. From the rock types they contain, extensive stoping and varying degrees of assimilation of the Cumulate Sequence, the High Level Intrusives, and the Sheeted Dyke Complex may be inferred to have occurred.

Numerous sheets of fine trondhjemite and plagioclase-amphibole pegmatite of variable orientation cut the xenolithic gabbros, diorites and
Figure 3.4. Rock-relation diagram to show the critical field relationships that demonstrate the intrusive nature of the massive gabbros of the Wadiyah Late Trondhjemite-Gabbro into the High Level Intrusives of the ophiolite stratigraphy.

their host rocks. Substantial (up to 3m wide) fine, clinopyroxene-phyric, basic dykes striking WNW-ESE, and smaller, less numerous, plagioclase-phyric, more acid dykes of similar orientation record the final intrusive event in the complex; they are seen to cut the late massive peridotites that intrude both the main Wadiyah Late Trondhjemite-Gabbro Complex and the host
Plate 3.5. Part of an irregular body of xenolithic, varitextured gabbro carrying fragments of mafic and ultramafic rocks derived from the host Cumulate Sequence stratigraphy.

ophiolite stratigraphy.

3.2.2 Budit Late Trondhjemite-Gabbro Complex

The geology of the Budit Late Trondhjemite-Gabbro Complex is shown in Figure 3.5 (and see Plate 3.1). In this area two NW-SE trending strike faults that downthrow to the southwest, give rise to outcrop repetition of the normal ophiolite succession; in both cases the Sheeted Dyke Complex or the High Level Intrusives are downthrown against gabbros of the Cumulate Sequence.
Figure 3.5. Geological sketch map of the Budit Late Trondhjemite-Gabbro Complex.
Along the more southerly of the two strike faults, an elongate body of trondhjemite and gabbro measuring 6km in length, has been intruded. On its southwestern margin it cuts the Cumulate Sequence and the High Level Intrusives, on its northeastern margin it intrudes the Sheeted Dyke Complex and upthrown gabbros of the Cumulate Sequence. At the northeastern corner of the intrusion a 1.5km long body of massive wehrlite (part of the Late Peridotite-Gabbro Complex suite) cuts the contact between the body of trondhjemite-gabbro and the Sheeted Dyke Complex, demonstrating clearly the age relations.

Along the more northerly of the two strike faults a smaller (750m) trondhjemite-gabbro has been intruded; to the southeast of this body another massive wehrlite has been emplaced into the Sheeted Dyke Complex and upfaulted gabbros of the Cumulate Sequence. The area of Sheeted Dyke Complex between the two strike faults also contains smaller (less than 750m) trondhjemite-gabbro bodies.

The trondhjemitic and more gabbroic components of the Budit Late Trondhjemite-Gabbro Complex are not differentiated on Figure 3.5; distinction was not possible due to the rapid and apparent lack of systematic variation between rock types, though overall, trondhjemite predominates. Typically areas of more mafic diorites and gabbros pass gradationally into trondhjemite; in some instances the trondhjemite veins the more mafic rocks. The distribution and proportion of xenoliths in the predominant trondhjemite is highly variable. Two types of xenoliths most commonly occur; dark, angular, metre sized fragments of dolerite, and lighter coloured, coarser, rounded masses of mafic trondhjemite. The "softer" nature of the contacts between the areas of mafic trondhjemite, diorite and gabbro as compared to the sharp sided, angular blocks of dolerite, might suggest that the former are autoliths, i.e. cogenetic with the predominant trondhjemitic body, having formed as earlier fractionates. However, considering the structural position of the Budit Late Trondhjemite-Gabbro Complex, the alternative that they are xenoliths derived from the host ophiolite stratigraphy cannot be excluded on the basis of field evidence.
Dykes and sheets of trondhjemite, often pegmatitic, of variable orientation cut the adjacent ophiolite stratigraphy. These dykes, the main trondhjemite-gabbro body, and the massive wehrlites related to the Late Peridotite-Gabbro Complex suite are cut by later fine, basic, plagioclase or clinopyroxene-phyric dykes trending between NE-SW and N-S (Plate 3.6).

Plate 3.6. A fine, basic, clinopyroxene phryic dyke (running away from camera) cutting pegmatitic trondhjemite dykes (striking across picture). The latter emanates from the Budit Late Trondhjemite-Gabbro Complex. The host rocks are gabbros of the higher Cumulate Sequence.

3.3 Late Peridotite-Gabbro Complexes

Figure 3.2 shows the distribution of Late Peridotite-Gabbro Complexes within the Haylayn Block. Up to 1.5 km in size, they are invariably associated with WNW-ESE trending strike faults, and cut the higher levels of the ophiolite stratigraphy. As noted in the preceding sections the massive wehrlites are often found in conjunction with the earlier Late Trondhjemite-Gabbro Complexes, which they are seen to cut.
Their distinctive massive, brown, and spheroidally weathered appearance lends to easy field recognition (see Plate 3.4).

Typically poikilitic, the wehrlite bodies usually have coarse, sometimes pegmatitic, gabbroic margins up to 1.5m in width (Plate 3.7). In strong contrast with the Late Trondhjemite-Gabbro Complexes the Late Peridotite-Gabbro Complexes are homogeneous and xenolith-free, excepting occasional dykes or sheets of pegmatitic gabbro.

Plate 3.7. Gabbroic envelope to massive wehrlite of the Late Peridotite-Gabbro Complex suite. The host rock, to the lower right of picture, is Sheeted Dyke Complex.

Only one example has been mapped that deviates from the typical pattern of a massive wehrlitic body with a gabbroic margin; this is the Mashin Late Peridotite-Gabbro Complex and is described below.

3.3.1 Mashin Late Peridotite-Gabbro Complex

The location of the Mashin Late Peridotite-Gabbro Complex is shown in Figure 3.2, and the geology of the intrusion is illustrated in Figure 3.6. Elongate WNW-ESE, 1.5km in length and not exceeding 250m in outcrop width, the complex is a sill-like body comprising a massive ultramafic wehrlitic
base overlain by a succession of more felsic rock types, from gabbro through to trondhjemite. Its southern margin is in faulted contact with the Sheeted Dyke Complex; small outcrops of layered olivine gabbro present along this margin are considered to be upfaulted slivers of the Cumulate Sequence. Along its northern margin trondhjemite is seen to intrude the Sheeted Dyke Complex.

A gross compositional layering, dipping at c. 45° to the NNE, is present in the Mashin Late Peridotite-Gabbro Complex, and is best developed at the northwestern end of the intrusion (see Figure 3.7). The lower contact of the body is not exposed; at least 50m of dominantly massive wehrfite forms the base of the complex. Oikocrystic in texture, well rounded 1mm sized olivine grains sit in heterads of clinopyroxene of up to 4mm in size (Plate 3.8). The resorbed olivine grains appear to be matrix supported by the clinopyroxene heterads. A pargasitic amphibole, showing colourless to dark brown pleochroism, is present as an intercumulus phase as reaction rims to clinopyroxene; small (less than 0.1mm) accessory chrome spinel grains may occur. Typical modes are olivine 68%, clinopyroxene 31%, and amphibole 1%.

In the uppermost ten metres of the wehrfite, amphibole becomes more abundant and is joined by plagioclase as an intercumulus phase. In particular, where the mode of plagioclase is high, schlieren of feldspathic wehrfite are observed, elongate parallel to the gross compositional layering of the intrusion (Plate 3.9). In sample 1405 all the clinopyroxene oikocrysts appear to have reacted with interstitial melt to form pargasite, and some minor primary actinolite (exhibiting colourless to pale green pleochroism) occurs as rims on the earlier amphibole. Euhedra of chrome spinel are present, up to 0.3mm in size. Modal percentages are olivine 52%, clinopyroxene 41%, and amphibole 7%.
Figure 3.6. Geological sketch map of the Mashin Late Peridotite-Gabbro Complex.

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Plate 3.8. Photomicrograph of poikilitic wehrlite. Small resorbed grains of olivine set in a heterad of clinopyroxene. Sample 1405, field of view 8mm.

A two metre thickness of ratio layered (on a decimetre scale) gabbro overlies the poikilitic wehrlite. Showing a good plagioclase lamination parallel to compositional layering, the rock is a fine grained (1mm size) adcumulate, with modes of plagioclase 67% and clinopyroxene 33%. The plagioclase is zoned, the clinopyroxene shows reaction to primary actinolitic amphibole, and scattered oxide grains are present.

The ratio layered gabbro passes upwards into a coarser (up to 4mm size) diorite. Showing good lamination of acicular, zoned plagioclase and amphibole, a sub-ophitic texture is developed. The amphibole occurs as crystalline grains or fibrous mats of clear to pale green actinolite; the former is considered primary igneous actinolite, the latter a hydrothermal alteration product of the former. The oxide mineral present is ilmenite. Typical modes are plagioclase 55%, amphibole 44% and opaques 1%.

The laminated diorite, some 90 metres in thickness, is succeeded by 30 metres of isotropic diorite which shows no preferred orientation of
Plate 3.9. Feldspatic schlieren in wehrlite that forms the base of the Mashin Late Peridotite-Gabbro Complex. The light weathering rock is a late basic sheet.

plagioclase or amphibole. As the strength of the lamination decreases upwards, the mode of plagioclase increases to 64%, and the mode of oxides to 4%, whilst the mode of amphibole decreases to 32%.
Figure 3.7. Sections through the northwest, central and southeast parts of the Mashin Late Peridotite-Gabbro Complex.
The top of the intrusion comprises a 35m thick trondhjemite. Of medium grain size (2mm), sometimes varitextured, the rock contains acicular prisms of randomly oriented plagioclase and primary actinolitic amphibole (often showing secondary alteration to fibrous habits), with interstitial subhedra of quartz and Fe-Ti oxides. Myrmekitic intergrowths of quartz and plagioclase are common. On a metre scale the trondhjemite is varitextured, showing variations in modes from 13-25% quartz, 49-59% plagioclase, 16-34% amphibole, and 1-4% for oxides.

The degree of low temperature alteration varies through the intrusion from moderate to high. The wehrlite shows extensive serpentinization of olivine, and breakdown of intercumulus plagioclase when present. The overlying felsic assemblages show clouding of plagioclase, alteration of amphibole to fibrous actinolite, and the development of secondary oxides. The degree of feldspar breakdown is most intense at the top of the intrusion where almost pure secondary albite coexists with calcite and epidote.

Both fine grained and pegmatitic sheets of trondhjemite cut all levels of the complex and the adjacent host rocks of the Sheeted Dyke Complex. Within the isotropic diorite up to one metre sized sweets of trondhjemite with gradational margins are observed. Evidence of stoping of roof rocks is provided by xenoliths (up to 30 cm in size) of dolerite, which show moderate assimilation. The latest igneous event is recorded by metre sized fine, aphyric basic dykes that cut both the intrusion, the trondhjemite sheets, and the country rocks.

In the central part of the intrusion a similar, but condensed stratigraphy is developed (Figure 3.7b); however, some significant differences are present. The feldspathic top to the massive wehrlite base is replaced by a 10 metre thick meta-gabbronorite assemblage of olivine, clinopyroxene, orthopyroxene, amphibole, spinel and plagioclase. In the northwestern part of the complex orthopyroxene was not observed as a cumulate or intercumulate phase. Like clinopyroxene, it is oikocrystic and shows reaction to brown amphibole. The overlying 3 metre thick gabbro
Layer shows the development of intercumulate brown amphibole (again not seen in the northwest of the intrusion), and is succeeded by laminated and then isotropic diorite and quartz diorite.

At the southeastern end of the complex orthopyroxene is again present as a cumulate phase (Figure 3.7c), in particular in the massive peridotite base. Intrusive sheets of quartz diorite cutting gabbro form the top of the intrusion, rather than a gradation from gabbro through diorite to trondhjemite.

3.4 Late Biotite Granites

The distribution of Late Biotite Granites within the Haylayn Block is shown in Figure 3.2. Small elongate bodies, no greater than 100m in width and not exceeding 500m in length, are found cutting the lower levels of the ophiolite stratigraphy, the Mantle Sequence and the lower Cumulate Sequence. They are associated with WNW-ESE trending faults.

Plate 3.10. Body of Late Biotite Granite cutting rocks of the Lower Cumulate Sequence at Al Haymilliyah.
In the field they form white weathering bodies, especially distinctive when the host rocks are ultramafic (Plate 3.10). In hand specimen they may be foliated to some degree, due to the preferential orientation and modal sorting of mica grains. In thin section they are fine grained (up to 1mm in size), with a mosaic of quartz anhedral and weakly laminated, well formed plagioclase laths, amphibole prisms, and mica flakes (Plate 3.11). A typical mode is quartz 50%, plagioclase 23%, amphibole 5%, and mica 2%. The quartz grains show undulose extinction, the cores of the plagioclase grains are corroded, and both minerals sometimes are intergrown. The amphibole prisms are well formed, length slow, with $\gamma^z 18^\circ$. They have moderate $c.50^\circ 2V_{\text{alpha}}$, and the pleochroic scheme is $\gamma$ blue-green, $\beta$ green, and $\alpha$ pale-green. Simple and lamellar twinning is displayed. The flakes of mica often show alteration to chlorite, display a $\gamma$ brown, $\beta$ brown, $\alpha$ pale brown pleochroic scheme, and have a very small $2V_{\text{alpha}}$. Unidentified accessory heavy mineral grains are present.

Plate 3.11. Photomicrograph of Late Biotite Granite. Sample 2697, field of view 8mm. Annotate.
3.5 Tectonic setting

Three successively later plutonic suites have been described that intrude the classic Penrose ophiolite stratigraphy. Their likely tectonic settings are now considered.

U-Pb zircon ages in the range 94-98Ma were obtained by Tilton et al. (1981) for trondhjemites from both the High Level Intrusives of the ophiolite stratigraphy proper and those from the Late Trondhjemite-Gabbro Complexes, rendering the later suite indistinguishable from the major spreading ridge phase on an isotopic age basis. Alabaster et al. (1980) and Pearce et al. (1981) have distinguished the trondhjemites from the main spreading phase and those of the later suites on the basis of their trace element compositions with lower incompatible element abundances in the trondhjemites of the later suites. These conclusions are supported by whole rock data from early and late trondhjemites from the Haylayn and Rustaq Blocks (Browning in prep.)

Micas from the Late Biotite Granites (from Khawr Fakkan, U.A.E) give a K-Ar age of 85Ma (Rex pers. comm), a date that coincides with ages of 83 and 86Ma for the greenschist facies rocks of the sub-ophiolite metamorphic sheet (Allemann and Peters 1972), which are considered to have been formed during emplacement of the ophiolite onto the Arabian continental margin (Searle 1980, Searle and Malpas 1980). Potassic granites have also been reported cutting the Masirah Island ophiolite, on the south-east coast of Oman (Abbotts 1978).

The Late Trondhjemite-Gabbro Complexes and the Late Peridotite-Gabbro Complexes are considered cogenetic with the volcanic Lasail and Alley Units respectively; these overlie the major spreading-phase-related Geotimes Unit of the Extrusive Sequence (Alabaster et al. 1980, Pearce et al. 1981, Browning and Smewing 1981).

On the basis of trace element geochemistry Alabaster et al.(in press) identify likely tectonic settings for each of the lava stratigraphic units (Figure 3.8). The Geotimes Unit was produced during sea-floor
spreading in a marginal basin, the Lasail Unit was erupted as discrete seamounts during incipient island arc formation, the Alley Unit was formed during arc-rifting, and the Salahi Unit was produced during continent-arc collision.

Alabaster et al. (in press) consider the Geotimes and Lasail Units to have been derived by similar degrees of melting of a selectively enriched mantle source that overlay a subduction zone. The contrast in incompatible element abundances between these lava groups reflects their different fractionation histories; the Geotimes Unit was supplied by liquids that underwent open system fractionation in a sub-spreading ridge magma chamber, which produced their high levels of incompatible element abundances; the Lasail Unit conversely, fractionated in discrete closed system magma chambers. The arc-rifting Alley lavas also formed in discrete closed system magma chambers, but were derived from a depleted mantle source. The syn-collision Salahi Unit lavas were however derived from an enriched mantle source that underlay the passive Arabian continental margin.
Figure 3.8. Magmatic and tectonic evolution of the Oman Ophiolite (from Alabaster et al. in press). A="Axis" spreading event and eruption of Geotimes Unit; B=development of Lasail Unit during incipient arc formation; C= graben formation during arc-rifting and eruption of Alley Unit; D=syn-collision volcanism producing Salahi Unit. The equivalent plutonic suites are indicated. (The Salahi Unit and the Late Biotite Granites are considered contemporaneous but not cogenetic.)

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<tr>
<th>Number</th>
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<tr>
<td>1</td>
<td>Dewathering of subducted plate</td>
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<td>2</td>
<td>Partial melting 25%</td>
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<td>3</td>
<td>Ascent of magma and olivine-cpx-spinel fractionation</td>
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<td>4</td>
<td>Open system 'axis' magma chamber</td>
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<td>5</td>
<td>Eruption of Geotimes Unit</td>
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<td>6</td>
<td>Partial melting of depleted mantle source</td>
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<td>7</td>
<td>Upward movement of magma</td>
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<td>8</td>
<td>High level closed system</td>
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<td>9</td>
<td>Eruption of Lasail Unit</td>
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<td>10</td>
<td>Upwelling of asthenosphere to compensate downward movement of subducted slab</td>
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<td>11</td>
<td>Melt from subducted slab or asthenosphere</td>
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<td>12</td>
<td>Small high level chambers centred beneath graben</td>
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<td>13</td>
<td>Eruption of Alley Unit</td>
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<td>14</td>
<td>Melting of subducted asthenosphere</td>
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<td>15</td>
<td>Melts migrate through subducted plate</td>
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<td>16</td>
<td>Magma collects in obducted plate</td>
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<td>17</td>
<td>Eruption of Salahi Unit</td>
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<td>Path of melt or diapir</td>
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<td>19</td>
<td>Plate movement</td>
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<td>24</td>
<td>Melting zones</td>
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<td>25</td>
<td>Enrichment of H$_2$O + LILE</td>
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Support for the closed system character of the Lasail and Alley Unit episodes comes from the highly differentiated nature of the Late Trondhjemite-Gabbro and Peridotite-Gabbro Complexes. Browning (in prep.) considers the Late Peridotite-Gabbro Complexes to have formed from diapirs of wehrlitic crystal mush that rose along faults during arc-rifting; the interstitial liquid segregated to the margins and top of the diapir, and on emplacement could differentiate from olivine gabbro through to trondhjemite. The Late Biotite Granites were intruded along faults formed (or rejuvenated) during ophiolite emplacement.

In Chapter 2 a distinction was drawn between the deformed, soft-margin dunite and olivine gabbro pods within the Mantle Sequence, and the clearly cross-cutting, sharp-sided, generally undeformed, and often pegmatitic orthopyroxenite, clinopyroxenite and gabbnororite dykes that cut the Mantle Sequence and Lower Cumulate Sequence. It is quite possible that these late stage dykes are the deeper seated fractionates of the Later Plutonic Suites, but their precise petrogenetic relationships have not been fully established (Browning in prep.).

Preliminary Nd and Sr isotopic studies on the successive igneous suites in Oman (Dunlop and Browning in prep.) indicate that the main plutonic series of the ophiolite proper, the Late Trondhjemite-Gabbros and the Peridotite-Gabbros were all derived from an oceanic mantle source, containing varying degrees of arc-component. No isotopic distinction can be made between these plutonic suites as the heterogeneity within any one suite appears as great as the variation between all three. The Late Biotite Granites however, show clear continental affinities.
4.1 Introduction

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4.3 Grain size and orientation measurements

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4.6 Phase chemistry
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4.7 The formation of layering
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   4.7.4 Mechanisms of rhythmic layer formation
4.1 Introduction

Rhythmic layering is ubiquitous within the Cumulate Sequence, occurring on a decimetre to metre scale (Plate 4.1). In an attempt to establish whether the mineral sorting within such layers is due to mechanical or chemical processes, a rhythmic unit from Wadi Shafan (G.R. 477326343), measuring 37 cm in thickness, was investigated to determine if the rapid modal variations that define the rhythmic layering were accompanied by similar textural and geochemical variations.

Plate 4.1. Rhythmic layering on a decimetre scale in olivine gabbro, Wadi Shafan. The darker bases to each unit are rich in olivine, the lighter weathering tops are rich in clinopyroxene and plagioclase.
4.2 General petrography

Eight specimens were collected, effectively sampling the top of the underlying and the bottom of the overlying rhythmic units, as well as the unit of interest itself. The rhythmic unit and positions of the samples taken are shown schematically in Figure 4.1. Where size permitted, three orthogonal sections were cut from each specimen.

![Figure 4.1. Schematic section through the Wadi Shafan rhythmic unit showing sample positions. The graph shows the variation in the modes of olivine, clinopyroxene and plagioclase with height.](image)

At all levels the rhythmic unit comprises medium-grained, laminated, meso- or adcumulates showing a simple three phase cumulate assemblage of olivine, clinopyroxene and plagioclase. The variation in modal proportions through the unit is also shown in Figure 4.1. A decrease in the proportion of olivine with an antipathetic increase in plagioclase and clinopyroxene is observed. Above the level of sample 1425 the mode of clinopyroxene increases at the expense of plagioclase also (Figure 4.1).
At the base of the rhythmic unit a sharp ratio contact is seen (Plate 4.2). Though irregular on a microscopic scale (having a relief of the order of the local grain size), no grains are seen to grow across the contact. The top quarter of sample 1423 displays a local ratio contact (Plate 4.3), the proportion of olivine decreasing from 54% to c.30%, giving a bimodal distribution of olivine to the rhythmic unit as a whole.

Olivine grains frequently display a resorbed character, indicating a reaction relationship between olivine, melt and clinopyroxene (wehrlitic heteradcumulates elsewhere in the Cumulate Sequence testify to such a relation), the degree of resorption increasing with height in the rhythmic layer.

Clinopyroxene grains, particularly in the top half of the unit, display a "vermicular" unmixing (Plates 4.4 and .5). Blebs or larger domains (with differing optical orientation) of clinopyroxene sit in a host of chemically identical clinopyroxene (see below). (The use of the term "unmixing" here is misleading; it is a structural rather than chemical unmixing (or sub-solvus exsolution) that occurs.)

4.3 Grain size and orientation measurements

If mechanical processes were responsible for layer formation it might be expected that the grain size and shape distributions, and the strength of grain orientation in terms of lamination or lineation, would vary with height in the layer in a way that reflected the strength and constancy of direction of the transporting medium or the degree of compaction following deposition. In order to explore such distributions, maximum and minimum dimensions were measured on about ten of each of the larger olivine, clinopyroxene, and plagioclase grains from the available orthogonal sections. The assumption was that every mineral grain could be approximated to a rectangle in two dimensions (i.e. they possessed orthorhombic symmetry in three dimensions). In addition, orientation of the long axes of the mineral grains was recorded.
Plate 4.2. Sharp, irregular ratio contact that defines the olivine rich base of the rhythmic unit onto the plagioclase rich top of the underlying unit (Samples 1421 and 1422). Field of view 8mm.

As noted by Dowty (1980), determination of grain size distributions in thin sections is not straightforward. The mean dimension alone does not provide a satisfactory description of grain size. Grain size distributions in thin sections are biased towards smaller grain sizes; if the grains are spherical then the correction is relatively straightforward (Krumbein and Pettijohn 1938). However the problem becomes much more difficult if the grain shape is complex, and although solutions for common grain shapes have been offered (Gray 1970), the approach here has been very simplistic, the aim being to determine relative differences in grain size between different minerals of different samples, rather than to be concerned with an accurate absolute measurement of grain size distribution.

To determine a realistic grain size in any rock sample showing mineral lamination and/or lineation, sections must be cut in the plane containing the lamination or lineation. (Macroscopically the samples show no evidence of a lineation.) The limited size of some of the specimens
Plate 4.3. Local ratio contact within rhythmic unit, defined by sudden decrease in mode of olivine from 54% to 30% (Sample 1423). Field of view 8mm.

prevented sections in the plane of layering/lamination being cut for all levels in the rhythmic unit; about half of the grain size measurements were determined on sections normal to the plane of layering/lamination. However, for samples (1423, 1424 and 1427) in which determinations were made both normal and parallel to the plane of layering/lamination, little difference in the mean grain dimension is observed (see Figures 4.2 and 4.3), though absolute minima and maxima may differ considerably. Therefore, measurements on samples for which mean grain dimensions were determined only in sections normal to layering/lamination are still considered representative.

In comparing the grain sizes of minerals within any one specimen or between specimens, one is forced to assume that any processes that may have led to a change in grain size subsequent to layer formation (for instance, in terms of classic cumulus theory, adcumulus growth) have changed the maximum and minimum dimensions of grains of different minerals at different positions within the layer to the same degree.
The variation of maximum grain dimension with height in the rhythmic unit is shown in Figure 4.2. The mean maximum grain size overall varies between 1 and 2mm for all phases. In the bottom quarter of the rhythmic unit olivine, clinopyroxene and plagioclase have similar mean grain sizes; in the top three-quarters plagioclase is consistently larger than clinopyroxene, which is in turn larger than olivine. For clinopyroxene and plagioclase a reverse size grading is suggested in the lower third of the rhythmic unit, with no obvious trend in sorting with respect to grain size in the upper two-thirds. Olivine however, would seem to decrease in grain size with height throughout the rhythmic unit. This observation lends support to the earlier suggestion that the degree of resorption of olivine grains increases with height. A similar picture emerges for the mean minimum dimensions of all mineral phases (Figure 4.3).

The scale of sampling is inadequate to reveal exactly where the change from reverse to no obvious size grading occurs in the lower third of
the rhythmic unit; it is tentatively suggested that the change may coincide with the local ratio contact that was noted at the top of sample 1423. Unfortunately the area of thin section that covers the uppermost part of specimen 1423 is too small to confirm or deny this.

Although the samples that were collected were not oriented with respect to each other, information about the degree of alignment of mineral grains at a particular level in the rhythmic unit was collected by measuring the orientation of the mineral grain long axes relative to an arbitrary direction. The degree of dispersion (or one standard deviation) of the measured directions of long axes should reflect the intensity of mineral lamination (in sections cut normal to layering) or mineral lineation (in sections cut in the plane of layering) at a given level. Hence the standard deviation may be regarded as an "index of lamination" (or lineation); a low value indicates a higher degree of preferential orientation than a high value. The variation of such lamination indices with height are shown in

Plate 4.5. "Structural" unmixing of larger domains of clinopyroxene within host of clinopyroxene of the same composition but differing optical orientation (Sample 1427). Field of view 2mm.
Figure 4.4. An overall decrease in the degree of lamination and lineation is indicated for all mineral phases going upward through the rhythmic unit. Figure 4.4 also demonstrates that there is no observable trend in the aspect ratio (that is, the ratio of the maximum to the minimum dimension) of grains with height in the rhythmic unit.

4.4 Olivine petrofabric

Olivine petrofabric data (determined by N.I. Christensen) are available for a similar undeformed, rhythmically layered olivine gabbro cumulate from Wadi Bani Kharus (see Chapter 8) (Figure 4.5). In sample 2429 a strong b-axis maximum, normal to the plane of compositional layering, indicates that the olivine grains are oriented with their side pinacoid (010) faces in the plane of layering. No preferential orientation of the a- or c-axes is shown.
Figure 4.2. Variation in absolute minimum, maximum, and mean values for the maximum grain size dimension in olivine, clinopyroxene and plagioclase. A or B indicates a section normal to layering/lamination, C indicates a section in the plane of layering/lamination. In the graph on the far left the means for olivine (O), clinopyroxene (C) and plagioclase (P) are compared.
Figure 4.3. Variation in absolute minimum, maximum, and mean values for the minimum grain size dimension in olivine, clinopyroxene and plagioclase. A or B indicates a section normal to layering/lamination, C indicates a section in the plane of layering/lamination. In the graph on the far left the means for olivine (O), clinopyroxene (C) and plagioclase (P) are compared.
Figure 4.4. Lamination indices (or lineation indices for sections in the plane of layering/lamination) for olivine, clinopyroxene, and plagioclase. Graph to the far right shows the variation in the aspect ratios of the mean dimensions for all minerals. Notation as for Figures 4.2 and 4.3.
Figure 4.5. Olivine petrofabric for sample 2429, a rhythmically layered olivine gabbro from Wadi Bani Kharus (see Chapter 8). The plane of compositional layering is shown by great circle. The 010 maximum is shown by an open star.
4.5 Hydraulic equivalence

Physical processes, such as crystal settling or density current deposition, have been offered as mechanisms for the formation of layering in cumulate rocks. In the light of such mechanisms it is instructive to examine the "hydraulic equivalence" (Jackson 1961) of the grains that form the Wadi Shafan rhythmic unit.

The well-known Stokes' Law relation

\[ v_0 = \frac{2 \pi r^2 g \Delta \rho}{9 \eta} \]

describes the terminal settling velocity of spherical particles of radius \( r \) in a liquid of viscosity \( \eta \) having a density contrast \( \Delta \rho \) with that liquid. This only holds at infinite particle dilution. As the volume proportion (\( \phi \)) of particles in the liquid rises, the effective settling velocity \( v_s \) decreases rapidly according to the relation (Shaw 1965):

\[ v_s = v_0 (1 - \phi)^{4.65} \]

For two grains of different mineral phases (and therefore of differing density) to occur at the same level in a layer formed by crystal settling from the same initial height in a Stokes' Law liquid their settling velocities must be the same. That is, they are hydraulically equivalent. From the two expressions above we may derive the following relation for two different mineral phases of radius \( r_1 \) and \( r_2 \), and density contrast \( \Delta \rho_1 \) and \( \Delta \rho_2 \) respectively:

\[ r_1^2 \Delta \rho_1 = r_2^2 \Delta \rho_2 \]

Using this expression, and values for mineral densities appropriate to their observed compositions (see below) from Deer et al. (1966) who give \( \rho_{ol}=3.45 \), \( \rho_{cpx}=3.35 \) and \( \rho_{plag}=2.75 \), graphs of hydraulic equivalence between clinopyroxene and olivine, plagioclase and clinopyroxene, and plagioclase and olivine have been drawn for typical gabbroic liquids of density 2.7, and granitic liquids of density 2.4 (Figure 4.7). (No correction has been made for the effect on density of thermal expansion; in all cases it will operate in the same sense and be of similar magnitudes.) Figure 4.6 also shows the
mean maximum grain dimension data from the Wadi Shafan rhythmic unit.

Little correlation is seen between mean grain sizes for clinopyroxene and olivine, or plagioclase and olivine. In situ resorption of olivine is likely to have dispersed any mutual grain size relationship, effectively moving points to the left on Figure 4.6. Plagioclase and clinopyroxene mean grain sizes are seen to vary sympathetically; however, such a covariation it is well outside the range for plausible magma densities of between 2.4-2.7; a best fit straight line gives an indicated liquid density of 1.37! (Similar results are obtained for the absolute maxima and minima of the maximum dimension, and for the minimum dimensions also.)

4.6 Phase chemistry

If a mechanical sorting process such as density current deposition or gravity settling are responsible for the formation of rhythmic layering, then an absence of chemical variation between the constituent mineral grains at different positions in the same unit would be expected; "mechanical" fractionation should only operate between grains of different size and/or density. If alternatively, the rhythmic layer formed in situ by some unspecified chemical process, one might anticipate marked variations in phase chemistry to accompany the rapid observed variations in whole rock composition.

A useful test between hypotheses would therefore be, in the case of the Wadi Shafan rhythmic layer, to determine the compositions of olivine, clinopyroxene, and plagioclase at different levels throughout the unit; microprobe analyses and calculated compositional parameters for these three phases are presented in Tables 4.1, 2 and 3 respectively. In the graphs that follow, mineral compositions and their error bars for individual data points, were evaluated as described in Appendices 1 and 2.
Figure 4.6. Plots between the mean maximum dimensions determined for samples from the rhythmic unit. Theoretical correlations for gabbroic (p=2.7) and granitic (p=2.4) liquids are shown.
4.6.1 Olivine

Figure 4.7 shows the variation of Fo content, NiO, and MnO in olivine with height in the rhythmic unit. As is typical of olivine from the Cumulate Sequence, no zoning with respect to Fo, NiO, or MnO is apparent within or between grains in any one sample. In addition, for NiO and MnO, no variation in content is observed with height, mean values remaining at 0.13 and 0.25 wt% respectively. However, for Fo content, there is a decrease with height from 85.0 to 83.5 within the rhythmic unit of interest, the value returning to 84.6 at the base of the overlying rhythmic unit.

4.6.2 Clinopyroxene

A uniform major element composition is observed, both within and between samples, with an average value of 87 for Mg
t (Figure 4.8). None of the minor oxides Cr2O3, MnO and Na2O show significant variation within or between different levels, with mean values of 0.22, 0.15 and 0.20 wt% respectively. TiO2 (average 0.3 wt%) and Al2O3 (average 2.4 wt%) however show significant zoning within grains but no overall variation with height. TiO2 shows no consistent direction of zoning, but Al2O3 is enriched in the cores of clinopyroxene grains, the most extreme example being at the top of the underlying rhythmic layer. The compositional parameter Cr* (100Cr/Cr+Fe*), usually the most sensitive indicator of fractionation in clinopyroxene (see Chapter 5), shows no variation within the rhythmic unit (average value 4.0), but does suggest some degree of reverse zoning. The underlying rhythmic unit may have overall higher Cr*. The blebs/domains of differing optical orientation (see section 4.2), and their adjacent host grains in sample 1427, show no sign of chemical contrast.

4.6.3 Plagioclase

No variation with height is seen in the mean An content of plagioclase of 86.2 (Figure 4.7). However, between grains of the same level, a tendency to reverse zoning is observed.
Figure 4.7. Compositional variation in Fo, NiO and MnO contents in olivine, and An content in plagioclase for the Wadi Shafan rhythmic unit. C=core, R=rim, H=50/50. Error bar on single determination is shown.
Figure 4.8. Compositional variation in TiO$_2$, Al$_2$O$_3$, Cr$_2$O$_3$, MnO, Na$_2$O, Mg* and Cr* contents of clinopyroxenes in the Wadi Shafan rhythmic unit. C=Core, R=rim, H=50/50, B=bleb, H=host. Error bar on single determination is shown.

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Overall then, the constituent phases of the rhythmic unit show a marked absence of chemical variation between different heights in the layer. Only Fo content of olivine shows significant chemical variation, decreasing in value upwards through the layer. As noted above, the olivine grains show an increasing degree of resorption with height in the unit. As no upward fractionation trend is shown by clinopyroxene or plagioclase (which presumably formed in equilibrium with olivine), the variation in Fo content is likely to be a result of the same process that caused the resorption of the olivine grains.

4.7 The formation of layering

A brief review of cumulus theory, and some of the problems faced by it, are presented in the following sections. Evidence from the Wadi Shafan rhythmic layer is discussed in the light of classic cumulus theory and with regard to some of the more recent hypotheses proposed to account for layering in igneous rocks.

4.7.1 Classic cumulus theory

The principles of cumulus theory were first outlined by Wager and Deer (1939), and found wide acceptance, culminating in the benchmark publication of Wager and Brown (1968). Morse (1980) presents a summary of the theory, the main features of which are shown schematically in Figure 4.9.

Nucleation and crystallization occur at the cooling walls and roof of a large magma chamber; the crystals, being denser than the magma, tend to sink towards its base. This "background" sedimentation results in "average rocks" that show good lamination but no sorting of contrasting minerals or grain sizes. Temperature differences between the roof and base of the cooling magma chamber induce large scale convection. Sudden changes in the convective regime or slurries of unconsolidated masses of crystals from the magma chamber walls produce size or mineral graded rhythmic layers. The piles of crystals on the magma chamber floor are cemented by
crystallization of "pore" liquid; if equilibrium is maintained with the
overlying magma body accumulates result, if the pores act as closed systems
orthocumulates are produced. Progressive fractionation of the magma
results in cryptic variation; input and mixing of primitive magma from below
gives rise to cyclic units often defined by phase layering. The overall
"allochthonous" nature (in the manner of pelagic or turbiditic sedimentary
rocks) is compounded by the occurrence of features analogous to
sedimentary structures such as trough-bedding and load casts. The vast
majority of layered igneous rocks were considered to have formed by such
processes.

However, a small percentage were believed to be "autochthonous" in
origin (as are evaporites and some limestones in the sedimentary record),
having been formed by some process of in situ crystallization. The
delicate, bladed, branching structures shown by crescumulates/harrisites or
the macrospherulites from the Rhum Intrusion (Wager and Brown 1951,
Wadsworth 1961, Donaldson 1974) were considered too fragile and complex to
have been formed by crystal settling, as was the fine, regular, "inch scale"
layering from the Stillwater Intrusion (Hess 1960) or the colloform
congelation cumulates of the steeply inclined Marginal Border Group of the
Skaergaard Intrusion (Wager and Brown 1968). Rather than crystal settling,
in situ processes such as bottom crystallization (Jackson 1961) and
constitutional supercooling (Lofgren and Donaldson 1975) were regarded as
likely mechanisms.
Figure 4.9. The essential features of classic cumulus theory.
4.7.2 Some problems with cumulus theory

Stimulating reviews of the problems facing cumulus theory are provided by Campbell (1978, and worthy of much wider readership) and McBirney and Noyes (1979).

Bottinga and Weill (1970) first drew attention to the problem of plagioclase density; for the calculated magma composition of the Skaergaard Intrusion, the plagioclase would always be less dense than the coexisting liquid. Murase and McBirney (1973) confirmed these calculations by experiment (Figure 4.10).

Figure 4.10. Density relations between minerals and successive liquids for the Skaergaard Intrusion (from McBirney and Noyes 1979).

The realization in recent years (Shaw et al. 1968, Murase and McBirney 1973, Sparks et al. 1977) that silicate liquids are not necessarily Newtonian (i.e. they can sustain a shear stress) poses further problems for
a crystal settling mechanism. In as much as they possess a finite yield strength that must be exceeded before an applied stress produces permanent strain, a crystal of a given density contrast with a particular silicate liquid, must attain a certain size, and so exceed the yield strength, before it may sink (or rise). Figure 4.11, from McBirney and Noyes (1979), shows the relations between size and density contrast of crystals in liquids of differing yield strengths. Typical yield strengths of basaltic magmas close to their liquidus are greater than 600 dynes cm\(^{-2}\) and become much greater at lower temperatures (Shaw et al. 1968, Murase and McBirney 1973); olivine and pyroxene must achieve grain sizes of about 3cm before they can commence to settle - hardly a grain dimension typical of layered igneous bodies.

![Figure 4.11. Relations between radius and density contrast of minerals in liquids of differing yield strengths (diagonal lines). A crystal can only begin to settle (or rise) if it has a density contrast and size that place it in the region of appropriate yield strength (from McBirney and Noyes 1979).](image-url)
That the near vertical colloform layering structures of the Marginal Border Group of the Skaergaard Intrusion formed in situ on the walls of the pluton is not in dispute. However, rhythmic layering, identical to the flat lying Layered Series of Skaergaard, and often showing cross or trough bedding, has been reported at steep or vertical inclinations by Philpotts (1968) from the Mount Johnson Intrusion, by McCall and Peers (1971) from the Binneringie Dyke, and by Campbell et al. (1970) from the Jimberlana Intrusion. Campbell (1978) notes that there is no textural, structural or tectonic evidence of deformation in the Jimberlana Intrusion and that a palaeomagnetic study shows that the Lower Layered Series (the gently dipping cumulates) and the Marginal Layered Series (the steeply dipping cumulates) have the same magnetic pole position and therefore concludes that the steeply inclined layering is primary.

With respect to the smaller scale aspects of layering, reverse grading (i.e. rhythmic layering with pyroxene rather than plagioclase rich tops - hydraulically upsidedown) has been found in Stillwater (Hess 1960), Kiglapait (Morse 1969) and Jimberlana (Campbell et al. 1970), and a lack of hydraulic equilibrium has been noted between olivine and chromite grains in layers from Stillwater (Jackson 1961). The even distribution of different mineral species between the walls and interior of the Skaergaard Intrusion, the lack of variation in thickness between the "stoss" and "lee" sides of obstructing blocks, and the absence of braided or overbank deposits from the trough bands lead McBirney and Noyes (1979) to exclude density currents as means of forming rhythmic layers.

Campbell (1978) considers that heterogeneous and self-nucleation are the dominant nucleation mechanisms during the formation of cumulates. (In homogeneous nucleation, nuclei form at random centres throughout the liquid. In heterogeneous or self-nucleation, nuclei develop against the margins or pre-existing crystals in the intrusion.) The activation energy, and consequently the amount of undercooling, required for heterogeneous nucleation is very much less than for homogeneous nucleation. Cooling and crystallization rates in the large magma chambers that form layered
intrusions are slow, and it seems unlikely that the magma would attain the high degree of supercooling necessary for widespread homogeneous nucleation. Campbell (1978) concludes that heterogeneous nucleation at the sides and bottom of the chamber is most likely.

The oxygen isotope ratios of basalt xenoliths and their Layered Series host rocks from the Skaergaard Intrusion (Taylor and Forester 1979) provide persuasive evidence for in situ crystallization of rhythmically layered cumulates (Figure 4.12). The basalt xenoliths (and the basalt country rocks) are strongly $^{18}O$-depleted (values of $\delta^{18}O$ down to $-4.0$), attributed to exchange by a meteoric-hydrothermal system that was established within the host rocks of the intrusion. The Layered Series cumulate in contact with the basalt xenolith is also strongly $^{18}O$-depleted in $^{18}O$ ($\delta^{18}O_{\text{plag}} 3.2$ and $\delta^{18}O_{\text{cpx}} 2.4$), and is much lower in $^{18}O$ than the more distant Layered Series samples (two metres from the contact the $\delta^{18}O_{\text{plag}}$ has the near "normal" value of $+5.9$). The $\Delta_{\text{plag-cpx}}$ value of 0.8 indicates isotopic equilibrium at c.$950^0C$, and $^{18}O$ depletion at subsolidus conditions is therefore unlikely. Taylor and Forester (1979) conclude that the low $\delta^{18}O$ value of both the plagioclase and the clinopyroxene is the result of growth in situ from an $^{18}O$ depleted silicate melt at the contact of the xenolith.

The density relations between plagioclase and liquid, and the observed yield strengths of magmas would seem to preclude "passive" crystal settling as a viable mechanism for the formation of cumulates from a gabbroic melt. (This is not to dismiss the possibility of crystal settling occurring in ultramafic liquids of lower viscosity and yield strength.) "Active" deposition by density currents in gabbroic liquids must remain an open question; understanding of the physical properties of crystals in suspension in magmas remains very limited. The occurrence of typical cumulate textures and layering at steep primary inclinations must indicate that at least some layering forms as a result of physico-chemical and not
mechanical processes. The likelihood that heterogeneous rather than homogeneous nucleation is the dominant crystallization mechanism in large, slowly cooled magma chambers, favours in situ crystallization at the margins rather than within the body of the intrusion. Stable isotopic evidence from the Skaergaard Intrusion lends support to this assertion.

4.7.3 The Wadi Shafan rhythmic layer

The rationale behind the investigation of the Wadi Shafan rhythmic layer (sections 4.2-4.6) may be summarized as follows: if a mechanical process was responsible for the layer formation then a regular physical fractionation of the layer components might be expected; if an in situ crystallization process was the mechanism of layer formation then chemical fractionation on the scale of the layer would be anticipated.

At face value the evidence points toward a mechanical origin to the rhythmic layer. The hydraulic equivalence observed between plagioclase and clinopyroxene indicates a mechanical fractionation with respect to grain size and density contrast; that the inferred liquid density of 1.37 is
completely unrealistic suggests that passive settling in a Newtonian liquid may be rejected. The ratio contact near the base of the rhythmic layer, and the suggestion of reverse grading below this contact, might point towards grain flow (Middleton and Hampton 1976) at the base or in front of a density current. The decrease in intensity of lamination and lineation observed with height may be a result of waning current action, or more moderate packing of grains occurring at higher levels in the layer (this being preserved by intercumulus "cementation" prior to the deposition of the overlying rhythmic unit). In contrast with the rapid chemical variation reported by McBirney and Noyes (1979) from a rhythmic layer from Skaergaard, constituent phases of the Shafan layer are constant in composition through the height of the unit. (The slight variation in olivine composition is attributed to its reaction with intercumulus liquid.) A mechanism involving in situ crystallization for the formation of the Wadi Shafan rhythmic layer can only be rejected if the operation of processes such as infiltration metasomatism (Korzhinsky 1965, Irvine 1980a), the upward migration of intercumulus melt (by input from below or by filter pressing) and the subsequent modification of the cumulate through which these melts pass, can be realistically excluded. Such a process could result in the complete re-equilibration of the affected cumulate, obliterating all pre-existing chemical variations that might have been produced by an in situ crystallization process. Infiltration metasomatism cannot be excluded for the Wadi Shafan layer, though no definite magmatic replacement features have been observed.

4.7.4 Mechanisms of rhythmic layer formation

Whether a mechanical or a physico-chemical process was responsible for the formation of the Wadi Shafan layer must, on the evidence presently available, remain a matter of speculation. It is instructive however, to examine some of the models (both mechanical and physico-chemical) which have been proposed to account for rhythmic layering. A better understanding of the processes involved should allow experiments or investigations to be designed which would enable more rigorous testing of the respective hypotheses.

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Results of experimental modelling of magmatic density currents have been presented by Irvine (1980b). Such studies are fraught with scaling problems, such as satisfactory reproduction of the fluid dynamics of silicate magmas at 1000°C in kilometre-size magma chambers with glycerin solutions at room temperature in metre-size flume tanks. Nonetheless it has been demonstrated by Irvine (1980b) that simulated viscous flows are capable of reproducing the size and modal grading (including reverse grading) observed in layered intrusions (Figure 4.13). The plagioclase buoyancy problem is overcome by transport in crystal suspensions charged with mafic minerals. Irvine (1980b) argues that if the magma is flowing and undergoing shearing, yield strength effects can be overcome. Ironically it is the yield strength of the magma (which prevents passive settling) that retains the less dense grains of plagioclase on the floor of the magma chamber after deposition by active density flows.

Few would refute that the spectacular size-graded layers of the Duke Island Intrusion (Irvine 1974), containing xenoliths of quartz which should have floated in the parent magma, were deposited by magmatic density currents; the parent ultramafic liquids are less viscous and have lower yield strengths than gabbroic magmas. A major question is whether such mechanical processes can also operate in less primitive gabbroic liquids.

Crystallization at the base, rather than at the walls or roof, of the magma chamber was forwarded by Jackson (1961). Whilst cooling rates may be higher at the roof, Jackson (1961) pointed out that due to the greater positive pressure dependence of liquidus temperatures compared to the adiabatic gradient for a convecting chamber, a homogeneous magma should crystallize more rapidly at its base than under its roof (Figure 4.14); an effect enhanced by volatile concentration near the top of the chamber (Elsdon 1970).
Figure 4.13. Density current models for the formation of graded layers. In (A) the suspension contains only mafic minerals and size grading results. In (B) the suspension consists of plagioclase (the blocky mineral) and the same mafic minerals, and because the feldspar has near neutral buoyancy, modal grading is dominant (from Irvine 1980b).
Figure 4.14. Successive position in time of the adiabatic gradient are shown in a cooling magma chamber. Between $t_1$ and $t_2$, the liquidus gradient is intersected and crystallization at the base of the magma chamber is initiated (after Jackson 1961).
The formation of rhythmic layering by an in situ oscillatory process of nucleation and crystal growth within a cooling boundary layer was proposed by McBirney and Noyes (1979). Consider a cooling magma chamber; at any particular time, the temperature increases inward from its boundaries, and at any given position the temperature is slowly decreasing. At the margin (wall, roof or floor) of the magma chamber the temperature profile will decay as a linear function of time according to the relation

\[ T(x,t) = a + bx - ct \]

where \( T \) is temperature, \( t \) is time, \( x \) is the distance from the margin, and \( a, b, \) and \( c \) are positive constants. Such a linear decay of the temperature profile with time is illustrated in Figure 4.15a.

If crystallization is initiated at the cooling margin of the magma chamber, the adjacent magma (which is stagnant within the boundary layer) will become depleted by diffusion of whatever components are being incorporated in the growing crystals. The rate at which the "diffusion front" advances through the magma is a function of the square root of time as given by

\[ C(x,t) = C_0 \text{erf} \left[ \frac{x}{(2Dt)^{0.5}} \right] \]

where \( C \) is the concentration of any chemical species, \( C_0 \) the initial concentration, \( D \) is the diffusion coefficient and \( \text{erf} \) indicates the error function. The form of the diffusion front with time is shown in Figure 4.15b. At small values of \( D \) the diffusion front spreads more slowly, at larger values more rapidly.

If crystallization commences at a particular level of supersaturation \( C_s \) and undercooling \( T_u \), it may be seen (Figure 4.15c) that initially the locus of \( C_s \) on the diffusion front may advance more rapidly than the locus of \( T_u \) on the temperature profile. However, because of the contrasting linear time dependence of the temperature profile decay and the square root dependence of the concentration profile decay, the levels of supersaturation and undercooling will later again coincide, initiating a second burst of crystallization. The process repeats again and again, and a "chemical" oscillator capable of producing rhythmic layering has been
established. The ubiquitous feature of igneous lamination is attributed by McBirney and Noyes (1979) to crystal growth along planes of equal chemical potential parallel to the cooling front.

Using this crude model, and likely values of constants appropriate for the Skaergaard Intrusion, McBirney and Noyes (1979) estimate that the dimensions of rhythmic layering formed by such a process would be between a few centimetres to nearly a metre. More rigorous quantitative description of the processes that operate during nucleation, crystallization and layer formation must wait until the physical parameters and complex mutual relations of temperature, diffusivities and rates of crystal growth of different chemical species in silicate liquids are better known.
Figure 4.15. (a) Decay in temperature profile with time at the cooling margin of a magma chamber. (b) Decay in composition profile following initiation of crystallization at the cooling margin. (c) Relative rates of temperature and composition profile decay.

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Table 4.3. Plagioclase microprobe analyses for Wadi Shafan rhythmic layer. C=core, R=rim, H=50/50, 0=no description.
Chapter 5

Wadi Bani Kharus Cumulate Sequence

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Chapter 5
Wadi Bani Kharus Cumulate Sequence

5.1 Introduction

In the preceding chapter it was demonstrated that variation in phase chemistry could not be detected on the scale of individual rhythmic layers within the Cumulate Sequence. In an effort to establish at what scale geochemical variation does occur, the mineral chemistry of a 2000+m thickness of the Cumulate Sequence and the High Level Intrusives in Wadi Bani Kharus has been studied (Figure 5.1). In particular, the lowermost 600m of rhythmically layered gabbroic cumulates that form the base of the Cumulate Sequence were sampled at a 20m spacing. Alternative Rayleigh fractionation magma chamber models which might account for the observed geochemical variations are investigated.

5.2 Petrography

The mineralogy of the lower, closely sampled 600m section of the Cumulate Sequence in Wadi Bani Kharus is straightforward; plagioclase and clinopyroxene are always present, with or without olivine and very minor Fe-Ti oxides. Typical modes for this section are estimated as plagioclase 55%, clinopyroxene 35% and olivine 10%. Plagioclase usually shows clouding, olivine is serpentinised to varying degrees, and clinopyroxene may show retrograde breakdown to actinolite.
Figure 5.1. Location of the Wadi Bani Kharus Cumulate Sequence section. Column shows position of the closely sampled section at the base of the Cumulate Sequence.
The primary igneous textural relationships are complicated by subsolidus deformation (see Chapter 2). The degree of recrystallization is seen to decrease with height (none has been identified above 600m), though not uniformly. Original igneous features may be found interspersed between recrystallized zones; only meso- or adcumulate textures have been identified (exemplified by samples 2395, 2400, 2403, 2411, and 2422). Such specimens are usually coarser grained (maximum grain size 3-6 mm) and show a well developed lamination of prismatic grains (though samples 2412-2416 provide finer grained exceptions). The recrystallized equivalents are finer grained (less than 3 mm, often 0.25-0.5 mm mean grain size), and show the development of a foliation defined by lenticular rather than prismatic shape fabrics in the same plane, or at a small angle to, the original igneous lamination. The petrographic features of the Cumulate Sequence - High Level Intrusives - Sheeted Dyke Complex transition in Wadi Bani Kharus have been documented in Chapter 2. Orthopyroxene is present in the higher levels of the Cumulate Sequence as a cumulus phase. Magnetite, primary amphibole and orthopyroxene occur with clinopyroxene and plagioclase in the High Level Intrusives; quartz is absent.

5.3 Phase chemistry

5.3.1 A note on the identification of geochemical cycles

The validity of graphs illustrating the cyclic variation of phase chemistry of cumulates with height depends on the size of the sampling interval. Ideally there should be a very small interval between sample points. However, time and analytical facilities restrict the number of samples that can be investigated, so that interpolation and interpretation between more widely spaced samples becomes necessary. Interpretation is required as well as interpolation as it would be meaningless simply to "join up the dots". Consider Figure 5.2 which shows the variation with height of a compatible element in a given phase within a succession of layered rocks; in case (a), where the sampling interval is generously close, and no problem is encountered in drawing the correct solution; however, with a wider
interval to the data points (case (b)) the solution that merely "joins up the dots" is in error as it gives the impression of a positive geochemical gradient where none exists. But in case (c) we are justified in drawing a positive geochemical trend as intermediate data points are present. In case (d) we are faced with ambiguity; equal weight may be given to both solutions. One must guard against "over-interpretation"; little justification can be offered for the solution shown in case (e).

![Diagram](https://example.com/diagram.png)

Figure 5.2. Diagram to illustrate the problems encountered when identifying geochemical cycles. For discussion see text.

The procedure adopted here is that when dealing with an element that is expected to behave compatibly, negative gradient fractionation trends are drawn between points where possible, and horizontal mixing lines are inferred to join displaced negative gradients, except where data points of intermediate composition are present, when positive gradients are justified. The reverse is followed for incompatible elements. This convention is only disregarded when evidence to the contrary is offered by the variations shown by other elements in the same, or coexisting phases. In this way the interpretation of cyclic geochemical patterns that results is based on the sum total of the data available, and not on one particular element in a single phase viewed in isolation.

5.3.2 Observed variations in phase chemistry

Microprobe analyses of olivine, plagioclase, pyroxene and amphibole for samples collected throughout the Wadi Bani Kharus succession are given in Tables 5.2, 5.3, 5.4 and 5.5 respectively. In the graphs that follow, mineral
compositions and their error bars for individual data points were evaluated as described in Appendices 1 and 2.

5.3.3 Variation over 2000+m section

In order to set the following detailed account of the closely sampled section in the context of the entire 2000+m succession of Cumulate Sequence and High Level Intrusives in Wadi Bani Kharus, Figures 5.3, 5.4, 5.5 and 5.6 illustrate the overall variation in phase chemistry. Here it is sufficient to note that for all elements and parameters, no overall trend of fractionation is seen, and the range in abundances shown is similar, irrespective of the height above the Petrological Moho.

5.3.4 Variation over 600m section

Plots of the variation in phase chemistry with height over the 600m closely sampled section are shown in Figures 5.7, 5.8, 5.9 and 5.10. In addition, to facilitate overall comparison, the data for each parameter have been averaged and displayed adjacent to the means of other parameters in Figures 5.11 and 5.12.
Figure 5.3. Variation over 2000m section of Fo, NiO and MnO content of olivine.
Figure 5.4. Variation over 2000m section of An content of plagioclase, and MnO content of clinopyroxene. C=core, H=50/50, R=rim, I=small interstitial, O=no description.
Figure 5.5. Variation over 2000m section of Mg*, Cr* and Cr2O3 content of clinopyroxene. C=core, H=50/50, R=rim, I=small interstitial, O=no description.
Figure 5.6. Variation over 2000m section of Na2O and TiO2 content of clinopyroxene. C=core, H=50/50, R=rim, I=small interstitial, O=no description.
Figure 5.7. Variation over 600m section of Fo, NiO and MnO content of olivine.

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Figure 5.8. Variation over 600m section of An content of plagioclase, and MnO content of clinopyroxene. C=core, H=50/50, R=rim, I=small interstitial, O=no description.
Figure 5.9. Variation over 600m section of Mg*, Cr* and Cr2O3 content of clinopyroxene. C=core, H=50/50, R=rim, I=small interstitial, O=no description.
Figure 5.10. Variation over 600m section of Na2O and TiO2 content of clinopyroxene. C=core, H=50/50, R=rim, I=small interstitial, 0=no description.
When considering the variation in mean compositions shown in Figures 5.11 and 5.12, the degree of zoning shown by minerals must be borne in mind. Table 5.1 summarizes the nature of the zoning present for all minerals, based on the data shown in Figures 5.7 to 5.10. A number of points emerge:

(i) In olivine any zoning with respect to Fo, NiO or MnO content is at or below the limit of detection.

(ii) Plagioclase commonly displays reverse zoning in terms of An content.

(iii) Zoning in clinopyroxene with respect to compatible elements (or ratios) (i.e. Mg*, Cr*, or Cr2O3) is not widely developed. In the lower quarter of the section a tendency to reverse zoning with respect to Mg* is indicated.

(iv) If zoning is present with respect to MnO in clinopyroxene, it is at or below the limit of detection.

(v) Those elements considered to be moderately incompatible in clinopyroxene (TiO2 and Na2O) are commonly significantly zoned. TiO2, where a distinction may be drawn, shows reverse zoning; Na2O appears to be reversely zoned in the lower half of the section, but normally zoned in the upper half.

(vi) Only in one sample (2416) are both reverse and normal zoning shown for different elements (in this case normal zoning for Na2O in clinopyroxene, and reverse zoning for TiO2 in clinopyroxene and An in plagioclase). For all other samples zoning, where present and detectable, is coherent (i.e. is in the same sense) for all phases.

With these points in mind the section has been divided into a number of cycles based on the criteria outlined in section 5.3.1; three major cycles I, II and III, based on the Cr2O3 content of clinopyroxene, and a number of smaller divisions (a, b, etc.) based on the overall geochemical coherence, have been identified (Figure 5.11 and 5.12). The data for the divisions Ia, IIb, and IIIc is insufficient to allow unambiguous interpretation of the geochemical trend.
The dominant feature, as would be anticipated, are "normal" negative geochemical gradients with height for compatible elements (or ratios) (i.e. Fo and NiO in olivine, Mg*, Cr*, and Cr2O3 in clinopyroxene, and An in plagioclase), and positive gradient trends for incompatible elements (i.e. MnO in olivine, Na2O, TiO2 and MnO in clinopyroxene). No overall trend to more evolved compositions with height is observed. Although the overall geochemical pattern is dominated by "normal" trends, "reversed" trends where compatible elements show sustained enrichment with height, and incompatible elements show a progressive depletion, occur in cycles IIe and IIf. Also, apparently incoherent geochemical behaviour ("decoupling") is shown in cycles IId and IIf for Na2O, in cycles Ib and IIf for TiO2, and in cycle IIIb by Cr2O3.
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Table 5.1. Characteristic zoning shown by olivine, plagioclase and clinopyroxene in Wadi Bani Kharus. Z=zoned, N=normal zoning, R=reverse zoning, ?=near limit of detection, D>1=compatible.
Figure 5.11. Mean contents of NiO, Fo and MnO in olivine, and An in plagioclase for Wadi Bani Kharus 600m basal section. Cyclic units identified on basis of phase chemistry are shown in the column on the right. Open circles are inferred compositions for olivine-free samples.
Figure 5.12. Mean contents of Cr*, Mg*, Cr2O3, Na2O, MnO and TiO2 in clinopyroxene for Wadi Bani Kharus 600m basal section. Cyclic units identified on basis of phase chemistry are shown in the column on the right. -d-d-d- = element showing decoupling. (N) or (T) = extreme values of Na2O or TiO2 excluded from calculation of respective mean values.
5.4 Magma chamber modelling

5.4.1 Open system fractionation

A plausible magma chamber model is illustrated in Figure 5.13. A magma chamber of mass $M_o$, and composition $C_o$, crystallizes a mass fraction of cumulate, $X$. For perfect Rayleigh fractionation, the composition of the remaining liquid will be given by

$$C_{l1} = C_o (1-X)^D^{-1},$$

where $D$ is the bulk distribution coefficient for the phase assemblage crystallizing. The magma chamber then erupts a mass fraction $Y$, the current mass of the chamber becoming $M_o(1-X-Y)$. An input of magma of composition $C_Z$ and mass fraction $Z$ (expressed as a fraction of the original chamber mass $M_o$), then enters the magma chamber and mixes completely with the fractionated liquid $C_{l1}$. The new mass of the magma chamber will be given by

$$M_1 = M_o (1-X-Y) + M_o Z$$

and the mixed composition by

$$C_{m1} = \frac{C_{l1}M_o(1-X-Y) + C_ZM_oZ}{M_1}.$$  

The magma chamber model then continues with the mixed magma composition $C_{m1}$ becoming $C_o$ of the second cycle.

Providing $X$, $Y$, $Z$, $C_Z$ and $D$ remain constant, and the system is closed to any other outside influences (i.e., no assimilation of wall rocks occurs) then in subsequent cycles the mass of the magma chamber will be given by

$$M_n = M_{n-1}(1-X-Y) + M_o Z,$$

and the evolved liquid composition by

$$C_{l_n} = C_{m_{n-1}} (1-X)^D^{-1},$$

and the mixed liquid composition by

$$C_{m_n} = \frac{C_{l_n}M_{n-1}(1-X-Y) + C_ZM_oZ}{M_n},$$

where $n$ is the number of cycles.

The variation of composition and mass with time in a magma chamber undergoing successive cycles of fractionation, eruption, input and mixing is shown in Figure 5.14. All elements, irrespective of compatibility or incompatibility, are seen to achieve a steady state composition; the smaller
the distribution coefficient, the longer it takes for a steady state composition to be reached. A steady state mass to the magma chamber is attained rapidly.

O'Hara (1977) and O'Hara and Matthews (1981) have derived expressions for the steady state compositions and mass of a such a magma chamber undergoing "open system" fractionation with constant $X$, $Y$, $Z$, $C_2 (=C_0)$ and $D$. The steady state mass is given by
\[ ssM = Z/(X+Y), \]

and the steady state evolved liquid composition by

\[ ssCL = \frac{(X+Y)(1-X)(D-1)}{1-(1-X-Y)(1-X)} \]

From equation 5.1 it follows that the steady state mixed composition \( ssCm \) is related to \( ssCL \) by the expression

\[ ssCm = \frac{ssCL}{(1-X)} \]

Note that the steady state compositions achieved are dependent on \( Cz \), the magma composition input, but independent of \( Z \), the mass fraction input.

At any time in the cycle the composition of the coexisting crystal extract \( Cx \), is related to the prevailing liquid composition \( CL \), by

\[ Cx = D \cdot CL \]

5.2
Figure 5.14. Variation in time of mass and liquid compositions of a magma chamber undergoing open system fractionation, for incompatible (D=0.1), just incompatible (D=0.5), just compatible (D=1.5) and compatible (D=5) elements.

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5.4.2 Magma chamber height

Consider a magma chamber, tabular in form, of height $l$, width $a$, length $b$, and composition $C_0$ (Figure 5.15a). For perfect Rayleigh fractionation, after crystallizing a mass fraction $X$, giving a thickness of cumulate $c$, the liquid composition will evolve to

$$C_l = C_0 (1-X)^{D-1}. \quad 5.3.$$  

In terms of the dimensions of the magma chamber, the mass fraction $X$ that has fractionated is

$$X = \frac{c \cdot a \cdot b}{l \cdot a \cdot b} = \frac{c}{l}.$$  

Combining this with equation 5.3 we may write

$$C_{xl} = C_{xo} (1-c/l)^{D-1} \quad 5.4$$

where $C_{xo}$ and $C_{xl}$ are the compositions of the first and last formed cumulates in a given cycle. Rearranging gives

$$l = \frac{c}{(1-(C_{xl}/C_{xo})^{1/(D-1)})} \quad 5.5$$

which in principle allows calculation of the maximum magma chamber height, as $c$, $C_{xo}$, and $C_{xl}$ are known, provided the liquid remained homogeneous in composition at all stages of the fractionation cycle.

5.4.3 Variation in cumulate composition with height

The equations outlined in section 5.4.1 allow the variation with time of liquid compositions. The same expressions, by virtue of equation 5.2, allow investigation of the manner in which the composition of cumulates collecting at the base of a magma chamber vary with height. The time axis in Figure 5.14 becomes a dimensionless scale of height (the ratio $c/l$) from equation 5.4. Figure 5.16 shows how the coexisting cumulate and liquid compositions would vary with height for compatible and incompatible elements in a sequence formed by open system fractionation.

5.4.4 Some other magma chamber shapes

Equation 5.5 allows calculation of the height of the column of liquid from which a given thickness of cumulate was deposited during one cycle of fractionation in a tabular shaped magma chamber. Some other geologically more realistic shapes may also be considered.
Figure 5.15. Possible magma chamber shapes. (a) Tabular, (b) Triangular, and (c) Semi-circular. For expressions relating the height of the liquid column (l) to the depth of the cumulate formed (c) see text.

For a magma chamber triangular in cross section (Figure 5.15b), of height L, length b, with walls that are inclined inward at an angle of θ, in which cumulate of thickness c forms parallel to those walls, the mass fraction $X$ that has fractionated in any one cycle will be given by the ratio of the volume of the amount of cumulate formed ($V_1$) to the volume of the
Figure 5.16. Variation in extract composition with height, for incompatible (D=0.1) and compatible (D=3) elements, in a cumulate pile formed within an open system magma chamber. The contemporaneous coexisting liquid compositions are also shown.

original body of liquid \( (V_2) \). Now,

\[
V_1 = \frac{[l^2/\tan \theta - (l/\tan \theta - c/\sin \theta)(l - c/\cos \theta)]}{b}.
\]

and

\[
V_2 = \frac{[l^2/\tan \theta]}{b}.
\]

So

\[
X = \frac{V_1}{V_2} = \frac{1 - (l/\tan \theta - c/\sin \theta)(l - c/\cos \theta)}{[l^2/\tan \theta]}
\]

\[
= \frac{(2c/\cos \theta l) - (c^2/l^2 \cos^2 \theta)}{5.8}
\]

If we let

\[
R = c/\cos \theta l
\]

then

\[
X = 2R - R^2
\]

or

\[
R^2 = 2R + X = 0.
\]

From equation 5.3 we know that

\[
X = 1 - \left(\frac{C_x}{C_{x0}}\right)^{1/D - 1}
\]

so we solve this quadratic expression for \( R \), and rejecting the root that gives an impossible solution, substitute in equation 5.9 to gain a value for \( L \), the height of the magma chamber.
For the case of a magma chamber semi-circular in cross section (Figure 5.15c) the volumes $V_1$ and $V_2$ will be given by

$$V_1 = \frac{1}{2} \pi l^2 - \frac{1}{2} \pi c^2$$
$$V_2 = \frac{1}{2} \pi l^2$$

and

$$X = \frac{V_1}{V_2} = \frac{\left( \pi c l - \frac{1}{2} \pi c^2 \right)}{\frac{1}{2} \pi l^2} = 2\left( \frac{c}{l} \right)^2 - \left( \frac{c}{l} \right)^2. \quad 5.10$$

If we let $R = c/l$ then we may write

$$R^2 - 2R + X = 0.$$ 

Again we solve the quadratic for $R$ and reject that root that gives an impossible solution, allowing determination of $l$, the magma chamber thickness.

5.4.5 Open system cumulate compositional profiles: constant parameters

It is valuable to compare and contrast the styles of compositional profile developed in cumulates produced during open system fractionation, under differing absolute and relative values of the parameters $X$, $Y$, $Z$, $C_2$ and $D$. Attention will be confined to values of the parameters that are considered realistic in the light of field evidence from the Oman Ophiolite.

It is noted that:

(i) The presence of a Sheeted Dyke Complex indicates that eruption was episodic. It is inferred that $Y$, the mass fraction erupted, was zero for long periods in terms of the magma chamber input-fractionation cycle.

(ii) No marginal border facies or cross cutting relations have been observed within the Cumulate Sequence. From this it is concluded that the magma chamber was dynamic (i.e. it crystallized as fast as it grew) and long-lived, so $X$, the mass fraction crystallizing, and $Z$, the mass fraction input, were of similar but unknown magnitudes averaged over long periods. In the examples that follow, the original mass of the magma chamber $M_0$, has been set to unity.

5.4.5.1 $X=Z=\text{large, } Y=0.$

The cumulate compositional profile developed under this condition is shown in Figure 5.17. Large amplitude compositional variations are
Figure 5.17. Cumulate compositional profile for condition $X=Z=0.5$ (large) and $Y=0$, for incompatible ($D=0.1$), just incompatible ($D=0.5$), just compatible ($D=1.5$), and compatible ($D=5$) elements.

Characteristic; highly compatible elements achieve steady state compositions rapidly.

5.4.5.2 $X=Z=$small, $Y=0$.

Figure 5.18 shows the cumulate compositional profile for these parameters. Small amplitude compositional variations contrast with the preceding case, and more cycles are required to achieve steady state compositions for all values of $D$.

5.4.5.3 $Z>X$, $Y=0$.

In this condition of magma chamber growth, the gross geochemical trends are similar (Figure 5.19a) to the case where $X=Z$ and $Y=0$, for all values of $D$; namely progressive depletion and enrichment for compatible and incompatible elements respectively. As the ratio $X/Z$ is decreased, the number of cycles required to achieve a steady state composition increase for a given value of $D$ (Figure 5.19b); for very small values of $X/Z$ a slight overall reverse trend is observed for a compatible element (case 3, Figure
5.19b) before a steady state composition is reached.

5.4.5.4 \( Z < X, Y = 0 \).

Under these conditions the mass of the magma chamber decreases. An interesting result is that for all values of \( D \) (Figure 5.20a) the gross compositional trend of the cumulate profile shows a period of enrichment for a compatible element, or a period of depletion for an incompatible element, before the steady state compositions are attained. This effect is most marked for large values of \( X/Z \) (Figure 5.20b), at a given value of \( D \).

5.4.6 Open system cumulate compositional profiles: changing parameters

It was noted in section 5.3.3 that, over the 2000+m section of the Wadi Bani Kharus Cumulate Sequence, no overall trend of fractionation is seen for any element or compositional parameter and that the range of abundances shown is similar at all levels irrespective of the height above the Petrological Moho. Such features are typical of a magma chamber that has achieved a compositional steady state, at least with respect to its
Figure 5.19. Cumulate compositional profile for Z>X and Y=0 for (a) X=0.01, Y=0, and Z=0.1 for compatible (D=10) to just incompatible (D=0.5) elements, and (b) Y=0, Z=0.1, and compatible element (D=3) for values of X=0.1, 0.01, and 0.005.

major element composition. One feature observed in the cumulate compositional profile from Wadi Bani Kharus which marked a perturbation from steady state conditions, was the presence of "reversed" cycles; that is, the sustained overall enrichment in a compatible element or overall depletion in an incompatible element with height. It was noted above that in the case were Z<X and Y=0 such trends are observed for all values of D before steady state compositions are achieved. However this feature should be regarded more as an artifact of a mathematical model rather than a reasonable petrological process. (The mass fraction Z is always expressed as a fraction of the original mass of the magma chamber M₀, rather than the current mass Mₙ (as for X and Y). In extreme cases of magma chamber contraction this is unrealistic.)

A number of examples for compatible elements are now examined in which one or more of the open system parameters undergo a sudden change, with the production of "reversed" trends.
5.4.6.1 Changing Y - eruption/isolation of magma.

It has been argued that the mass fraction $Y$ of magma erupted is likely to be zero for extended periods. An eruptive event, or some other mechanism that isolates a proportion of the magma from the system of interest (a change in convective regime perhaps), would allow a sudden increase of $Y$ which might be sustained for a few cycles.

Figure 5.21 demonstrates that such a sudden increase in $Y$ could produce an "reversed" trend; two such events have been modelled for two differing values of $Y$ and $n$. A return to condition of no eruption allows resumption of "normal" trends and a progressive return to the former steady state compositions.

5.4.6.2 Changing $C_Z$ - more primitive magma input.

A "reversed" trend is also produced if the composition of the magma input $C_Z$ is suddenly changed. This might be effected by changing the degree of fractionation or the path of fractionation that a primary magma
had undergone within the mantle before entering the crustal magma chamber. Figure 5.22 illustrates two such events; a new (more primitive) steady state composition would be achieved if the new value of $C_Z$ was sustained; in the examples shown a return to the original input composition after a few cycles leads to a return to the original steady state composition.

5.4.6.3 Changing $D$ – a compatible element becomes incompatible

If the phase assemblage crystallizing was to change suddenly by virtue of the appearance of a new phase (i.e. a new phase volume was encountered by the locus of liquid compositions) or by the disappearance of an existing phase (i.e. by complete reaction with the liquid), then the bulk distribution coefficient might change for example, so that an originally compatible element became incompatible. Two occurrences of such a situation are portrayed in Figure 5.23; a "reversed" trend is observed until the original parameters are re-established. Of course, such a change in the phase assemblage crystallizing should not go unnoticed in the developing cumulate pile.

5.4.6.4 Changing $Z$ and $C_Z$ – mixing in of overlying, evolved magma.

Previously the mass fraction input has always been considered as entering the magma chamber from below. However, one might consider the situation in which the magma chamber is compositionally stratified, as in Figure 5.24a, and the mass fraction input was supplied from above. A simple case might have a more evolved liquid overlying a lower, more primitive liquid. Both systems are closed to each other. The lower primitive liquid behaves as a separate fractionating open system until some point at which both systems mix and become a single open system (Figure 5.24b). In terms of the present model this would represent an almost instantaneous change in $Z$ (and $C_Z$), followed by a rapid return to the pre-existing parameters. The cumulate compositional profile resulting from such a sequence of events is modelled in Figure 5.24c. A rapid shift to more evolved compositions is followed by a "reversed" trend, with a progressively increasing abundance for a compatible element.
Figure 5.21. Cumulate compositional profile that results following eruption of magma. Two cases are shown with $X=Z=0.1$ for compatible element ($D=3$) in which $Y=0.8$ for one cycle, and $Y=0.5$ for 3 cycles.

5.5 Wadi Bani Kharus cycles in the light of open system fractionation

The interpretation of the data in Figures 5.11 and 12 indicates that cyclic compositional variation is present on a vertical scale of about 50m. It could be argued that a finer-scaled compositional variation remained undetected because the sampling interval of 20m was inadequate for it to be revealed. This may be the case, but it can be said that if such a scale of compositional variation is present, then its amplitude is less than the observed amplitude of compositional variation, otherwise the overall chemical coherence that is seen would be obliterated.

Of the "normal" cycles Ib is the best defined. Compositional data for NiO in olivine and Cr$_2$O$_3$ in clinopyroxene will be used to determine the maximum height of the column of liquid (see previous section) that fractionated to form the observed thickness of cumulate. (It is assumed that no undetected finer scaled geochemical variation is superimposed upon cycle Ib.)
Figure 5.22. Cumulate compositional profile that results following input of more primitive magma. Case shown with condition $X=0.1$, $Y=0$, $Z=0.1$ for compatible element ($D=3$) when, for three cycles, $C$ increases from 6.6 to 16.7.

Determination of suitable bulk distribution coefficients is not straightforward; the rocks at all levels show evidence of crystal sorting; no "average rocks" are available which might represent unsorted crystal extracts. Mean modal proportions of the olivine-bearing assemblages in cycle Ib are olivine 10%, clinopyroxene 35% and plagioclase 55%. Using suitable density values these convert to olivine 12%, clinopyroxene 38%, and plagioclase 50% by weight.

The following values for mineral-liquid distribution coefficients are used (Cox et al. 1979):

<table>
<thead>
<tr>
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<th>olivine</th>
<th>clinopyroxene</th>
<th>plagioclase</th>
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</thead>
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<tr>
<td>Cr</td>
<td>2</td>
<td>10</td>
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</tr>
<tr>
<td>Ni</td>
<td>10</td>
<td>2</td>
<td>0.01</td>
</tr>
</tbody>
</table>

and the mean abundances and heights of the samples at the top and bottom of cycle Ib:

<table>
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<tr>
<th></th>
<th>Ni$_{ol}$</th>
<th>Cr$<em>{2O</em>{3}}$</th>
<th>Cpx Height (metres)</th>
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<tr>
<td></td>
<td>2393</td>
<td>0.11</td>
<td>0.27</td>
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</table>

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Figure 5.23. Cumulate compositional profile that results following a change in the fractionating assemblage, with the result that a compatible (D=3) element becomes incompatible (D=0.5) for three cycles. Condition X=0.1, Y=0, and Z=0.1.

Bulk distribution coefficients for the observed assemblage are:

\[ D_{Cr} = 3.53 \]
\[ D_{Ni} = 1.91 \]

Values for Ni in clinopyroxene and plagioclase, and Cr in olivine and plagioclase, are determined by assuming equilibrium between olivine, clinopyroxene, plagioclase, and liquid giving the following values for whole rock Cr$_2$O$_3$ and NiO:

<table>
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<th>NiO</th>
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<tbody>
<tr>
<td>2393</td>
<td>0.095</td>
<td>0.019</td>
</tr>
<tr>
<td>2396</td>
<td>0.011</td>
<td>0.009</td>
</tr>
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</table>

Equation 5.5 allows calculation of the maximum height of the column of liquid that cycle Ib could have fractionated from in a tabular shaped magma chamber. For Cr$_2$O$_3$

\[ l = \frac{[82.00 - 26.60]}{[1 - (0.011/0.095)^1/(3.53-1)]} \]

\[ = 96.6 \text{ m} \]
Figure 5.24. Cumulate compositional profile that results following mixing in of overlying, evolved magma for compatible \( D=3 \) element. Initial condition shown in (a) where \( X=0.1, Y=0, \) and \( Z=0.1, \) which achieves steady state \( C_1=16.7. \) In (b) overlying, evolved magma \( C=1.7 \) is mixed instantaneously. Initial conditions are then restored.

and for Ni0

\[
l = \frac{82.00 - 26.60}{1 - (0.009/0.019)^{1/(1.91-1)}} \approx 98.92 \text{ m}
\]

which are in good agreement. The ratio of the observed cumulate thickness to the calculated magma chamber height indicates that a mass fraction \( X \) of 0.57 had crystallized during cycle Ib.

For a magma chamber triangular in cross-section, the walls of which dip inwards at 30° (see Chapter 8) then from equation 5.8 a liquid column height of 188.15m is calculated. For semi-circular magma chamber, a liquid column height of 162.94m is obtained from equation 5.10.

These values for the thickness of the column of liquid that fractionated to form the rocks of cycle Ib are an order of magnitude less...
than the vertical extent of the Cumulate Sequence in Wadi Bani Kharus. If it is argued that the thickness of the Cumulate Sequence should reflect the depth of the magma chamber in which it formed, then the values obtained above can only be explained in terms of a stratified magma chamber (Turner and Chen 1974, McBirney and Noyes 1979, Huppert and Sparks 1980a,b) in which bodies of liquid of limited vertical extent behaved temporarily as closed fractionating systems.

Within each of the larger cycles I, II and III (ignoring, for the moment, the "reversed" cycles IIe and f) it is noted that for compatible elements or ratios, as well as a progressive depletion in terms of the abundances within each of the smaller constituent cycles (a,b,c etc.), a progressive depletion in terms of the successive primitive "resets" that start each constituent cycle is also seen. This is characteristic of an open system tending towards a steady state composition.

In the foregoing investigation of open system fractionation four mechanisms were identified for producing sustained "reversed" geochemical gradients in cumulate compositional profiles:

(i) eruption/isolation of magma
(ii) more primitive magma input
(iii) change in bulk distribution coefficient
(iv) mixing in of overlying, evolved magma.

In the case of cycles IIe and f we may reject the third possibility (Figure5.23); no change in the assemblage of ol + cpx + plag is seen at any point in the section. No discontinuity is observed at the junction between IId and e, merely a reversal in gradient; on the basis of Figure 5.24 the possibility of mixing in an overlying, evolved magma is excluded, though it remains an option to account for the discontinuity between IIe and f. It is difficult to differentiate between the end results of mechanisms (i) and (ii); this is to be expected as eruption of magma temporarily reduces the mass of the chamber allowing the effects of primitive magma from below to momentarily reset the mixed liquid compositions to more primitive compositions - exactly the same effect that is achieved by making the input
composition more primitive itself. Both these mechanisms are considered viable alternatives in the case of Wadi Bani Kharus.

The "decoupling", notably by TiO₂ and Na₂O in clinopyroxene, does not have a satisfactory explanation. Precipitation of magnetite would account for sudden depletion rather than enrichment in TiO₂ in clinopyroxene, but no petrographic evidence supports this. The fact that it is two relatively highly incompatible elements, Na and Ti, that show decoupling, and that it is also these elements that show the strongest development of zoning, suggests that some later stage intercumulate process may be the mechanism responsible.

5.6 Major element covariation between cumulus phases

The mean values of Fo, An and Mg* contents for coexisting olivine, plagioclase, and clinopyroxene respectively have been plotted against each other in Figure 5.25.

The strong correlation between Mg* content of clinopyroxene and Fo content of olivine is expected; it reflects the closely similar behaviour of Mg and Fe between these phases. The gradient is slightly less than unity and indicates a lower bulk distribution coefficient for Mg/Fe between olivine and liquid than for clinopyroxene and liquid.

The plots between An content of plagioclase and Fo content of olivine or Mg* content of clinopyroxene are, at first sight, much less well defined correlations. However, if both graphs are considered in the light of the cycles erected on the basis of the variation of phase chemistry with height (see Figure 5.11 and 5.12), several important features emerge (Figure 5.25):

(i) For those cycles that contain three or more data points good correlations are present; if more data points were available for the remaining cycles it is suggested that the same would be true for these.

(ii) The observed correlations appear to radiate from the composition points Fo₈₅,An₈₅ and Mg*₉₀,An₈₅.
(iii) The gradients of the observed correlation for any one cycle decrease systematically with the height of the cycle from the Petrological Moho.

The contrasting gradients between different cycles provides independent support of the cryptic units erected on the basis of the variation of phase chemistry with height.

In recent years workers such as Roeder (1975), Nathan and van Kirk (1978), Weill et al. (1980) and Langmuir and Hanson (1981) have attempted empirical or thermodynamic modelling of the major element evolution of magmas. Knowing the starting liquid composition, they have sought to predict the temperatures at which different minerals will crystallise, and what their compositions will be. Subtraction of these mineral compositions allows a line of liquid descent to be calculated. Such methods show considerable promise but their application to the modelling of cumulate suites requires knowledge of an initial primitive liquid composition. Chilled margins or border groups are absent from the Cumulate Sequence of the Oman Ophiolite, and the dykes of the Sheeted Dyke Complex are not only fractionated but have also suffered hydrothermal alteration which has modified their major element composition. The major element compositions of the Cumulate Sequence are unaffected by such alteration but, as yet, no method for predicting the major element composition of their coexisting liquid compositions is available. Therefore only a qualitative treatment of the major element covariation between coexisting phases of the Wadi Bani Kharus Cumulate Sequence can be attempted here.
Figure 5.25. Plots showing the covariation between coexisting phases of mean contents of Mg*-Fo, An-Fo, and An-Mg* for olivine, clinopyroxene, and plagioclase. The data points are labelled according to the cycle to which they belong (Figures 5.11 and 5.12). The lowest cycle Ia is A, the highest cycle IIIId is N.
Consider a magma of a given composition fractionating an assemblage of olivine, clinopyroxene, and plagioclase under conditions of fixed total and volatile pressure. The line of liquid descent will be fixed; the locus of compositions will not in general be a straight line (major element distribution coefficients change with both temperature and composition, and the proportions of the phases fractionating may change down temperature), nor will the locus of coexisting mineral compositions (see the hypothetical graph of successive coexisting olivine and plagioclase compositions in Figure 5.26a). A family of similar loci will exist for all other starting liquid compositions.

What then are the mechanisms available for producing the dispersion in the Fo-An and Mg*-An correlations seen in the Wadi Bani Kharus cyclic units (Figure 5.25)?

If the proportions of olivine, clinopyroxene, and plagioclase fractionating were to change in different ways in different cycles then contrasting loci of coexisting mineral compositions will result to reflect the change in lines of liquid descent. Mean modes for cycle Ib and IIIc (cycles B and M respectively in Figure 5.25) are Ol7.5 Cpx36 Plag56.5 and Ol3 Cpx38.5 Plag58.5 respectively; only the proportion of olivine is different. Figure 5.26b shows the consequence of changing the mode of olivine with a fixed ratio of clinopyroxene to plagioclase on the line of liquid descent. In terms of the Fo-An and Mg*-An plots (Figure 5.25) this has the opposite effect to that required; decreasing the mode of olivine fractionating will increase the rate at which An content in plagioclase varies for a given change in Fo content in olivine (and similarly for Mg* and An). If the lower mode of olivine in the higher cycles of the Wadi Bani Kharus gabbro section has any effect on the correlation between Fo-An (or Mg*-An), it is swamped by some more important petrological process.
Figure 5.26. (a) Hypothetical loci of coexisting olivine and plagioclase compositions. (b) Plot of CaO vs. MgO to illustrate the effect of varying the mode of olivine at fixed clinopyroxene/plagioclase ratio in a olivine gabbro fractionating assemblage on the resulting line of liquid descent. (c) Effect of increasing water pressure (see also inset), or increasing cooling rate, or decreasing the mode of olivine at constant clinopyroxene/plagioclase ratio on the loci of coexisting olivine and plagioclase compositions. (d) Possible effects of open system fractionation on loci of olivine and plagioclase compositions if individual loci are significantly concave towards the plagioclase axis.
An increase in the partial pressure of water not only depresses but also "steepens" the solidus (see inset, Figure 5.26c) of the plagioclase feldspars (Yoder et al. 1957, Yoder 1965). For a given decrease in temperature this results in a smaller change in An content of plagioclase relative to the change that occurs in Fo content of olivine under water-free conditions (Figure 5.26c). A progressive increase in $P_{H_2O}$ would be required to produce the progressive change in gradient of the Fo-An (or Mg*-An) correlation that is observed.

Naslund (1976) reported a systematic change in the rates of evolution of coexisting olivine and plagioclase compositions within the upper part of the Layered Series of the Skaergaard intrusion. Moving from the centre of the pluton towards the eastern margin a progressively smaller change in An content of plagioclase for a given change in Fo content of olivine was observed. McBirney and Noyes (1979) have interpreted this feature as being due to the effects of more rapid cooling rates at the margins of the intrusion in decreasing the rates of diffusion of some chemical components more than others. Where cooling rates were more rapid the diffusion rates of complex silica and alumina ions were lower than those of the simple ions of iron and magnesium; the rate of fractionation of plagioclase was lower where the rate of crystallization was too rapid for equilibration to be approached as closely as it was in olivine and pyroxene. As a result a small change in An content of plagioclase occurs for a given change in Fo content of olivine (Figure 5.26c). Such a mechanism would require the rates of cooling to be more rapid at higher levels, than at the base of the magma chamber that precipitated the Wadi Bani Kharus Cumulate Sequence.

Neither higher partial pressures of water, or higher rates of cooling with concomitant lower diffusion rates, can be rejected as mechanisms to explain the observed features of the Fo-An and Mg*-Pl covariations; both are plausible in that upward migration of volatiles and faster rates of cooling are likely to occur at higher levels in any magma chamber. However it may be noted that if the partial pressure of water was
elevated during the formation of the higher levels of the gabbroic section, then it was never high enough to stabilise amphibole; also no textural differences have been discerned within the section that could be attributed to a variation in cooling rates.

Another mechanism that might explain the observed features of the Fo-An (and the Mg*-An) covariation is illustrated in Figure 5.26d, whereby the variations that are seen with height are a direct consequence of open system fractionation. If the lines of liquid descent (and hence the loci of coexisting olivine and plagioclase (or clinopyroxene and plagioclase) compositions) are significantly concave toward the plagioclase axis, then a primitive liquid in equilibrium with P might fractionate to give a locus of coexisting olivine and plagioclase along PQ. If a primitive liquid input, again in equilibrium with P, then mixed with the liquid in equilibrium with Q, a liquid in equilibrium with R might result. The liquid in equilibrium with R then fractionates to produce a cycle of coexisting olivine and plagioclase compositions along RS, which has a differing gradient to the locus of coexisting compositions from the first cycle PQ. Continued cycles of fractionation, input and mixing will produce a spectrum of coexisting olivine-plagioclase compositions that migrate to trends characterized by limited evolution of An content of plagioclase for a given change in Fo content of olivine. Testing of such a model requires a better knowledge of the detailed form of the Fo-An (and Mg*-An) covariation within each cycle, and whether such a curved locus of coexisting phases is realistic in terms of major element model of fractionating basaltic liquids.

Calculations in section 5.5 indicated that the cumulates of the Wadi Bani Kharus section formed in a stratified magma chamber in which layers of liquid of limited vertical extent (between 100-200m) acted temporarily as closed systems to precipitate the individual cyclic units. If such liquid stratification was the result of double diffusive convection (Turner and Chen 1974, McBirney and Noyes 1979), it may be that the diffusive processes in such a system combined to produce contrasting compositional gradients within the magma body, which in turn led to the contrasting correlations
with height seen between Fo-An and Mg*-An contents of the coexisting olivine, clinopyroxene and plagioclase.
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Table 5.2. Olivine microprobe analyses for Wadi Bani Kharus. For samples 2391-2422 the height above the Petrological Moho is in metres measured normal to layering. The heights for the remaining samples are estimates in metres normal to the Petrological Moho.

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P Browning 184 Chapter 5.6
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P Browning 185 Chapter 5.6
Table 5.3. Plagioclase microprobe analyses for Wadi Bani Kharus. C=core, R=rim, H=50/50, I=interstitial, 0=no description. For samples 2391-2422 the height above the Petrological Moho is in metres measured normal to layering. The heights for the remaining samples are estimates in metres normal to the Petrological Moho.
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P Browning 189 Chapter 5.6
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P Browning 190 Chapter 5.6
### Table 5.4 Pyroxene microprobe analyses from Wadi Bani Kharus.
C=core, R=rim, S=small interstitial, O=no description. For samples 2391-2422 the height above the Petrological Moho is in metres measured normal to layering. The heights for the remaining samples are estimates in metres normal to the Petrological Moho.

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### Table 5.5 Amphibole microprobe analyses for Wadi Bani Kharus.
C=core, R=rim, I=interstitial, O=no description. Heights are estimates in metres normal to the Petrological Moho.

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