
THE GEOLOGICAL FIELD GUIDE SERIES

The Field Description of Igneous Rocks

SECOND EDITION



Dougal Jerram and Nick Petford

The Field Description of Igneous Rocks

The Field Description of Igneous Rocks

SECOND EDITION

Dougal Jerram

Durham University

Nick Petford

University of Northampton

 WILEY-BLACKWELL

A John Wiley & Sons, Ltd., Publication

The Geological Field Guide Series

Barnes, J.W. and Lisle, R.J. (2004) *Basic Geological Mapping*, 4th edn. ISBN: 978-0-470-84986-6, 5th edn publishing (2011). ISBN: 978-0-470-68634-8

Fry, N. (1991) *The Field Description of Metamorphic Rocks*.
ISBN: 978-0-471-93221-5

McClay, K.R. (1991) *The Mapping of Geological Structures*.
ISBN: 978-0-471-93243-7

Milsom, J. and Eriksen, A. (2010) *Field Geophysics*, 4th edn.
ISBN: 978-0-470-74984-5

Tucker, M.E. (2011) *Sedimentary Rocks in the Field*, 4th edn.
ISBN: 978-0-470-68916-5

This edition first published 2011 © 2011 by John Wiley & Sons, Ltd.

Wiley-Blackwell is an imprint of John Wiley & Sons, formed by the merger of Wiley's global Scientific, Technical and Medical business with Blackwell Publishing.

Registered office: John Wiley & Sons, Ltd, The Atrium, Southern Gate, Chichester, West Sussex, PO19 8SQ, UK

Editorial Offices: 9600 Garsington Road, Oxford, OX4 2DQ, UK
The Atrium, Southern Gate, Chichester, West Sussex, PO19 8SQ, UK

111 River Street, Hoboken, NJ 07030-5774, USA

For details of our global editorial offices, for customer services and for information about how to apply for permission to reuse the copyright material in this book please see our website at www.wiley.com/wiley-blackwell

The right of the author to be identified as the author of this work has been asserted in accordance with the UK Copyright, Designs and Patents Act 1988.

All rights reserved. No part of this publication may be reproduced, stored in a retrieval system, or transmitted, in any form or by any means, electronic, mechanical, photocopying, recording or otherwise, except as permitted by the UK Copyright, Designs and Patents Act 1988, without the prior permission of the publisher.

Designations used by companies to distinguish their products are often claimed as trademarks. All brand names and product names used in this book are trade names, service marks, trademarks or registered trademarks of their respective owners. The publisher is not associated with any product or vendor mentioned in this book. This publication is designed to provide accurate and authoritative information in regard to the subject matter covered. It is sold on the understanding that the publisher is not engaged in rendering professional services. If professional advice or other expert assistance is required, the services of a competent professional should be sought.

Library of Congress Cataloguing-in-Publication Data

Jerram, Dougal.

The field description of igneous rocks / Dougal Jerram, Nick Petford. – 2nd ed.

p. cm. – (Geological field guide)

Rev. ed. of: The field description of igneous rocks / R.S. Thorpe and G.C. Brown. Milton Keynes [Buckinghamshire] : Open University Press ; New York : Halsted, 1985.

Includes index.

ISBN 978-0-470-02236-8 (pbk.)

1. Igneous rocks – Identification. 2. Geology – Fieldwork. I. Petford, N. (Nick) II. Title.

QE461.T45 2011

552'.1 – dc22

2010034826

A catalogue record for this book is available from the British Library.

This book is published in the following electronic format: epdf 9780470022375

Set in 8.5/10.5pt Times by Laserwords Private Limited, Chennai, India.

First Impression 2011

*This book is dedicated to Bob Hunter, and his great contribution to
our understanding of rock textures.*

CONTENTS

Preface	xi
Acknowledgements	xv
1 Introduction and Occurrence	1
1.1 The Importance of Fieldwork	1
1.2 The Global Picture – Igneous Rocks in Relation to Regional Tectonics	2
1.3 Mode of Occurrence of Igneous Bodies	4
1.4 Summary	11
2 Field Skills and Outcrop Structures	15
2.1 Equipment	15
2.2 Preparing Maps and Basic Mapping	16
2.3 Notebooks and Data Recording	17
2.4 Primary Outcrop Structures	18
2.5 Secondary or Late Stage Outcrop Structures	28
2.6 Outcrop Contact Relationships	32
2.7 Summary of Igneous Outcrop Descriptions	33
3 Igneous Textures and Classification	37
3.1 Describing Rock Types	37
3.2 Colour and Composition	38
3.3 Texture, Grain-Size/Shape and Fabric	43
3.4 Mineral Identification	49
3.5 Naming and Classification	58
4 Volcanics 1 – Lava Flows	69
4.1 Lava Flow Emplacement Mechanisms	69
4.2 A Compositional Divide for Lava Flows	71
4.3 Mafic/Basaltic Lava Flows	73
4.4 Felsic/Silicic Flows	80
4.5 Pillow Lavas and Hyaloclastites	82
5 Volcanics 2 – Pyroclastic Rocks	93
5.1 Structures, Textures and Classification	93
5.2 Pyroclastic Flows and Ignimbrites	101
5.3 Scoria Cones	108

CONTENTS

5.4	Water/Magma and Sediment/Magma Interactions	109
5.5	Epiclastic Deposits	112
6	Shallow-Level Intrusions	119
6.1	Sill and Dykes	119
6.2	Working Out Emplacement History	124
6.3	Volcanic Plugs and Diatremes	130
6.4	High-Level Subvolcanic Intrusions	133
7	Granitic Complexes	137
7.1	Introduction	137
7.2	General Features and Occurrence	137
7.3	Zoned Plutons	142
7.4	Internal Structures and Textures	145
7.5	Internal Contacts	150
7.6	Emplacement Timing	158
7.7	Distinctive Granitoid Textures	162
7.8	Metamorphic Aureoles	164
7.9	Summary of the Field Characteristics of Granitic Complexes	165
8	Mafic Complexes	171
8.1	General Features and Occurrence	171
8.2	Continental Mafic-Ultramafic Intrusions	173
8.3	Ophiolite Complexes	177
8.4	Komatiites	183
8.5	Summary of the Field Characteristics of Mafic-Ultramafic Intrusions	184
9	Magma Mixing and Mingling	189
9.1	Magma Rheology	189
9.2	Magma Mixing	190
9.3	Magma Mingling	192
9.4	Synplutonic Dykes and Sills	196
9.5	Magma Mingling in Subvolcanic and Volcanic Environments	200
9.6	Xenoliths	201
9.7	A Word of Warning	202
9.8	Summary	203
10	Mineralisation and Geotechnical Properties	207
10.1	Mineralisation and Key Minerals	207

CONTENTS

10.2	Mineralisation in Layered Mafic Intrusions	209
10.3	Geotechnical Properties of Igneous Rocks	213
10.4	Rock Mass Classification	216
10.5	Summary	226
Appendix		229
Further Reading		231
Index		233

PREFACE

Igneous rocks in the field present a bewildering array of lithologies formed not just from cooling and crystallisation of intrusive material but also by sedimentation and surface flow from volcanic eruptions. This concise guide is designed to give students, professionals and keen amateurs of igneous geology the key tools needed to help understand and interpret better the origin and evolution of complex igneous systems in a focused way. This extensively revised and reorganised colour guide builds on the original version by Richard Thorpe and Geoff Brown of the Open University (both now deceased) and published in 1985 as part of the Geological Society of London Field Guide Series. Since then much has changed in igneous petrology including significant advances in our understanding of the physical processes that govern the emplacement of volcanic and plutonic material both on the surface and within the solid earth. For example, new models for the emplacement of pyroclastic flows have helped explain the origin of some of the more enigmatic features of ignimbrites, while magma emplacement studies now draw heavily on techniques used routinely in structural geology. It is thus not enough just to know how to identify minerals in the field – multidisciplinary skills borrowed from other branches of field geology, and even engineering, are not desirable aspirations, but essential must haves. As our understanding of the subject has increased, we have tried in this first colour revision to incorporate much of this new thinking. At the same time we have remained true to the original philosophy of a portable guide that explains clearly and concisely the basic concepts underpinning igneous geology in the field. To this end we hope that the inclusion of new colour images and a colour-coded index system make basic identification of rock types and structure a slightly easier task. Both of us grew up with the original version and it has been a pleasure to revise and build from the start on a text so well laid out. We hope you find the new colour guide as great a companion as we did the old, and that in revised form it is even more of an essential aid when confronted, perhaps for the first time, with igneous rocks in the field.

As we write, the Eyjafjallajokull volcano is erupting in Iceland, causing severe disruption to air traffic across Europe and beyond while generating unexpected but welcome global media interest in volcanology. This excitement will die down soon enough, but a raised awareness and respect for volcanoes

PREFACE

will linger in the public psyche for years to come. Its aftermath will present new opportunities to those willing to grapple with the complexity and scientific challenges of ‘next generation’ igneous petrology. Our hope is that this book will in small measure inspire new entrants into the field, just as the original did for us.

Dougal Jerram and Nick Petford
April 2010

Meet the Authors



Dougal Jerram

Dougal is currently at the Department of Earth Sciences, Durham University where he is involved in research and teaching, particularly of field geology. His main expertise is in rock microstructure and textural analysis, 2D–3D modelling of rock textures and volcanic basins, volcanology, sedimentology and field geology. In recognition of his early significant contribution to Earth Sciences he was awarded the Murchison Fund of the Geological Society in 2006. Recently Dougal has developed a keen eye for science outreach, and has appeared on national and international TV (BBC, Discovery, National Geographic, History Channel, Channel 4) promoting aspects of the Earth, and has developed a popular web presence through www.dougalearth.com.

Nick has published extensively on both field and theoretical aspects of igneous geology. After completing a PhD on Andean magmatism and geochemistry he switched to work more on the physical mechanisms governing the ascent and emplacement of granitic magmas and magma rheology. Other research themes include melt segregation in meteorites, the geotechnical properties of volcanic rocks and volcano tourism and economic regeneration. Nick began his research career as a Royal Society University Research Fellow and has worked at the Universities of Liverpool, Cambridge, Kingston and Bournemouth. He was awarded the Murchison Fund of the Geological Society in 1999 and has held visiting professorial appointments in Europe, USA and Australia. Like Dougal he makes the occasional TV appearance. In 2010 he was appointed Vice Chancellor of the University of Northampton.



Nick Petford

ACKNOWLEDGEMENTS

DJ would like to thank James Haythornthwaite and Tim Watton for their skillful help in drafting some of the diagrams. Lorraine Field, Richard Walker, James Haythornthwaite, Tim Watton, Jon Davidson, Graham Pearson, Nick Schofield, Robin Gill, Tony Philpotts, Sergio Rocchi, Scott Patterson and Bruce Marsh are thanked for providing photographs and other graphic materials and support. Additional discussions with Catherine Nelson, Clayton Grove, Sam Clark, Kirstie Wright and the vibrant student and postgraduate community at Durham University proved very helpful. Henry Emeleus and Jon Davidson are also thanked for their comments on earlier drafts of some chapters. In addition to our many friends and colleagues (thanks, you know who you are!) who have informed and enlightened us in the field NP would like to extend his gratitude to Mike Atherton, Wally Pitcher for the immortal ‘at the end of a long day in the field, a note saying white rock cuts black rock will do!’, Mr and Mrs A.P. Boyle, Donny Hutton, Colin Key, Ken McCaffrey, Stephano Pugliese, Mark Thomas and Eddie Bromhead. Fiona Woods and Rachael Ballard at Wiley are thanked for their help and patience. The final word goes to our long suffering families for putting up with all of this. Just wait until the next edition.

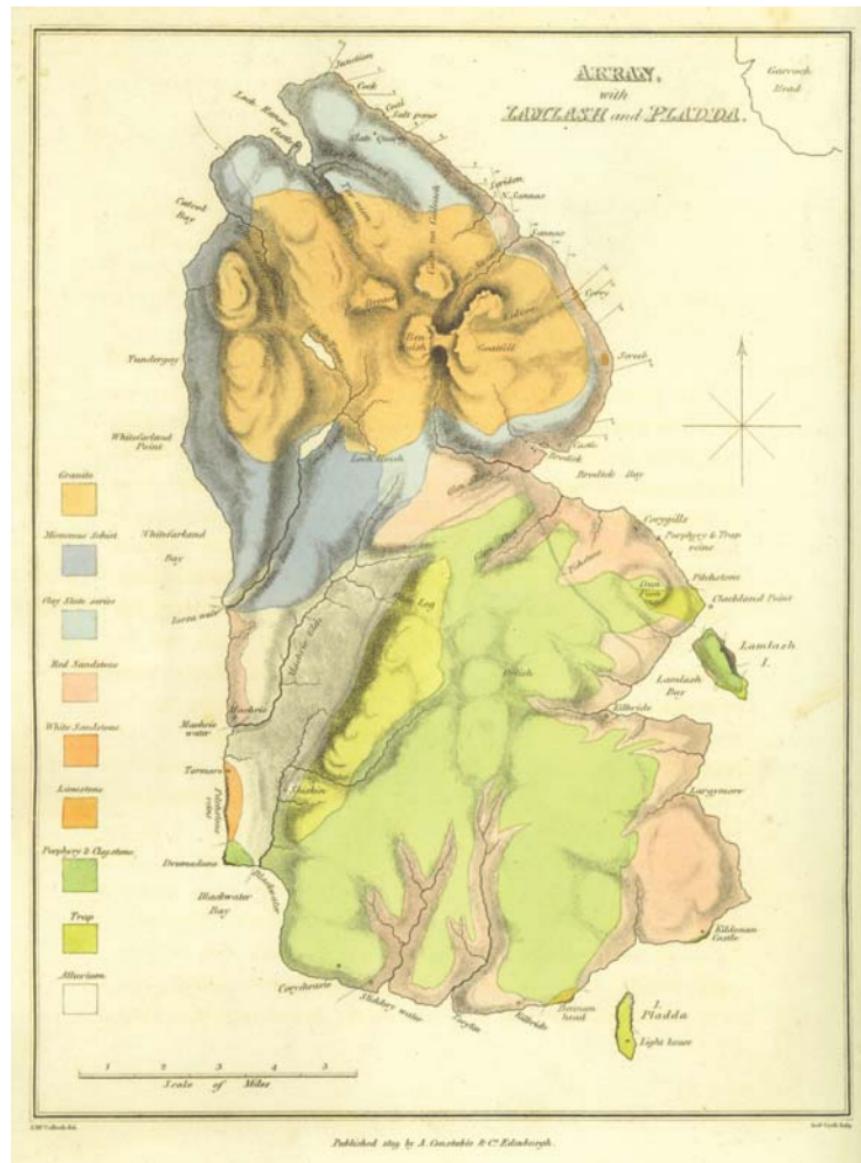
Illustrative matter appearing in this book was adapted from or inspired by the following sources to whom also grateful acknowledgement is made:

Figures 3.2, 3.11–3.14: R.V. Dietrich and B.J. Skinner, *Rocks and Rock Minerals*, Wiley (1979) A.G. Tindle; Figure 1.2: H. Williams and A.R. McBirney, *Volcanology*, Freeman, Cooper and Co. (1979); Figure 4.8: P. Lyle, *J. Geol. Soc. Lon.* **157** (2000); Figure 4.12 R.A.F. Cas and J.V. Wright, *Volcanic Successions*, Chapman & Hall (1988) & M. J. Branney, B. Bonnichsen, G. D. M. Andrews, B. Ellis, T. L. Barry and M. McCurry, *Bull. Vol.* **70**(3) (2008); Figures 4.13, 4.17: G.A. Macdonald, *Volcanoes*, Prentice-Hall (1972); Figures 4.18, 4.19: R. Batiza and J.D.L. White In: *Encyclopedia of Volcanoes*, Academic Press (2000); Figure 5.3: K.H. Wohletz and M. F. Sheridan, *Geol. Soc. Am. Spec. Paper* **180** (1974); Figure 5.5: R.V. Fischer and H-U Schmincke, *Pyroclastic Rocks*, Springer-Verlag (1984); Figure 5.14: M.J. Branney and P.B. P Koke-laar, *Geol. Soc. Memoir* **72** (2002); Figure 5.16: R.S.J. Sparks, S. Self and G.P.L. Walker, *Geology* **1** (1973); Figures 5.15, 5.25, 8.14: R. Gill, *Igneous Rocks and Processes*, Wiley (2010); Figure 5.20: D. Dirk Vesperman and H-U. Schmincke, In: *Encyclopedia of Volcanoes*, Academic Press (2000); Figure 5.21: J.L. Smellie, In: *Encyclopedia of Volcanoes*, Academic Press (2000); Table 5.1: R. V. Fisher, *Bull. Geol. Soc. Am.* **72** (1961); Figure 6.12: M. G. Best and

ACKNOWLEDGEMENTS

E.H. Christiansen, *Igneous Petrology*, Blackwell (2001); Figure 6.6: Hutton, *Pet. Geosci.*, **15** (2009); Figure 7.1: F.H. Lahee, *Field Geology*, McGraw-Hill (1916); Figure 7.2: M.A. Bussell, W.S. Pitcher and P.A. Wilson, *Can. J. Earth. Sci.* **13** (1976); Figure 7.5: M. Haederle and M.P. Atherton, *Tectonophysics* **345** (2002); Figure 7.11: D.H.W. Hutton, 1988. *Trans R. Soc. Ed.: Earth Sci.* **91** (2000); Figure 7.13: N. Petford and M.P. Atherton, *Tectonophysics* **205** (1992); Figure 7.14: W.S. Pitcher and D.H.W. Hutton, *Geol. Sur. of Ireland* (2003); Figure 7.21: J.G. Arth, F. Barker, D.E. Peterman and I. Friedman, *J. Petrol.* **19** (1978); Figure 7.22: W.S. Pitcher, *The Nature and Origin of Granite*, Blackie Academic Professional (1993); Figures 8.8, 8.9: J.D. Winter, Pearson Education/Prentice Hall (2001); Figure 9.1: E. Hallot *et al.*, *J. of Volc. Geo. Res.* **71**, 1996; Figure 9.4: S. Pugliese and N. Petford, *Visual Geosci.*, **6** (2001); Figure 9.10: B. Barbarin, *Schweizerische Mineralogische und Petrographische Mitteilungen*, **69**, (1989); Figure 10.3: R.H. Sillitoe, *Geol. Ass. Canada Spec. Paper* **14** (1976); Figure 10.9: M.O. Sarr and M. Manga, *Geophys. Res. Lett.* **26** (1999); Figure 10.8: D. Norton and R. Knapp, *Amer J. Science* **277** (1977); Figure 10.11: M. Thomas, PhD Thesis, Kingston University (2007).

INTRODUCTION AND OCCURRENCE



Early map of Arran from John Macculloch's *Western Islands of Scotland* 1819 book.

1

INTRODUCTION AND OCCURRENCE

Igneous rocks occur in almost all environments at the Earth's surface as volcanoes and their dispersed products, and within the crust and mantle as they form the pathway from which the Earth cools from its hot interior. The products of igneous activity not only provide key information about the evolution of the Earth through time, they can be used as key stratigraphic markers, form the basis of our understanding of the Earth's chronology and can be found to have striking effects on our planet's climate. From the beauty of diamonds, the wonders of erupting volcanoes and the polished rocks which adorn buildings all over the world, man has had a fascination with igneous rocks as far back as we can trace. The key basis from which we can understand igneous rocks and their systems is from detailed fieldwork and observation. This guide aims to provide the basic information and tools to enable earth and engineering scientists from a variety of backgrounds to investigate the wonderful world of igneous rocks in the field.

1.1 The Importance of Fieldwork

The most fundamental observations that you can make within the earth sciences are in the field. Here we are able to characterise the occurrence of earth materials *in situ*, in relation to their surroundings and within the context that will underpin any further investigations of the rocks in the lab. In short – *Fieldwork is the basis of all geological studies*. As such it is important that a person who wishes to understand rocks is proficient in the field. Whether one is studying igneous, sedimentary or metamorphic rocks the fundamentals of the fieldwork are more or less the same and indeed one may need to address all three major types of rock in the same field locality. It should be noted that there are many instances in which expensive geochemical and geophysical data have been misinterpreted through incomplete knowledge of basic field relationships. Therefore, *if the appreciation of field geology is poor, then all studies based on collected samples and field measurements will be equally poor*. Conversely, *good appreciation of field geology forms the basis of good geological interpretation*.

As a good field scientist you will need to have a background of basic skills to enable you to make the correct detailed observations that will in turn lead to clear

and well thought out interpretations of the geology. The study of igneous rocks might also include petrological and mineralogical investigation, geochemical and isotopic analysis to determine the age and origin of the rocks and the use of geophysical measurements in the field to determine the distribution of rock-types below the ground. Also, many igneous rocks are associated with distinctive types of economic mineralisation and these are generally discovered and evaluated by fieldwork. Key to the success of these approaches will be the detailed understanding of the rocks in the field.

In this handbook we explain how to observe igneous rocks in the field, from the scale of outcrops down to hand specimens and to tie observations into basic interpretations of how the igneous rocks formed. Before embarking on the details of igneous rocks in the field it is valuable to consider the role of igneous rocks in a global framework and to consider the main occurrence of igneous rocks.

1.2 The Global Picture – Igneous Rocks in Relation to Regional Tectonics

Igneous rocks are materials that have solidified from molten or partially molten material, termed *magma*. Such rocks may be classified as *extrusive rocks*, which were erupted at the surface of the Earth, and *intrusive rocks* that crystallised beneath the surface. Igneous rocks of different compositions and field relationships exist at specific regions on the Earth associated with the plate tectonics in general. This reflects the mode of formation and emplacement of igneous rocks in the context of regional tectonic patterns. Below we briefly describe the key plate margin and within plate associations where patterns of igneous activity can be recognised.

The Earth's crust forms the uppermost part of the outer rigid shell, or lithosphere, of the Earth and is divided into large coherent 'plates' that move in relation to one another. This process termed *plate tectonics* (continental drift) reflects our cooling planet and the convection of the mantle beneath. The plates themselves are split into two types of crust which are defined by their composition and thickness, *oceanic* and *continental* and the configuration of the plates leads to different types of plate margins (boundaries) where specific igneous associations exist.

The boundaries between plates are of four types (summarised Table 1.1, see also Figure 1.7).

1. Constructive plate margins or ocean ridges, where two plates are moving apart and the upwelling and solidification of magma forms new oceanic crust.
2. Destructive plate margins, where two plates are converging so that one plate sinks below the other and is eventually resorbed into the mantle

INTRODUCTION AND OCCURRENCE

or ‘destroyed’. This process is accompanied by formation of a range of magmas. Such plate margins may occur on oceanic (island arcs) or continental lithosphere (active continental margins).

3. Conservative plate margins, are faults where two plates slide past each other (transverse faults), so that lithosphere is neither created nor destroyed, and igneous activity is minor.
4. Collision zones, where two island arcs and/or continents have collided so that subduction of oceanic material has ceased. Such areas are characterised by widespread extrusive and intrusive igneous activity which commonly continues for a considerable time after collision.

Over 99% by volume of igneous activity occurs at constructive and destructive plate margins and at collision zones and some occurs at locations *within* the plates, for example volcanoes such as those of Hawaii and those associated with the East African rift system.

Igneous activity at constructive plate margins is responsible for the formation of the oceanic crust. The composition and structure of the oceanic crust is known from the study of rocks dredged from the ocean floor, from seismic studies and from studies of onshore exposures of older rocks that are believed to be fragments of the oceanic crust (*ophiolites*). These lines of evidence indicate that the oceanic crust consists of layers of basalt lavas, basalt/dolerite dykes, gabbro and peridotite. These rocks form a distinctive association which may be recognised in ancient orogenic belts, where it is termed the *ophiolite association*. The recognition of such associations is clearly of great palaeogeographic significance and the ophiolite associations are described in detail in Chapter 8.

The oceanic lithosphere moves away from the oceanic ridge by the process of sea-floor spreading and is generally returned to the mantle at a destructive plate margin within circa 200 Ma. The descent of oceanic lithosphere into the mantle is accompanied by partial melting above the descending plate where water is driven off at depth. This melting in the overlying mantle forms magmas ranging in composition from basalt, through andesite to rhyolite in composition. These intrude the crust and may be erupted at the surface or emplaced at depth as gabbro, diorite and granite. In some places, the emplacement of such rocks causes melting of the lower crust and this results in the emplacement of intrusions dominantly of diorite, granodiorite and granite composition at destructive continental margins, accompanied by eruption of andesite, dacite and rhyolite. The intrusive rocks emplaced at active continental margins form linear belts of intrusive complexes of diorite-granite composition, often termed batholiths.

The composition of the continental crust broadly resembles that of the igneous rocks of andesite composition. Much of continental crust is thought to have formed as a result of igneous activity of the type seen today at island arcs and at destructive continental margins. The crust has evolved continuously as a result

INTRODUCTION AND OCCURRENCE

of magmatic and metamorphic activity, uplift, erosion and sedimentation, and hence consists largely of metamorphic and igneous rocks with a thin veneer of sedimentary rocks. Because of its greater age and complex geological history, the structure of the continental crust is much more varied than that of the oceanic crust. The continental crust is therefore considered to have a complex structure characterised by rapid lateral and vertical variation, and uplifted sections from which the sedimentary veneer has been eroded expose sections of a wide variety of igneous rocks emplaced at great depth within the crust.

Igneous rocks formed at locations distant from plate margins (locations within-plate, Table 1.1) may have distinctive modes of occurrence, for example as flat-lying sheets of plateau lavas, as discordant plutonic magmatic bodies within continental rifts and as concordant or discordant gabbroic intrusions. Such igneous rocks may have characteristic compositions; indeed, many magmas emplaced at locations within a plate have distinctive alkali-rich chemical compositions which may be reflected in their mineralogy.

1.3 Mode of Occurrence of Igneous Bodies

In general we can split the different types of igneous body into three main sections relating generally to their relative positions within the Earth. Volcanic rocks erupt at and onto the Earth's surface, minor intrusions tend to occur at shallow depths within the Earth's crust and plutonic rocks form larger intrusive bodies at a variety of depths. In the context of this field guide we will generally look at these three main modes of occurrence and different subsections of these (for example, Volcanic – lava, Volcanic – pyroclastic, Plutonic – Granite, Plutonic mafic/ultramafic). Most field areas will be dominated by rocks from one of these general levels, though many examples will exist where different styles of occurrence are found (for example, shallow intrusions with lavas) so these basic subdivisions are used as a guide to the main features of each type. Below we briefly describe these major subdivisions.

1.3.1 Volcanic rock units

Volcanic rocks are classified as *lavas* and *pyroclastic rocks* (volcaniclastic). *Lava* is the term for molten extrusive rock and its solidified product, and pyroclastic rocks are composed of a mix of materials (lava fragments, pumice and crystals) fragmented by explosive volcanic activity. Within the guide we will deal with lava flows in detail in Chapter 4 and pyroclastic rocks in Chapter 5. Below is a brief summary of the main occurrence of volcanic rock units.

Often lavas and pyroclastics associated with individual volcanoes are concentrated within valleys and depressions around the volcano. The most extensive pyroclastic deposits may form large-scale stratigraphic units that blanket the topography and may form plateau-like features around volcanoes, additionally

INTRODUCTION AND OCCURRENCE

pyroclastic deposits may be underestimated in volume due to a large proportion of fine ash material which can be carried great distances from the volcano. The largest lavas flows build up thick sequences (plateaus), and construct large igneous provinces of immense volume.

Volcanoes show a wide variety of forms, depending largely upon the composition of the erupted material and hence the style of eruption (cf. Chapters 4 and 5). Basaltic volcanoes such as Hawaii erupt dominantly (over 80%) lava, the dominantly andesitic products of many volcanoes in island arcs and active continental margins have less than 10% lava and over 90% pyroclastic rocks. Further, the erupted proportion of pyroclastic rocks is often underestimated from subsequent field studies because these materials are often rapidly dispersed by the wind, or are eroded after deposition more rapidly than the equivalent volume of solid lava. Hence, bear in mind that lavas might be over-represented in many island arc and continental margin volcanoes.

Many volcanoes, particularly of andesite composition are *composite* in the sense that they comprise both lava and pyroclastic materials and have a steep irregular conical form (for example, Figure 1.1). Such volcanoes are built by flow of lava down depressions around the volcanoes and the eruption of pyroclastic materials; they commonly have diameters of 10–40 km. Volcanoes associated with mainly basaltic eruptions form relatively low relief shallow sided volcanoes



Figure 1.1 Colima volcano with lava flows and pyroclastic flows which have destroyed parts of the woodland around the mountain flanks.

INTRODUCTION AND OCCURRENCE

known as shields due to the predominantly low viscosity lava, that erupts from them. The smallest volcanic forms result from a single short-lived eruption and comprise a variety of cones and extrusions. These include *pyroclastic/scoria cones* comprised of material of basic and intermediate composition, and steep sided *flows and domes* of more viscous acid lava. Such volcanic forms are generally 1–2 km in diameter and may form upon, or near to, larger volcanic forms, when they may be termed *parasitic volcanoes*.

From this description of volcanic forms and products, it is clear that a single volcanic area may be characterised by a variety of deposits. These may be contemporaneous and, for a terrestrial composite volcano, may be interpreted in terms of variation of *associations* of deposits with distance from the volcano (for example, Figure 1.2). The central zone (within circa 2 km of the central vent) is characterised by lava conduits (later exposed as volcanic plugs, dykes and sills) associated with coarse, poorly-sorted pyroclastic materials which have been deposited near to the vent. The *proximal zone* (circa 5–15 km from the central vent) has a higher proportion of lava flows, with a variety of pyroclastic flow deposits, and the *distal zone* (beyond circa 5–15 km from the central vent, and extending beyond the volcano) is characterised by pyroclastic flow deposits associated with fine air-fall deposits dispersed by wind away from the volcano. These may be interbedded with sedimentary rocks such as lacustrine deposits and reworked volcaniclastics (epiclastic) rocks. Pyroclastic

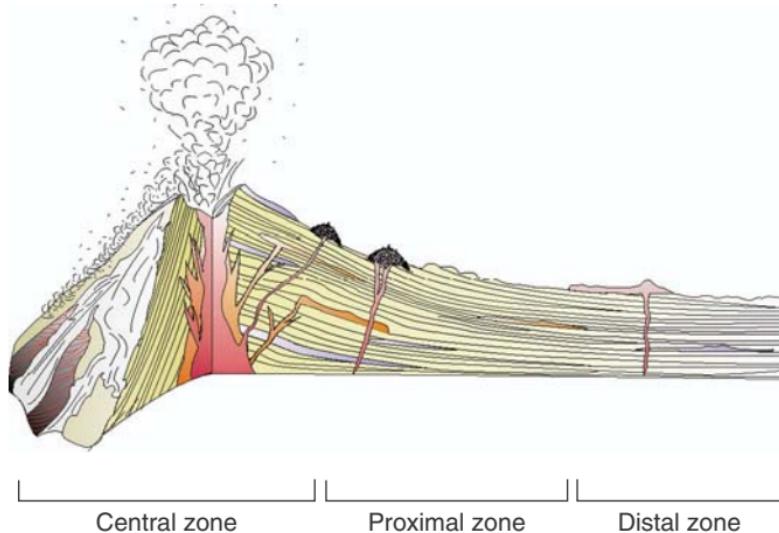


Figure 1.2 Schematic internal architecture of a composite volcano. See text for descriptions of central, proximal and distal zones.

INTRODUCTION AND OCCURRENCE

cones, flows and domes may occur within any of the three zones. Therefore, even for young volcanoes, it may be difficult to correlate individual lava flows, pyroclastic fall and flow deposits with a single eruption. Debris flows (as distinct from pyroclastic flows) formed when material collapses from volcano sides and unconsolidated deposits, may travel several kilometres around source, and can carry great masses of lava as blocks (for example, around Mount Egmont/Taranaki, in North Island New Zealand).

1.3.2 Intrusive rock units

Intrusions vary widely in size and relationship to the country rock and are generally grouped according to size into

1. **minor intrusions**, which have mean minimum dimensions measured in tens of metres (or less) and were emplaced relatively near to the Earth's surface and
2. **plutonic intrusions**, which are commonly emplaced at greater depth and have sizes measured in terms of kilometres.

Within the guide we deal with shallow/minor intrusions in Chapter 6, granitic plutonic rocks (granitic complexes) in Chapter 7 and mafic/ultramafic complexes in Chapter 8. A brief description of the main modes of occurrence is introduced below.

1.3.3 Minor intrusions

The most common forms of minor intrusions are shown schematically in Figure 1.3. *Dykes* are sheet-like intrusions which were approximately vertical at the time of emplacement and are hence *discordant* to host rocks such as shallow-dipping sedimentary rocks. As a consequence of their attitude, the outcrops of dykes are little affected by the topography of the countryside in which they occur and often appear as nearly straight lines on geological maps. The width of dykes ranges from centimetre size to sizes measured in hundreds of metres, but in general the average width is probably in the range 1–5 m. Since this is too small to portray accurately on large-scale geological maps, dykes are often shown with a uniform or ‘conventional’ width (like roads on geographic maps).

Sheet intrusions that were approximately horizontal at the time of emplacement are termed *sills* (for example, Figure 1.4). These are often emplaced into horizontal or shallow dipping sedimentary rocks, in which case they may be broadly concordant, with the stratification, and appear on a geological map as part of the sedimentary succession. Other forms of intrusion which are broadly concordant with the surrounding strata are blister-shaped masses with a sub-horizontal base and elevated upper surface, termed *laccoliths*, and saucer-shaped intrusions, termed *lopoliths* (Figure 1.3); both forms range in size from small

INTRODUCTION AND OCCURRENCE

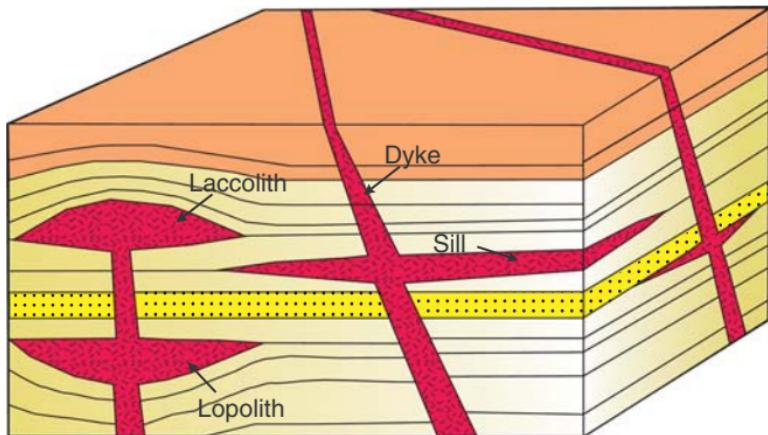


Figure 1.3 Broad types of shallow level intrusions; sills, dykes, lopoliths and laccoliths (see Chapter 6 for detailed section).



Figure 1.4 Thick sills ($>100\text{ m}$) and dykes cutting through sediments in the Transantarctic mountains, Antarctica (total height of section circa 300 m).

INTRODUCTION AND OCCURRENCE

bodies with dimensions measured in metres, to enormous masses hundreds of kilometres in size, which are clearly distinct from smaller minor intrusions.

In studying minor intrusions care should be taken to trace contacts and measure stratigraphic sections. Most contacts are complex: minor intrusions and plutons usually show steep contacts and, since the form and areal distribution of such contacts are used to determine the form of the intrusion in three dimensions, it is particularly important to determine the attitude and orientation of all planar contacts in order to determine the overall geometry. For many minor intrusions it should be possible to locate a contact area with sufficient three-dimensional exposure to determine dip and strike.

1.3.4 Plutonic intrusions and igneous centres

At the larger end of the igneous spectrum we have the major igneous intrusions/centres and plutons. Although there is a huge range of sizes and associations that may be regarded as major plutonic intrusions, in this guide we will be grouping the large plutonic intrusions into those which are mainly granitic in composition (Granite complexes – Chapter 7) and those which are mainly mafic/ultra mafic in composition (mafic complexes – Chapter 8). Additional information about magma mixing/mingling which is also often found within the large plutonic intrusions will be dealt with in Chapter 9.

Coarse-grained granitic plutonic rocks occur most frequently within large, elongate belts of intrusions, 50–150 km in width and 500–1500 km in length, that characterise eroded mountain belts. Such elongate plutonic bodies are termed *batholiths*. They are usually composed of a large number of cross-cutting smaller intrusions, including bodies 5–50 km in size with circular outcrop patterns, ring like intrusions and irregular dykes and sills. Circular intrusions with surface areas of less than 100 km² are sometimes termed *stocks*. The features of the intrusions in an eroded batholith are shown in Figure 1.5. This shows that the major contacts are outward dipping but that where the intrusion has an irregular roof, exposure might leave irregular downward projections of roof country rocks surrounded by igneous rock. These are termed *roof pendants* and can be identified from the outcrop pattern as shown on Figure 1.5. Although the subsurface form of plutonic intrusions may be determined from the attitude of contacts, additional clues come from the extent of the metamorphic aureole and from remote geophysical information such as gravity and magnetics.

The major igneous centres, which are commonly linked as fossil magma chambers to extinct volcanoes and isolated large igneous intrusions form a subset of plutonic igneous rocks. In many instances they have a significant mafic/ultra mafic compositional component, and can be characterised by multiple injections/inflations of magma at a specific location in the Earth's crust (for example, Rum and Skye igneous centres, Scotland – Figure 1.6) or in extreme

INTRODUCTION AND OCCURRENCE

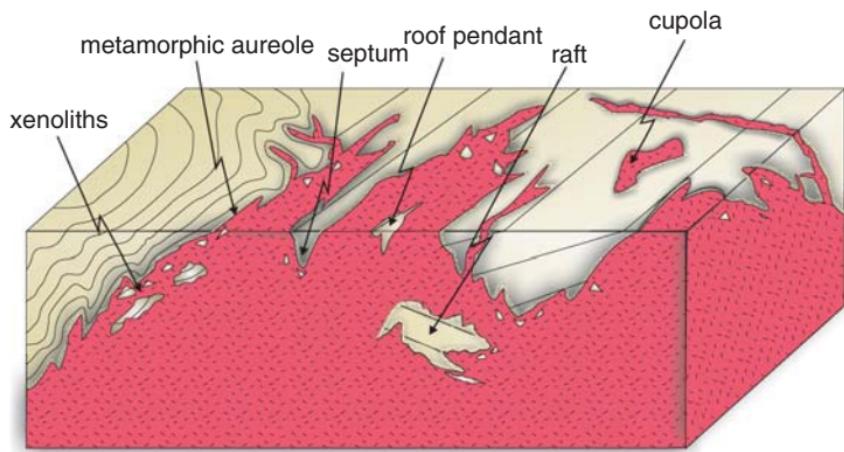


Figure 1.5 Schematic of a granitic pluton highlighting complex contact relationships and associated structures.



Figure 1.6 The igneous centres of Rum and Skye, Scotland. Layered mafic igneous rocks of Rum in the foreground and the (gabbroic) Cullin mountains of Skye in the background.

examples as large single pulses of magma, for example Skaarguard intrusion, Greenland. Although a range from mafic/ultra mafic through to granitic compositions often characterise these intrusions, it is predominantly the mafic/ultra mafic rocks which host some of the most stunning and enigmatic igneous structures, that of layered igneous rocks. Probably the most useful connection that has been made in igneous geology is the linking up of some of the major igneous centres with the volcanic systems that they feed. This realisation that we are able to have a window into the processes going in the depths of a volcanoes magma chamber, has provided great insights into how magma is processed through the crust to the surface.

1.4 Summary

The layout and organisation of this field guide emphasises the occurrence of igneous rock associations within characteristic crustal regions, as introduced in Section 1.3. Chapters 2 and 3, provide information on the key field skills and igneous outcrop-texture description as well as basic classification tables for igneous rocks. Volcanic rocks are looked at in Chapter 4 (lavas) and Chapter 5 (pyroclastics), and the shallow minor intrusions in Chapter 6. Larger plutonic systems are presented in Chapter 7 (granitic complexes) and Chapter 8 (mafic complexes), and the roles of magma mixing/mingling in Chapter 9. Finally we take a look at the mineralisation and engineering properties of igneous rocks in Chapter 10.

INTRODUCTION AND OCCURRENCE

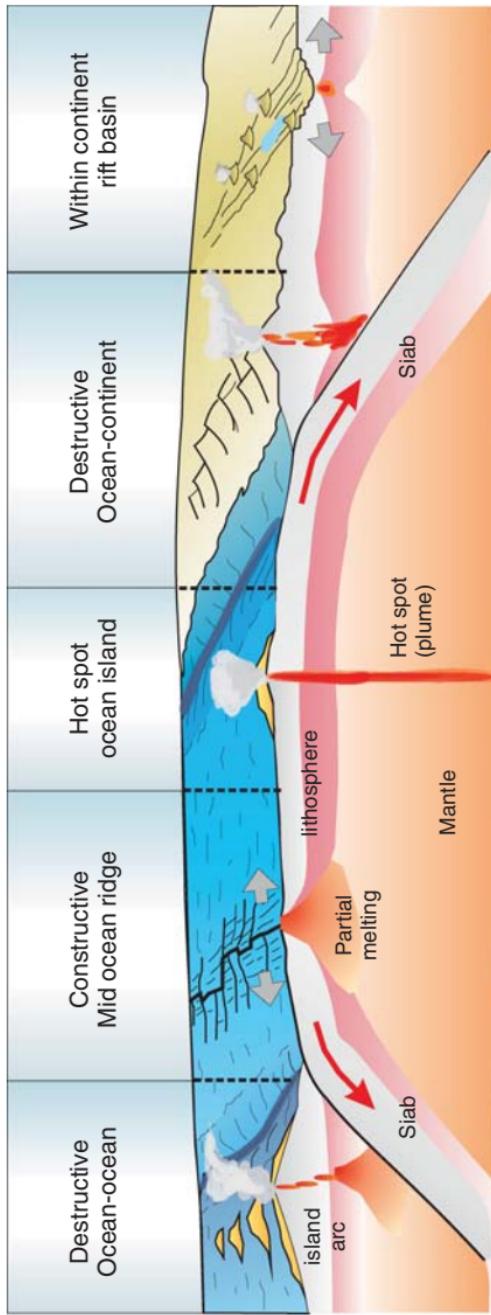


Figure 1.7 Schematic of plate tectonic associations and igneous rocks (see also Table 1.1).

Table 1.1 Classification of igneous activity in relation to tectonic setting (see also Figure 1.7).

	Plate margin		Within plate
Constructive	Destructive	Collision-zone	
Ocean ridge (large ocean basin)	Island arc and transition to active continental margin	Continent-continent and continent-island arc collision	Oceanic
Back arc spreading	Basalt-andesite (in island arcs); andesite-dacite-rhyolite (in active continental margins)	Dacite-rhyolite	Basalt-andesite-rhyolite ¹
Extrusive	Basalt	Gabbro-diorite (in island arcs); diorite, granodiorite and granite (in active continental margins)	Gabbro-diorite-granite ¹
Intrusive		Granite	

¹These rock types have a distinctly different composition and mineralogy from the analogous rock types formed at plate margins, and a range of rarer alkali-rich rock types may also be found in within-plate settings.

FIELD SKILLS AND OUTCROP STRUCTURES



Figure 2.1 Be prepared!! Picture of field kit; hard hat, compass clinometers, tape measure, mapping pens, high visibility jacket, maps, and so on. See text for detailed list.

2

FIELD SKILLS AND OUTCROP STRUCTURES

In order to make meaningful observations of igneous rocks in the field some key skills are useful to help characterise outcrops and gross scale structures before focusing on detailed structures and classification of igneous rocks (Chapter 3). It should be noted that a good igneous field geologist should also be a good field geologist in general, and be versed in structural, sedimentary and metamorphic rocks in the field (see also additional handbooks in this series), as in most cases the igneous rocks under consideration will be in association with other rock types and will have invariably undergone some form of structural deformation. In this section we focus on some key basic fieldwork skills which can be applied to characterise igneous rock outcrops.

2.1 Equipment

When embarking on fieldwork studies the key to the correct equipment is to be prepared. As with any outdoor activity there is a likelihood of significant weather variation so a variety of clothing/footwear appropriate for the terrain and environment is required. Take note of the geological fieldwork code which contains detailed information about field safety, sampling and general rules in the outdoors (see the Geologists' Association <http://www.geologists.org.uk/downloads/GAfieldworkcode.pdf>; also refer to *Basic Geological mapping* by Barnes & Lilse, from this geological field guide series). Regarding the equipment required to complete the fieldwork, again it is most important to be as fully prepared as possible. Some background preparation and thought to your fieldwork will help you to be best ready to deal with the variety of rock outcrops, scales of observation required and variety of geology and give you the best opportunity to flourish in the field.

Basic and detailed geological fieldwork should go hand in hand and so the kit list for the prepared field scientist should include the following (see Figure 2.1):

- General and detailed maps of the area of interest. This can also include, where available, ariel photographs, satellite data and even geophysical maps which can help with regional interpretations.

- Notebooks (plenty to complete your task and MAKE SURE they have your name and contact details in case you mislay one), graph paper, stereonet and tracing paper are useful in case you need to make detailed base maps and wish to plot structural data, and so on. You may also set up your own logging templates to record field data (see Section 2.3).
- A compass clinometer (check that it is set correctly for your position on the planet – information about magnetic declination can be found on most maps), a GPS hand receiver (again this must be checked for the correct datum and grid referencing system). It is vitally important that you record the settings of both your compass clino and your GPS into every notebook so that data can be corrected or converted later if need be.
- A hand lens (a magnifying glass can also be useful), pen knife, fine pencils for initial mapping, fine mapping pens (various colours), coloured pencils, rulers, calculator, grain size and fieldwork tables chart (a variety of small cards with basic info on grain size, sorting and other key geological charts are available – and you can always construct your own).
- Hammer, tampered chisels, a trowel (in good working order in unconsolidated material, for example, pyroclastics). Variety of measuring tapes, string and
- A hard hat, high visibility jacket/vest, protective glasses, first aid kit.

We have tried to make this list as detailed as possible to cover most eventualities. In many cases you would not have all of this on your person in one go, but to have the above available at short notice, for example a base camp or field centre accommodation, then you will be able to fully adapt to what is required for your fieldwork.

2.2 Preparing Maps and Basic Mapping

A detailed description of geological mapping is beyond the scope of this field guide and one should refer to the specific mapping guide in this book series. However, it is vitally important that the locality and general relationships of the igneous rocks you are studying is recorded on the appropriate maps. This would also include mapping out the distribution of the rock units at a variety of scales, and understanding how to refer key units between maps, logs and notebooks. When mapping out igneous terrains a standard scale would be to use 1 : 10 000 (metric) scale GREEN base maps (having contours, roads and other basic map ornaments in green help emphasise the geology you map onto them), but you may also be required to map and record information at different scales to capture the variations that are present in outcrop. It is also important to be aware of some of the key symbols that you may require when recording structures of igneous rocks in the field such as bedding, contacts, joints, and so on

(see *Basic Geological Mapping by Barnes & Lilse; The Mapping of Geological Structures by McClay*, from this geological field guide series).

2.3 Notebooks and Data Recording

When preliminary preparations have been completed, the task of working with igneous rocks in the field begins in earnest. At this stage, it is tempting to rush up to an outcrop in order to identify rock-types immediately. However, a more cautious approach is to be recommended.

An important part of any fieldwork is reconnaissance, and 10–15 minutes spent walking round your outcrop and visually taking in its variety and possible key features, is a far better use of time than to try and immediately take vigorous notes at an outcrop. Remember to ‘recce’ your rocks every time as this allows you to gather your thoughts and prepare to record the information properly.

You should develop a clear structure to your notebooks. It is vitally important to separate observations from interpretation as well as to record information as accurately and quantitatively as possible. Record your location (this should be done using an eight figure grid reference – appropriate GPS position), the date, the weather and your current mood including a basic statement of your plan.

2.3.1 Field sketches

It is often good to start looking at an outcrop from a larger scale and then work into the detail, so that all of the key observations can be made. You can start this by making a general outcrop sketch to which the detailed observations can be added (for example, Figure 2.2). *Field Sketches* of key outcrops and features are important as they help to illustrate the processes and structures which have taken place to form the igneous rocks. Every sketch MUST have a scale and orientation (direction or field of view), and should be annotated as extensively as possible. Also record any samples and photos taken (see examples Figures 2.2 and 2.3).

Due to the complex 3D nature of many outcrops, discontinuous exposure and the relationships you are trying to get across, it may be best to use a *synoptic sketch*. These ‘cartoon’ like sketches can be very helpful to summarise the salient features of your outcrop and can be used to help explain processes, key relationships and time sequences pertinent to your outcrop. Additionally sketch geological/outcrop maps can also help put everything into context (for example, Figures 2.4 and 2.5).

2.3.2 Logging

A powerful way of recording sequences of rock is to produce a log up through the sequence. A graphic log can be used to record any banded, layered or bedded sequence of rocks and can be used both in plutonic and volcanic sequences. This can be done using sketch logs in your notebook (for example,

FIELD SKILLS AND OUTCROP STRUCTURES

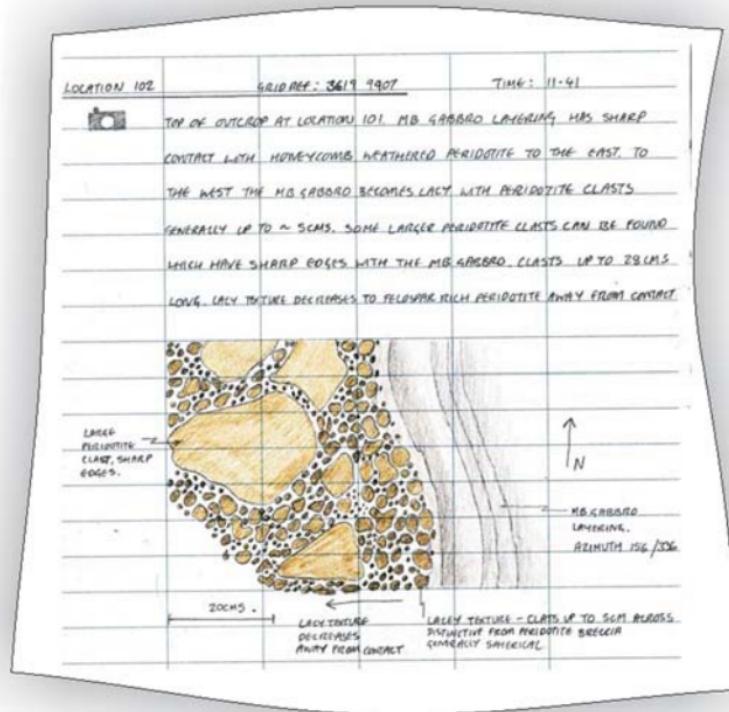


Figure 2.2 Example of field notebook layout (example from Lorraine Field, Isle of Rum, Scotland).

Figure 2.6) or more formally on specifically designed graphic logging paper. Volcaniclastic sequences particularly lend themselves to graphic logging (for example, Figure 2.7), but also lava sequences, layered igneous rocks (see Chapter 8) and even igneous/metamorphic lithostratigraphic relationships can all be recorded using some form of graphic log.

2.4 Primary Outcrop Structures

When considering an outcrop there are a variety of structures that can be recorded by annotated sketches, measurements and notes. A number of primary, late stage and secondary structures are likely to be present in any outcrop. Here we consider primary outcrop structures before looking at late stage and secondary structures.

FIELD SKILLS AND OUTCROP STRUCTURES

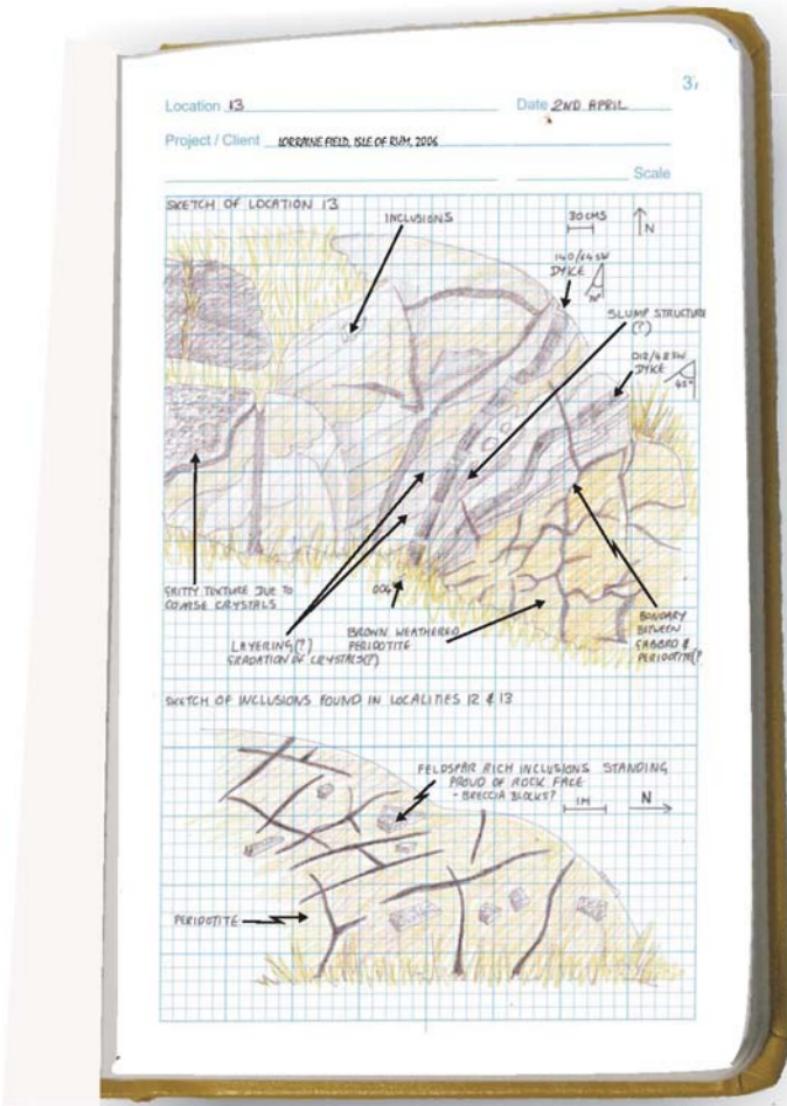


Figure 2.3 Example of field notebook with sketches of outcrops (example from Lorraine Field, Isle of Rum, Scotland).

FIELD SKILLS AND OUTCROP STRUCTURES

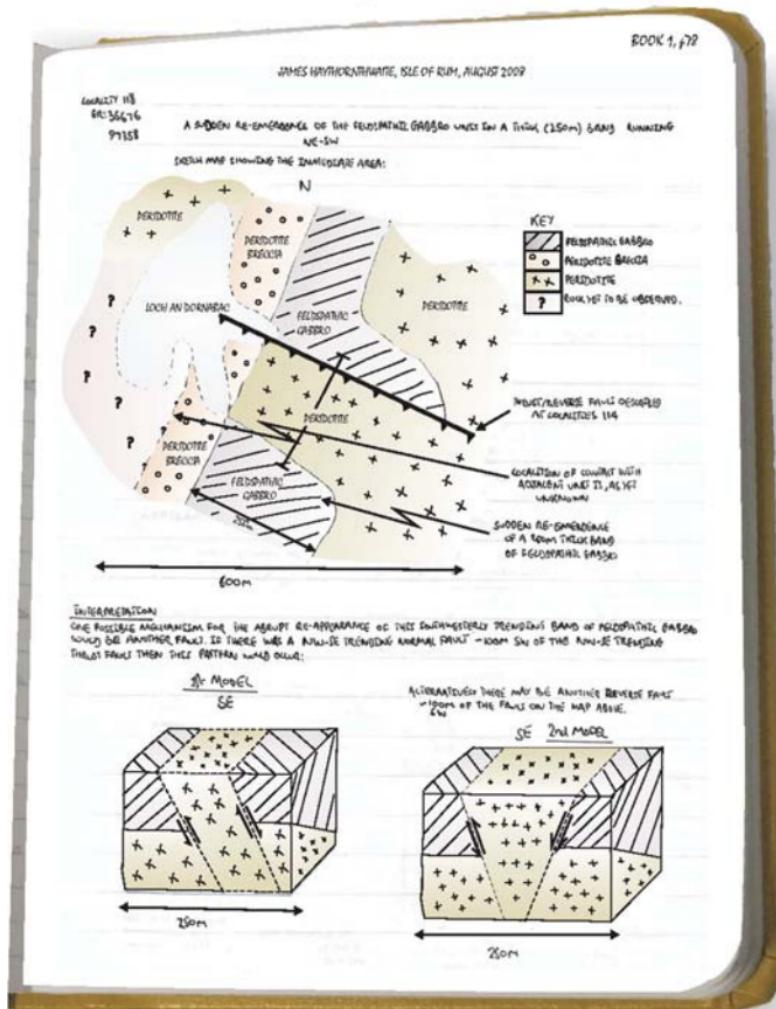


Figure 2.4 Example of field notebook with sketch map and synoptic sketches of interpretation (example from James Haythornthwaite, Isle of Rum, Scotland).

2.4.1 Bedding and depositional structures in volcanic rocks

In volcaniclastic/pyroclastic rocks and lava sequences, bedding and depositional features are important to record (Figure 2.8). With bedding the nature of contacts (for example, sharp – gradational), and the style of the bedding can be recorded (for example, graded bedding, reverse grading, cross bedded, and so on).

FIELD SKILLS AND OUTCROP STRUCTURES

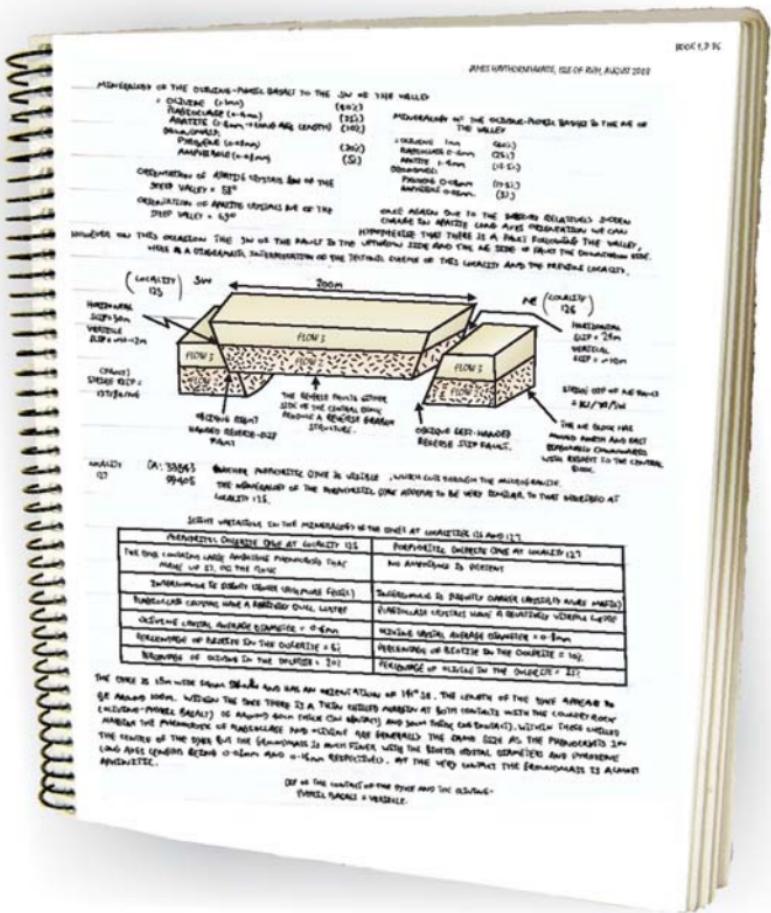


Figure 2.5 Example of field notebook with detailed notes and synoptic sketch (example from James Haythornthwaite, Isle of Rum, Scotland).

Detailed log sections (Figure 2.7) will be very useful in recording these kinds of primary structures. Very important are observations and measurements that can relate to the depositional environment, for example flow indicators, any welding of the deposit indicating hot deposition, re-working and erosive bases to beds, and so on. The nature of structures in volcanic units will be dealt with in detail in Chapters 4 and 5, and further additional information on the nature of sedimentary structures and how to record them is covered in detail in (Tucker (2011) this book series).

FIELD SKILLS AND OUTCROP STRUCTURES

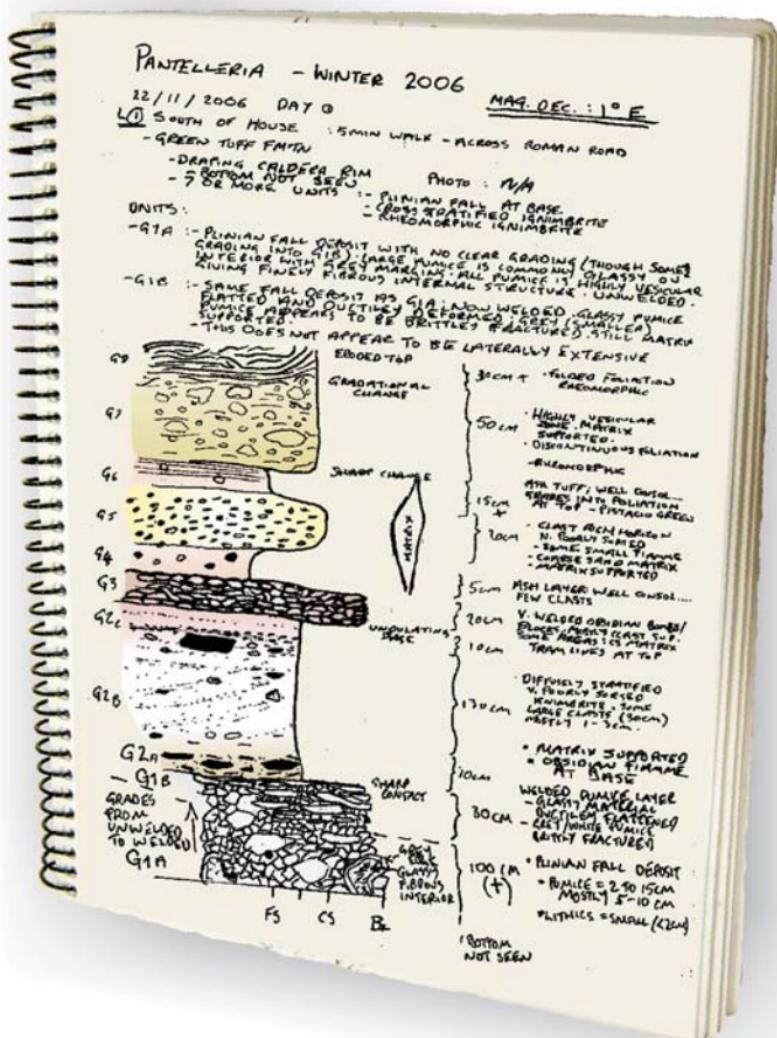


Figure 2.6 Example of sketch log (example from Richard Walker, Tenerife).

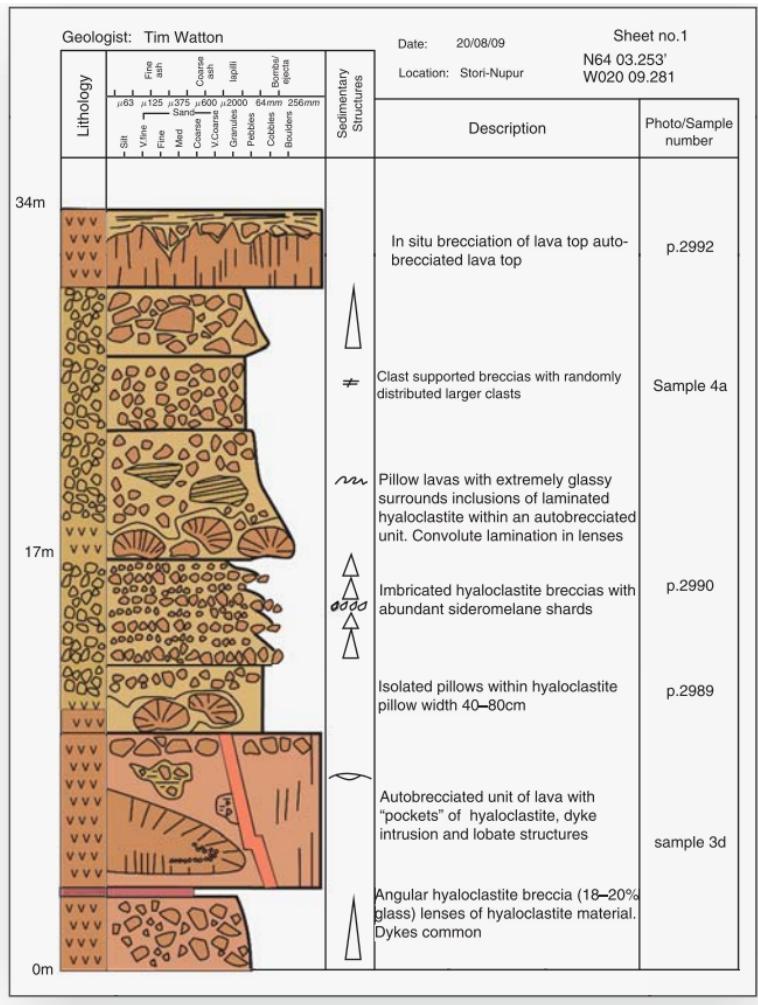


Figure 2.7 Example of detailed log on logging paper (example from Tim Watton, Iceland).

2.4.2 Igneous layering, banding and laminar structures in intrusions

These features, again easily recognised at outcrop scale, may originate when a magma is partially crystallised and consists of both crystals and melt. Banding may also result from preferential vesiculation (that is, the development of gas bubbles) along flow lines of a lava or shallow dykes. For example, flow-banded



Figure 2.8 Students looking at pyroclastic deposits on Tenerife.

ryholite lavas (Figure 2.9) with parallel bands of slightly different colour, on a centimetre or millimetre scale, result from the streaking out of glassy, pumicious or partially devitrified layers during movements of viscous lava. Parallel alignment of early-formed minerals in less viscous lavas and shallow intrusions of intermediate and mafic compositions may also be expressed as banding parallel to the direction of movement, and it is typical of feldspars in some types of andesites and basalts (see also Chapter 6).

On a larger scale, intrusions of all kinds may develop flow structures imparted either by migration of magma from its source to its zone of emplacement, or by subsequent stirring/convection within a chamber before the magma consolidated. In granitic rocks the parallel alignment of large well-shaped feldspar phenocrysts, sometimes termed megacrysts, is particularly common near the margins of intrusions where flow-aligned inclusions may also occur (Figure 2.10). It will be self-evident that, where preferred orientation of minerals or inclusions occurs in an intrusion, several faces of an exposure must be studied at the same outcrop in order to gain a full three-dimensional picture. Note that care should be taken to distinguish flow banding in coarse-grained igneous rocks from gneissose banding, developed through metamorphism.

A different kind of mineral layering, not simply related to magma flow, is commonly seen in ultra-mafic to mafic rocks (for example, peridotites and



Figure 2.9 Flow banding in glassy rhyolite, Obsidian Dome, USA.



Figure 2.10 Flow alignment of megacrysts and layering in granite, Cape Granite, South Africa.



Figure 2.11 Igneous layering between darker and lighter bands in the Rum layered intrusion, Scotland.

gabbros) and also occasionally in diorite and granite intrusions. It takes the form of alternating and repeated layers of crystals organised into different modal concentrations, typically varying from mafic rich (dark layers) to felsic rich (light layers) (for example, Figure 2.11). This igneous layering/lamination is repeated many times and can occur on a variety of scales through an intrusion, and is typically found in the larger igneous centres and thick sill sequences, where the injecting magma is a mixture of crystals and liquid. This process of large-scale *fractional crystallisation* may result in an apparently stratified magma chamber in closed systems (one magma batch) in which there is a complete gradation from the most mafic rocks (that is, mainly ferromagnesian minerals) at the base to the most felsic rocks (that is, mainly feldspars with relatively few ferromagnesian minerals) at the top. In open systems where repeated magma batches replenish the system, large scale repeated layered sequences can occur.

The origin of the igneous layering has been the subject of much debate over the years. Clearly, many examples of layers that are formed by gravity driven processes occur, with either crystals settling out of a cooling magma body or crystal laden slurries collapsing in from the sides and roof of a magma body where crystallisation is enhanced. Both of these examples will result in the preferential sorting of different density minerals into the layers. Structures such

as slumps, flame structures and graded bedding are testament to these examples (see Chapter 8). Other types of layering are formed by localised fractionation *in situ* rather than by large scale settling, and also layering formed by the wholesale replacement of material due to percolating fluids through a crystal mush can be responsible for a variety of complicated igneous laminations. Examples of igneous layering and lamination will be dealt with in more detail in Chapters 7 and 8.

2.4.3 Inclusions in host rocks

During the ascent of magma bodies to their level of emplacement or extrusion, fragments of solid rock material from the surrounding region or from the walls and roof of the magma chamber may become incorporated and remain as foreign material in the solidified igneous rock. These foreign bodies are known as xenoliths and they comprise two main types: (i) those with relict meta-sedimentary textures and (ii) those with igneous textures.

1. Xenoliths in the first category originated as fragments of country rock that were either engulfed by the upward-moving magma or fell into a magma chamber before crystallisation became complete. Undisturbed sedimentary rock fabrics, such as stratification, may be preserved within these typically angular inclusions but, more often, these xenoliths become severely metamorphosed by the heat of the magma, producing finely-banded hornfelsic textures. Xenoliths in lava flows erupted onto continental crust are particularly useful for studies of crust and upper mantle structure, as they may have sampled most of the layers through which they have passed (for example, upper mantle peridotite, high-grade metamorphic rocks of the lower crust, shales and limestones of the upper crust). Country rock xenoliths engulfed by intrusive magmas often show evidence of reaction with the magma. At its most advanced stage, this results in partial assimilation of the xenoliths, leaving rounded dark-coloured inclusions, rich in recrystallised biotite and amphibole, with apparently igneous textures.
2. Xenoliths of igneous origin: these range from the olivine and peridotite xenoliths, often called ‘nODULES’, that are common in volcanic rocks to fragments of pre-existing igneous bodies or of early-crystallised igneous material that became incorporated into the residual magma from the walls of a magma chamber (for example, Figure 2.12). All igneous-textured xenoliths derived from a source in common with their host magma may be referred to as cognate xenoliths; the term autolith is appropriate for igneous xenoliths which represent earlier-crystallised igneous material from the same magma. Generally, however, these terms have genetic implications and should be used with caution until a full petrological and petrogenetic analysis of the igneous body has been completed.



Figure 2.12 Igneous xenoliths in granitic melt, Sierra Nevada, USA.

Where two different magmas come together in a juvenile (molten) state a great variety of textures can result where there are inclusions of one magma type within another. These can be angular to rounded, wispy and irregular and show a variety of stages of mixing and mingling together. This process is of particular importance and will be dealt with in detail in Chapter 9.

2.5 Secondary or Late Stage Outcrop Structures

2.5.1 Joints, faults and shear zones

Joints are usually the most obvious features of an igneous rock outcrop where there are several sub-parallel sets of cracks or fractures, varying from a few centimetres to a few metres apart, known as *joints*. The origins of joints in intrusive igneous bodies are complex, but may be attributed to (i) the contraction that occurs when a volume of magma crystallises, producing joints that may bear a simple geometrical relationship to the intrusion walls and which may, therefore, be useful in deducing the shape of the body; (ii) post-intrusive regional tectonic stress, producing joints that usually extend beyond the intrusive body and may be matched to joint patterns in the surrounding country rocks; and (iii) expansion due to the removal of the confining pressure of overburden. This will widen pre-existing joint sets, and possibly increase their numerical density,

or may create new sets with different orientations. Joints in minor intrusions and extrusive igneous bodies that have not been buried significantly and have not experienced deformation or faulting will usually relate simply to the tension created by cooling and contraction (see Chapter 4 for details).

Faults are surfaces, sometimes utilising pre-existing joints, upon which relative motion of opposite sides has occurred, and may not be easily identified in homogeneous igneous rocks where the two displaced sides cannot be ‘matched’. However, the presence of *slickensides* (grooves or striations on the sides of a fault, parallel to the direction of movement) and of crushed or fragmented brecciated rock along the fault should be noted, as both are evidence of faulting. Such brecciation may result in formation of a cemented *fault-breccia* and in some cases, especially where softer rocks are involved, the rocks may be crushed to a soft, uncemented clay-like material, called a *fault-gouge*.

If deformation occurs when the igneous rocks are still hot or partially molten it can result in *shear zones*. These are characterised by deformation structures that are typical of ductile deformation. In such zones a careful investigation of the fabrics present can help to elucidate the direction of motion through the use of *shear sense indicators* (see the mapping of geological structures by McClay in this book series for detailed examples of shear indicators). Contacts between major batches of magma in multiple intrusions may be characterised by sheared contacts, also the deformation of igneous rocks in deeper crustal settings will result in such ductile deformation, for example in subduction zone settings and collision zones.

2.5.2 Veins, aplites and pegmatites

Veins are sheet-like, or tabular bodies that occur within both igneous and country rocks. The formation of veins and associated structures are usually due to late stages in magmatic activity, the cooling of magma bodies and the development of hydrothermal systems associated with igneous rocks. Here we shall briefly look at the main examples that we find in igneous terrains.

2.5.2.1 Hydrothermal veins

Hydrothermal veins are commonly white and usually of quartz or calcite. They vary in width from a few millimetres to a metre or more and are very common in igneous terrains. The mineral constituents may be distinguished as quartz or calcite using the hardness and cleavage tests described later (Chapter 3). Such vein materials may be precipitated either in pre-existing joints or in newly developed joints or faults. If the igneous host is cold, the veins will tend to be straight sided, whereas reaction between vein material and hot, recently-crystallised igneous rocks results in veins with irregular and diffuse edges. Occasionally, complex veins may be found. These contain suites of different minerals arranged in bands with a definite sequence from the walls to the centre of the vein. The variations

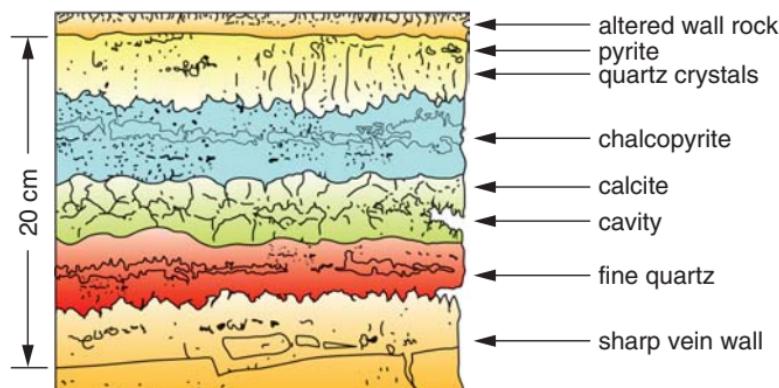


Figure 2.13 Sketch cross-section across a complex hydrothermal mineral vein.

in mineral content reflect the changing species that were being precipitated, as time progressed, from the hot fluids permeating through the vein. A field sketch of a complex mineral vein, as in Figure 2.13, is an excellent way of recording the fine detail. Hydrothermal fluids permeating through joints/fractures are also often responsible for the alteration of the host rocks on either side, involving, for example chemical reactions that break down feldspars to softer minerals such as clays. Hydrothermal systems and their associated vein networks can be very important in mineral exploration, particularly of sulfides and PGEs (see also Chapter 6).

2.5.2.2 Aplites

Aplites are veins of fine-grained felsic (that is, generally light-coloured), crystalline materials, usually of a paler colour (white or pinkish) than their host (Figure 2.14). Aplites probably represent a residual fraction of relatively silica-rich magma which remains within the intrusion when the bulk of the magma has crystallised. If this magma is then injected into the solidified parts of the intrusion, it will cool rapidly and form the fine-grained veins dominated by quartz and feldspar crystals seen in aplites. Aplites are particularly common in granitoid intrusions where, to judge from their lack of chilled margins, they are usually intruded into hot solidified material.

2.5.2.3 Pegmatites

Pegmatites are veins of exceptionally coarse-grained, igneous-textured material (Figure 2.15), usually incorporating the same minerals as their hosts, that is quartz, feldspars, micas and occasionally amphiboles in granitic pegmatites. Some may incorporate occasional large crystals of exotic minerals such as apatite, tourmaline and beryl (see Table 3.4). Like aplites, pegmatites are thought



Figure 2.14 Example of aplite dykes cutting through granite, Sierra Nevada, USA.



Figure 2.15 Coarse pegmatitic segregation in gabbros from the Isle of Skye, Scotland.

to represent the residual melt of a crystallised intrusion, but, unlike aplites, the silicate liquids that form pegmatites are often rich in water and other volatiles (for example, fluorine and boron) and rare trace elements (hence the exotic minerals noted). These volatiles are responsible for depressing the freezing temperature of the pegmatite liquids, so allowing the characteristic large crystals to form (these range from 1 cm to as much as several metres in length). Simple, homogeneous igneous-textured pegmatites without mineral zonation are termed simple pegmatites; those with mineral zonation along the length of the vein, and often with exotic minerals, are complex pegmatites.

Hydrothermal quartz veins, aplites and pegmatites are commonly intimately associated, sometimes occupying the same complex fracture with multiple infillings and sometimes cutting across one another. They are found most frequently in granite intrusions because the magmas that form these rocks are usually rich in the volatiles that concentrate in residual liquids and enhance their mobility. It should be noted that they can also be found in mafic intrusions, where they have been used to find exotic magmatic crystals for dating (for example, baddeleyite and zircon). It will be clear that both aplites and pegmatites may extend beyond the boundaries of their host intrusion as veins cutting country rocks. This means that the veins may be sheet-like intrusions which are discordant to the pre-existing fabric of their surrounding rocks. Such intrusions are termed *dykes* and, where they are about a metre or so in width, they might be termed an *aplite dyke* or *pegmatite dyke*. (NB In cross-section, a dyke is a straight-sided vein. The more common forms of dyke are described together with other minor intrusions in Chapter 6.)

2.6 Outcrop Contact Relationships

Most intrusions emplaced into hot country rocks will show little sign of marginal chilling, but the contact may have undergone plastic deformation, it may be diffuse due to remelting across the boundary, or it may be disrupted if magma has penetrated into thermally metamorphosed wall rocks. In such cases, the material on either side of the contact will be crystalline, although the contact may be recognised through contrasts of grain size, colour and mineralogy. In examples where multiple intrusions have occurred, the lack of chilled margins may result when later intrusions occur when the earlier ones are still hot.

At the opposite extreme, intrusions emplaced into cooler country rocks at a high level within the crust frequently exhibit chilled margins (Figure 2.16) of rapidly cooled magma, ranging from a few millimetres to a few metres in width. The nature of the contact and the width of the chilled margin both depend on the temperature difference between an intrusive magma and its host which, in turn, reflects factors such as the original depth of magma emplacement and the composition of the magma (since basic magmas crystallise at higher temperatures than acid magmas).



Figure 2.16 Fine chilled glassy margin of a dyke cutting through a coarse grained sill intrusion, Skye, Scotland.

In general, when describing a contact zone involving igneous rocks, measure the orientation of the contact in two dimensions, and, where possible, in three dimensions, to help you trace the feature from one outcrop to the next. Make notes on the orientation, rock-types and geometry of veins, aplites, pegmatites and chilled margins, all of which are closely related to physical processes occurring at or near the contact of an igneous body with its host. Note also any other features of contact zones, such as the occurrence of country rock inclusions (xenoliths) and contact metamorphism in the surrounding rocks. Finally, if an outcrop contains cross-cutting sets of joints, veins, dykes and other contacts, then it can be immensely useful to establish a sequence of geological events by examining their relationships (see examples in Chapter 6).

2.7 Summary of Igneous Outcrop Descriptions

The check list in Table 2.1 can be used as a guide to some of the key things to be looking out for and to record when first describing outcrop scale features. It requires the broader knowledge of detail contained within the rest of this book and should be used in conjunction with the check list for describing igneous textures at the end of Chapter 3 (Table 3.6).

FIELD SKILLS AND OUTCROP STRUCTURES

Table 2.1 Summary of key features to look for in the basic descriptions of igneous outcrops (to be used in conjunction with Table 3.6).

Plutonic rocks

Look out for any internal boundaries/contacts that denote different phases of intrusion and determine their relative ages (cross cutting relationships – chilled margins). Emphasise contact relationships. For example, concordant or discordant/transgressive. Use foliations, fabrics, the presence of xenoliths or roof pendants, plus the state of disturbance of the country rocks as some of the criteria used to establish the mode of emplacement of the pluton. For example, stoping, diapirism, forceful vs. passive intrusion.

If internal modal variations and layering is present look for features that enable you to establish the mode of formation of the layering, for example cumulative settling vs. in-situ growth, and look for way-up criteria. Use graphic logs to document the internal variation.

Use the presence/absence of miarolitic cavities (crystal lined irregular cavities or vugs), pegmatites, and so on to infer the depth of emplacement of the pluton.

Record and measure any jointing, foliation, fabric, grain-size variation, flow-banding.

Shallow intrusions/hypabyssal rocks

Measure the thickness, attitude and orientation of all intrusive sheets.

Orientation data for dykes can be plotted on rose diagrams. Cumulative dyke thickness can be used to assess levels of crustal extension.

Emphasise contact relations. Many bodies are transgressive and or varying orientation such that they are best referred to as ‘sheets’ in the field.

Record internal structures (see above). Look for evidence of processes such as flow-differentiation. Look for chilled margins: glassy chilled margins.

Volcanic sequences

Lava sequences and volcaniclastic/pyroclastic sequences lend themselves to graphic logging.

With lavas record the number of flows, their internal structure and composition, their thicknesses and the presence of any inter-lava features such as boles (paleo-sol horizons) that denote significant time gaps between flows and ash horizons, signifying more explosive activity.

(continued overleaf)

FIELD SKILLS AND OUTCROP STRUCTURES

Table 2.1 (*continued*)

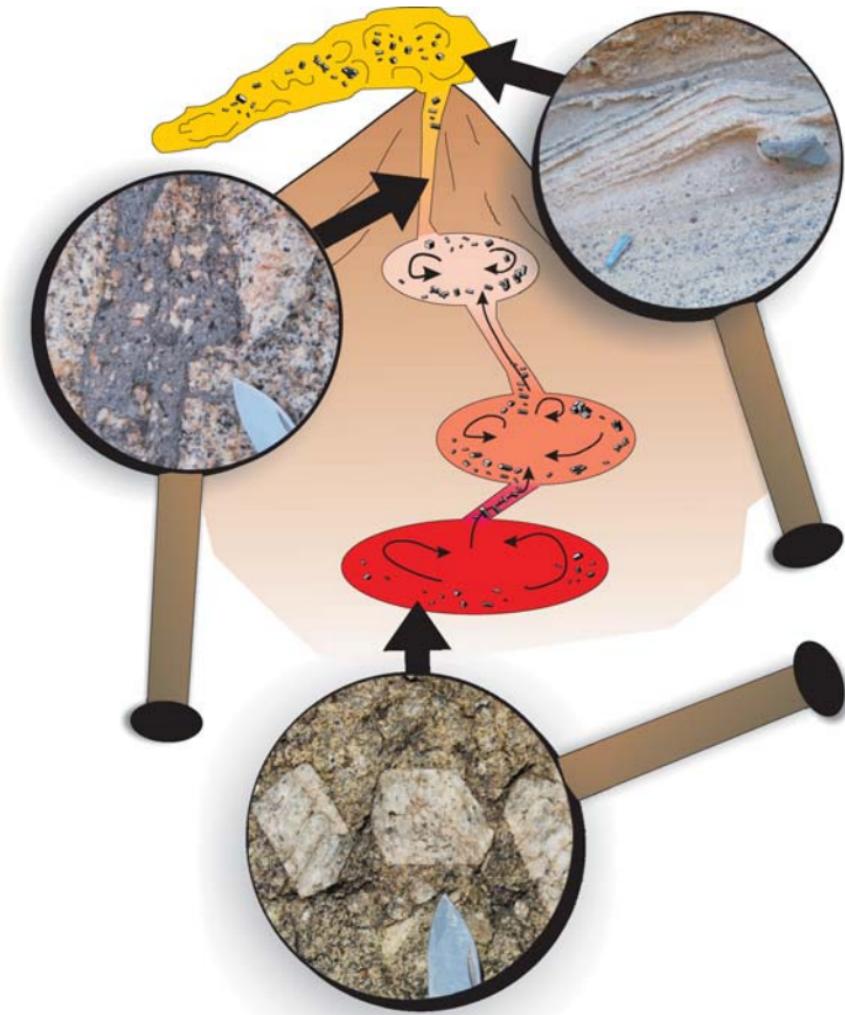
Look for features that establish the environment of eruption, for example subaerial vs. sub-marine. Try to characterise the flow types, aa versus, pahoehoe, directions and flow regimes plus the source of eruption, for example fissure or vent.

Field identification of rock type may be difficult. For example, brecciated lavas may resemble agglomerates.

In volcaniclastic/pyroclastic sequences pay careful attention to proportions of lithics, pumice, fiamme, crystals. Is the deposit air-fall or water-lain in origin? Is there evidence for reworking?

Deposits erupted on to water can settle into well bedded ash layers. Grain-size sorting is often enhanced between top and bottom due to the effects of water and pumice may be reversely graded due to temporary floatation before settling. Look for water-escape structures vs. fumarolic activity.

IGNEOUS TEXTURES AND CLASSIFICATION



Understanding igneous textures and classification from source to surface.

3

IGNEOUS TEXTURES AND CLASSIFICATION

When describing an outcrop or sequence of rocks that are in part or wholly made up of igneous rocks it will be necessary to make a detailed examination of the rock-type(s) with a view to visiting other outcrops, mapping the shape and form of the igneous rocks and considering their mode of occurrence (for example, intrusive, extrusive, pyroclastic, and so on). It will also be an important step to start to classify the types of rock present in terms of their mineral contents and approximate position on accepted igneous classification schemes. In this chapter we introduce the basic methods used to describe and classify igneous rocks in field outcrop/hand specimen.

NOTE: This is directed at identifying the primary igneous components and is tailored specifically to textural and mineral characteristics of the rocks they formed from, that is magmas of specific composition, so will be of particular use to intrusive and volcanic rocks containing visible crystals. This would equate to the primary magmas (juvenile component) of pyroclastic rocks, however the textural relationships such as grain size, sorting and other key features of pyroclastic/volcaniclastic rocks will be dealt with in Chapter 5.

3.1 Describing Rock Types

To what extent can igneous rocks be named/classified in the field? The business of assigning field names to fresh samples may be surprisingly straightforward, because igneous rocks vary widely in their physical characteristics. Even when making a detailed study of a small region in which, for example successive lava flows may appear remarkably similar at first sight, it is possible with practice and by adopting a systematic approach to develop a keen eye for small changes in the appearance of your samples. The main features of igneous rock samples that should be noted in the field as an aid to identification are:

1. **Colour**, as a preliminary guide to the approximate chemical and mineral composition of the sample.
2. **Texture, grain size and fabric** which are related to the rate and sequence of crystal growth in intrusive and extrusive lava flows, and to eruption and emplacement in pyroclastic/volcaniclastic rocks.

3. **Mineralogy** – usually between two and four main minerals occur in an igneous rock. For coarse-grained rocks, the number and nature of the main minerals present, their relative abundances and inter-relationships provide a clear guide to rock identification. Many fine-grained rocks also contain some large crystals that may be useful for diagnostic purposes. In many cases of fine-grained rock identification, it is difficult to make full use of features 1–3 and to record in the field a description, that is more detailed than say, ‘fine-grained grey lava’. In addition, but not part of the rock identification, there are inclusions of ‘foreign’ material. These may be found in samples and should be described separately, as they are important when assessing the mode of occurrence and origin of the igneous rock body.
4. **Weathering** – although the diagnostic features of igneous rocks are often less easily recognised from weathered than from freshly broken surfaces, the opposite is sometimes true for coarse-grained mafic rocks. It is important to note the amount of weathering and to explore both weathered and fresh outcrop as they often reveal different things about the rocks texture. Bear in mind, however, that the effects of weathering may penetrate below the surface to a depth varying from a few millimetres in fine-grained rocks to as much as several metres in the case of coarse-grained or heavily-fissured material and in the case of poorly consolidated pyroclastic rocks. (These depths may be increased by at least a factor of 10 in areas which have, or are presently experiencing, wet, tropical climates.)

3.2 Colour and Composition

The basic composition of an igneous rock is reflected in the mineralogy which crystallises from it. It is important, therefore, to remember some key things with regard to understanding the types of minerals that we expect to see in igneous rocks of different compositions, as these reflect known crystallisation sequences. Additionally, it can be helpful in determining likely mineral species where minerals which are easy to identify are found with minerals that are not clear or difficult to identify, or when you find minerals that should not be in sequence with each other which may represent magma mixing/crystals in disequilibrium with melt compositions, and so on. The most important conceptual diagram which will help understand crystallisation sequences is Bowens Reaction Series (Figure 3.1), based on experimental melting experiments of different compositions of igneous material. Importantly it includes the major rock types from mafic through intermediate to felsic (acidic) rocks, it also indicates the likely temperatures that different compositions have been at.

Once you have an understanding of what minerals crystallise from what magmas types, we can use this information to get a good understanding of the

Bowen's Reaction Series

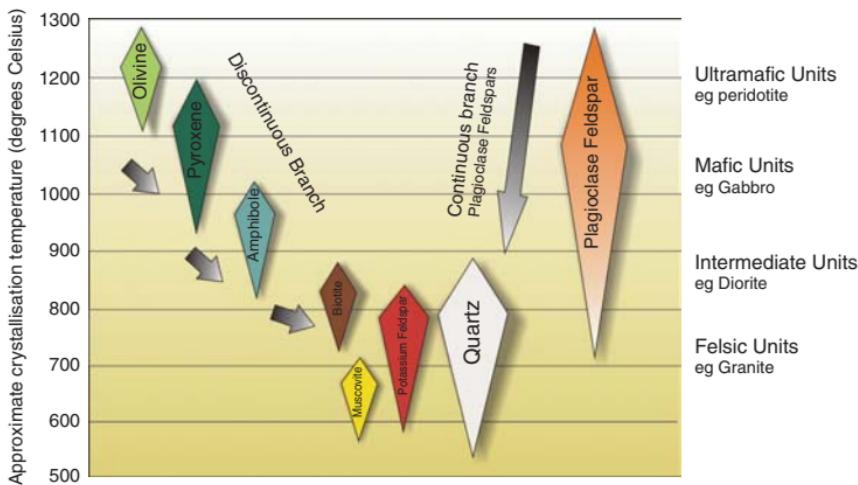


Figure 3.1 Bowens reactions series and the relation of mineral associations with composition and temperature of igneous rocks.

basic compositions we may expect from different mineral assemblages. The naming of most field specimens requires a working knowledge of the natural variations in their silicate mineral compositions (Section 3.5). Individual silicate minerals have chemical compositions that vary within predictable limits (Table 3.1) and it follows that the mineral and chemical compositions of rocks are closely related. This relationship is also expressed in the most obvious property of a field specimen – its *colour* – which is reflected in the colours of its constituent minerals.

The following terms may be used to describe the characteristics of individual minerals and the major minerals present within the mode or norm of an igneous rock.

1. Terms applied to minerals present in the *mode* of an igneous rock and to rocks having one or more of these minerals as major components of the mode:
 - a. *felsic* = feldspar + silica (for example, feldspars, feldspathoids for example, leucite + nepheline, quartz),
 - b. *mafic* = magnesium + ferric (for example, olivine, pyroxenes),
 - c. Mafic minerals may also be termed *ferromagnesian* minerals (ferric + magnesium).

IGNEOUS TEXTURES AND CLASSIFICATION

2. Terms applied to minerals present in the *norm* of an igneous rock and to rocks having one or more of these minerals as major components of the norm:
- d. *salic* = silicon + aluminium (for example, quartz, feldspars, feldspathoids),
 - e. *femic* = ferric + magnesium (for example, olivine, pyroxenes).

Although most of your observations will relate to fresh surfaces it is usually worth noting the colour of the weathered skin of an igneous rock, particularly if it is fine-grained. Since the chemical process of weathering tends to concentrate medium- to dark-brown insoluble residues in rocks that contain a high proportion of mafic minerals (Table 3.1), or pale-brown to white residues if most of the minerals are felsic, the colour of the weathered surface provides a first clue to the composition of the rock. Fresh surfaces of most igneous rocks range in

Table 3.1 Typical chemical compositions¹ for some major silicate minerals in igneous rocks (weight per cent). See Table 3.4 for distinguishing features in hand specimens of these and other minerals.

Mineral	SiO ₂	Al ₂ O ₃	FeO + Fe ₂ O ₃	MgO	CaO	Na ₂ O	K ₂ O	H ₂ O
<i>Felsic minerals</i>								
Quartz	100	—	—	—	—	—	—	—
Orthoclase	65	18	—	—	—	—	17	—
Albite	69	19	—	—	—	12	—	—
Anorthite	43	37	—	—	20	—	—	—
Muscovite	45	38	—	—	—	—	12	5
Nepheline	42	36	—	—	—	22	—	—
<i>Mafic minerals</i>								
Olivine	40	—	15	45	—	—	—	—
Clinopyroxene (Augite)	52	3	10	16	19	—	—	—
Orthopyroxene (Fe rich)	50	1	30	16	1	—	—	—
Amphibole (Hornblende)	42	10	21	12	11	1	1	2
Biotite	40	11	16	18	—	—	11	4

¹Silicate minerals are made up of atomic frameworks in which different combinations of cation-forming elements are always bonded to oxygen and so it is customary to quote analyses in terms of simple oxides rather than elements.

Note also that in most mineral groups there is compositional diversity caused by the ability of atomic frameworks to hold different cations in the same site (e.g. substitution between Fe and Mg, or between Na and Ca occurs in many mineral groups).

colour from dark- to light-grey, greenish-grey or brown, though many of the lighter-coloured rock types also have purple to pink varieties. In simple terms, this dark to light colour range results from the different combinations of dark and light minerals that crystallise from magmas of different compositions (further details in Section 3.5).

After noting the colour of the weathered surface, the next step in classifying an igneous field specimen is to decide where the fresh surface lies within the natural colour range and, if possible, to record its *colour index*, based on your estimate of the total percentage of mafic minerals present (see Figure 3.2), as follows:

1. Felsic/leucocratic (0–33%) – rich in felsic minerals and therefore light-coloured,
2. Intermediate/mesocratic (34–66%),
3. Mafic/melanocratic (67–100%) – rich in mafic minerals and therefore dark in colour.

The accuracy with which the colour index can be estimated will depend on the grain size of the specimen. For example, close inspection under a hand lens of the fresh surface of a coarse-grained rock should give an estimate within 10% of the true value (for example, Figure 3.3), but for fine-grained rocks, the colour index becomes more difficult to determine and is obviously inapplicable to glassy rocks (Section 3.3). In cases where individual crystals cannot be distinguished with a hand lens, it is better to record an overall impression of the specimen's colour, thus avoiding any misleading conclusions that could arise later from a poorly estimated numerical value of colour index.

When working in an area of fairly similar coarse-grained rocks, such as a granite intrusion where the colour index might vary from, say, 5 to 25%,

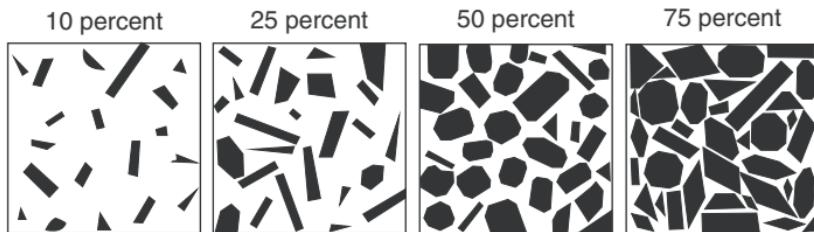


Figure 3.2 Comparison diagram for estimating the mafic mineral content of two-dimensional igneous rock surfaces. It is well worthwhile using this diagram in the field as there is a common tendency, because of the greater visual impact of dark minerals, for colour index to be overestimated in field reports.



Figure 3.3 Three samples of equigranular (equal grain sizes) coarse-grained igneous rocks with colour indices of 10, 40 and 70. The left specimen is felsic/leucocratic, that in the middle is intermediate/mesocratic, and that on the right is mafic/melanocratic.

then it is permitted field practice to use the prefixes *leuco-*, *meso-* and *mela-* in conjunction with the word ‘granite’ to describe, respectively, samples with mafic mineral contents that are below average, average and above average for the whole intrusion.

Because the colour of an igneous rock reflects the minerals it contains and hence its chemical composition, it follows that the colour index conveys some information about the chemical composition of the rock. It will be obvious from Table 3.1 that the most important chemical constituent of most igneous rocks, expressed in the conventional oxide form, is silica – SiO_2 . Almost all igneous rocks contain between 40 and 75% SiO_2 by weight; the higher the silica content, the more felsic minerals the rock must contain and the lower will be its colour index – and vice versa for its content of mafic minerals.

There are obvious similarities between this colour index classification and the much used geochemical classification of igneous rocks into *acid*, *intermediate* and *basic* categories, using their silica contents. This classification dates from the period when silicate minerals were classified, erroneously, as salts of hypothetical silicic acids. Although the terms acid, intermediate, basic and ultrabasic are defined strictly in terms of the abundance of SiO_2 in igneous rock chemical analyses (Table 3.1), they are also applied loosely in the field; for example most felsic/leucocratic rocks may be described as acid. Table 3.2 provides a ready comparison of the different terms used in describing the colour of an igneous rock; in general, it is best to use those on the right for field descriptions.

Although the implications of colour for an assessment of rock composition have been described in some detail, clearly there are many other clues to be considered before assigning field names to igneous rock samples. Therefore, do not jump to hasty conclusions but, after noting the colour of the weathered and

Table 3.2 General rules.

Geochemical term	Definition wt%	Approximate range of colour index ¹	Possible field descriptions
Acid	>65	5–25	Felsic/leucocratic
Intermediate	52–65	25–55	Intermediate/mesocratic
Basic	45–52	55–85	Mafic/melanocratic
Ultrabasic	<45	85–100	—

¹Note that there is a general correlation, in which felsic/salic rocks are leucocratic, with low colour index and of acid chemical composition, whereas mafic/femic rocks are melanocratic, with high colour index and of basic/ultrabasic chemical composition. However, there are exceptions to this generalisation. For example, rocks composed of Ca-rich plagioclase, termed *anorthosites*, are felsic/salic leucocratic rocks with low colour index but of *basic* or intermediate chemical composition. Similarly, rocks composed of Na-, Ca-, Mg- carbonates, termed *carbonatites*, are leucocratic rocks with low colour index and of *ultrabasic* chemical composition.

fresh surfaces, consider next the other features of field specimens – their texture and mineral content.

3.3 Texture, Grain-Size/Shape and Fabric

Collectively, the relationships between the constituent crystals and glass (where present) in an igneous rock are known as *texture*. When describing the texture of a field sample, examine:

1. The *grain-size* of the rock which can reflect the rate at which it crystallised.
2. The *fabric*, or geometrical characteristics and arrangement of the crystals including, where possible, observations of the number of minerals present and the characteristic shapes, or *habits*, of their crystals.
3. The overall *homogeneity* of the specimen (that is, whether it is uniform and equigranular, or contains mineral segregations, banding and irregular inclusions).

These features all provide clues to the physical conditions under which the magma crystallised; thus, whereas the colour of a specimen is generally related to its chemical composition (for example, granites are usually lighter in colour than gabbros), the texture reflects its history and mode of occurrence. For example, rapid chilling leads to the formation of a glass, whereas the nucleation of crystals at many sites in a magma erupted at the surface and still subject to fast cooling,

IGNEOUS TEXTURES AND CLASSIFICATION

Table 3.3 Grain-size descriptions in crystalline rocks.

Fine-grained ¹	Few crystal boundaries distinguishable in the field or with the aid of a handlens; mean grain size below 1mm. If the rock is glassy, the term <i>hyaline</i> may be used
Medium-grained ²	Most crystal boundaries distinguishable with the aid of a hand lens; mean grain size 1–5 mm
Coarse-grained ²	Virtually all crystal boundaries distinguishable with the naked eye; mean grain size greater than 5 mm

¹Fine-grained and hyaline rocks may be termed *aphanitic* in texture.

²Medium- and coarse-grained rocks may be termed *phaneritic* in texture.

such as in lava flow, tends to result in minute grains that usually cannot be identified in the field. On the other hand, slow cooling, as in intrusions emplaced below the surface, is accompanied by crystallisation at relatively few nucleation sites and results in larger, easily observed crystals.

Grain size is usually the most obvious textural characteristic, and after examining the specimen with the aid of a hand lens, it should be assigned to one of the categories in Table 3.3.

The terms in Table 3.3 are easily applied to rocks that are more or less *equigranular* (equal grain sizes), but many specimens are porphyritic (see Figure 3.4), having a population of early-formed, large *phenocrysts* in a finer grained *groundmass* or *matrix*. In such cases, the actual size of both populations of crystals should be recorded. Note that another short-hand way of describing porphyritic texture, once the phenocryst mineral has been identified, is to use the hyphenated adjective ‘-phyric’ along with the mineral name: thus Figure 3.4d shows a *feldsparphyric* granite. But it may prove impossible to continue the description of non-porphyritic (or *aphyric*) specimens in this category without recourse to laboratory techniques. Field names such as ‘fine-grained felsic or mafic rock’ are recommended on the basis that the former are likely to contain mainly feldspars whereas the latter will contain mainly mafic minerals.

The next stage in describing medium and coarse-grained rocks is to examine their *fabric*, that is the *shape* of the crystals and their *relationship* to one another. First compare and contrast the characteristics of different types of crystals in the rock. Minerals left to crystallise ‘freely’ in a magma will form well-shaped crystals which, in the two dimensions of a field sample surface, would show regular, straight-sided cross-sections, often of rectangular shape (but see Table 3.4 for common crystal habits). During the crystallisation of most igneous rocks several minerals are competing for the available space and only the first-formed crystals, such as any phenocrysts present, will have good crystal forms. Three

IGNEOUS TEXTURES AND CLASSIFICATION

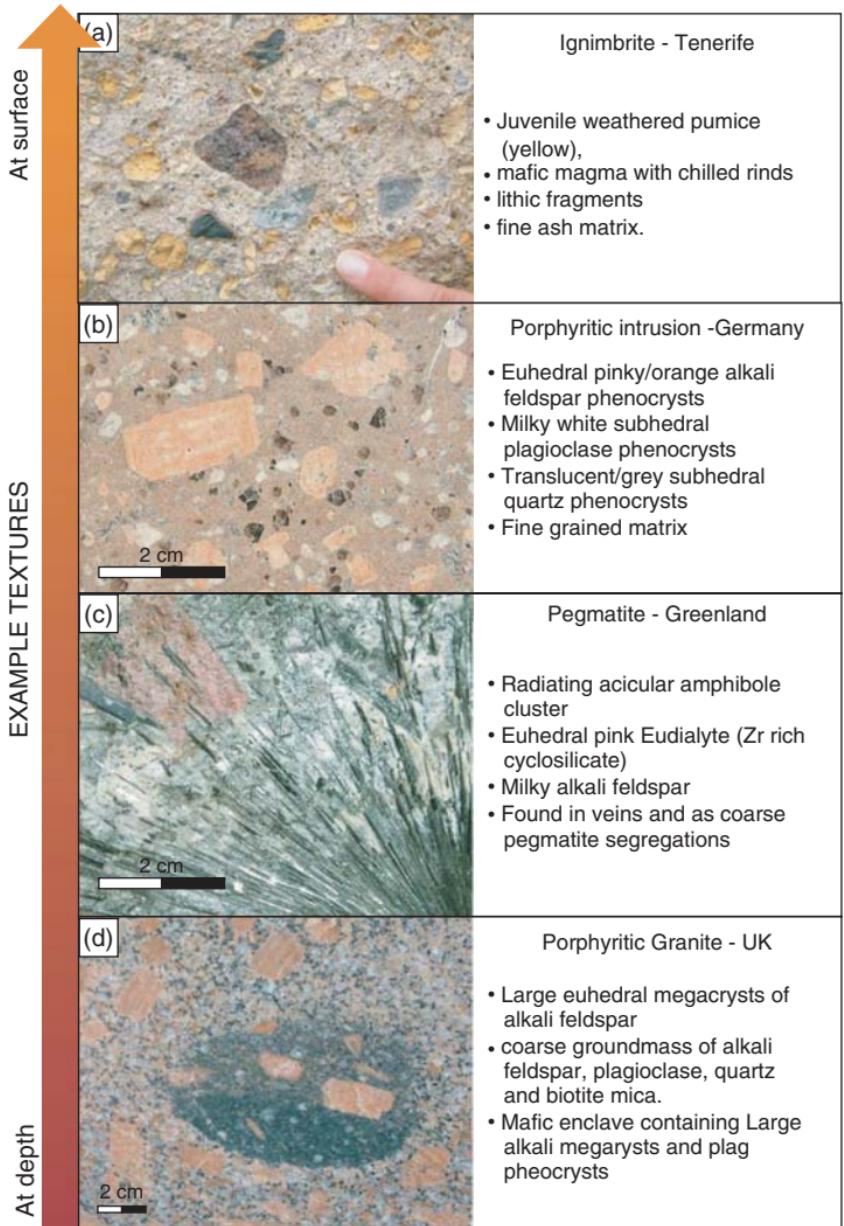


Figure 3.4 Examples of textures in some evolved rock types. Note: these textural types for example, porphyritic (large crystals in finer ground mass) and pegmatite and so on. Are found throughout the range of magma compositions.

IGNEOUS TEXTURES AND CLASSIFICATION

The figure consists of three diagrams labeled (a), (b), and (c). Diagram (a) shows a single euhedral crystal with a regular octahedron shape, bounded by its characteristic faces, with a 1 cm scale bar below it. Diagram (b) shows a subhedral crystal with irregular, rounded edges, partially bounded by its faces. Diagram (c) shows an anhedral crystal with no clearly defined faces, appearing as a cluster of irregular shapes.

Term	Euhedral	Subhedral	Anhedral
Other terms	(Idiomorphic, Automorphic)	(Hypidiomorphic, Hypautomorphic)	(Allotriomorphic, Xenomorphic)
Description	Crystal completely bounded by its characteristic faces	Crystal bounded by only some of its characteristic faces	Crystal lacks any of its characteristic faces

Figure 3.5 Euhedral, subhedral, anhedral with descriptions. Sketches of (a) euhedral, (b) subhedral and (c) anhedral amphibole crystals (cleavage traces at 60° and 120° , cf. Table 3.6).

sets of terms are used to describe the quality of development of crystal faces, *Euhedral*, *Subhedral* and *Anhedral* (see Figure 3.5).

In ideal circumstances, where the grain shapes of a coarse-grained rock are extremely clear, it may be possible to classify the grain shape fabric of the entire rock (for example, *granular texture* if all crystals are of uniform size). Furthermore, a rock may not fit neatly into a single category: thus, one in which <50% of the crystals are euhedral and <50% anhedral might best be described as having a mixed euhedral and anhedral granular texture. In some examples it is possible to note the relative timing of crystallisation, buy careful examination of the crystal population with a hand lense on both weathered and fresh surfaces. Euhedral crystals often grow first when they are relatively free in the melt, and subsequent crystallising phases have sub to anhedral shapes as they are forced to occupy the space between existing euhedral crystals (for example, Figure 3.6a). Where a new crystallising phase engulfs crystals of a pre-existing growing phase this leads to a texture known as *poikilitic*, with the large *oikocrysts* engulfing the smaller *chadacrysts* (for example, Figure 3.6b).

After describing the individual crystals and their inter-relationships, the remaining textural characteristics to notice concern the overall *homogeneity* of the rock. At the scale of a hand specimen, most igneous rocks are fairly

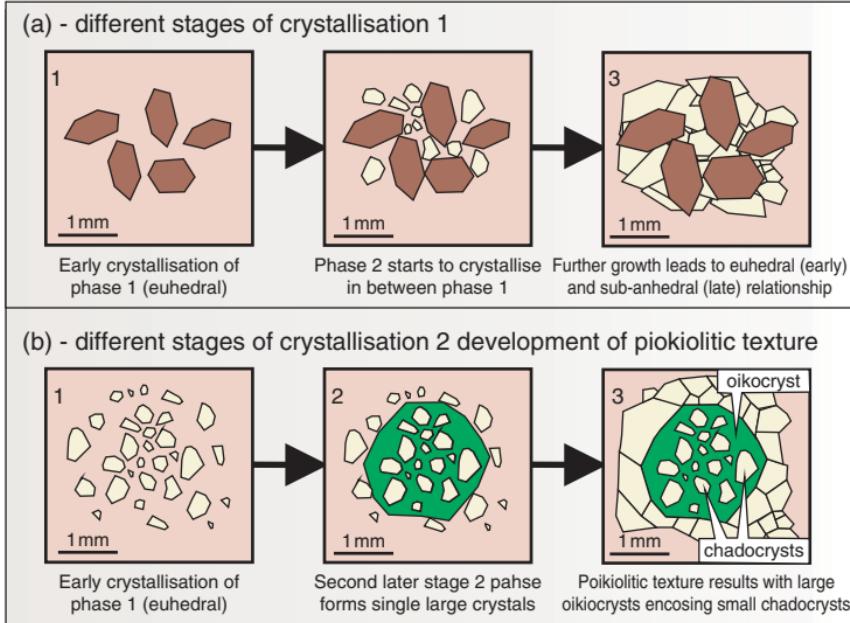


Figure 3.6 Schematic texture development to show: (a) Different stages of crystallisation shown by euhedral-anhedral relationships. (b) Different stages of crystallisation with large crystals enclosing older small crystals.

homogeneous but occasionally *flow banding* or *mineral lamination* are encountered (Figure 3.7). Flow banding in medium and coarse-grained rocks may be expressed through the visible alignment of elongate crystals (for example Chapter 2; Figure 2.10) or, in fine-grained rocks, by a fine-ribbing or colour banding (flow banding) which is often most noticeable on weathered surfaces (for example, Chapter 2; Figure 2.9). As flow structures are developed parallel to the direction of movement, measurement of their orientation will act as a useful aid in mapping lava sequences. Mineral lamination, expressed as alternating lighter and darker layers in coarse-grained rocks, may originate through fractionation processes (as described in Chapter 2 and Figure 2.11) or, perhaps, through high-grade metamorphism of otherwise homogeneous igneous material. Subsequent interpretation of field data in all these cases is made a great deal easier if careful notes are made in the field of the colour, texture, orientation and scale of the layering, not only in hand samples, but also in the outcrop as a whole.

The textures which have been introduced in this section are those frequently encountered in igneous field studies; the list will serve as a general guide in

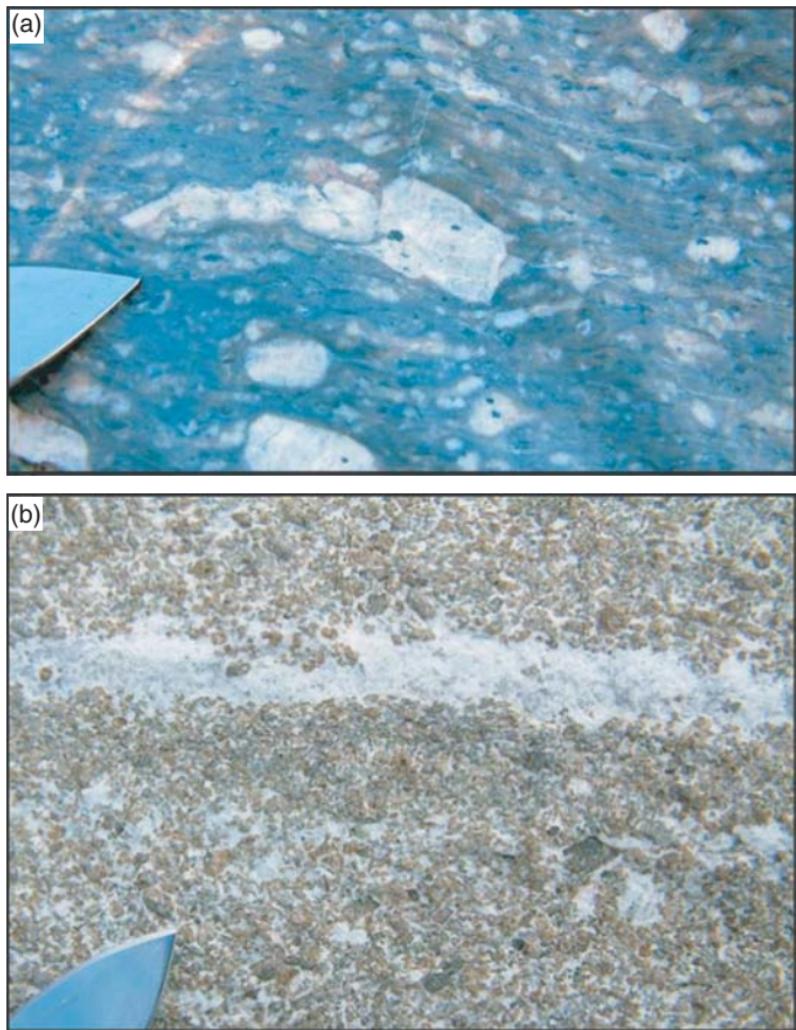


Figure 3.7 Examples of: (a) Mineral alignment in porphyritic dyke, Antarctica. (b) Mineral lamination/layering in the Dias intrusion, Dry Valleys, Antarctica (see also Figure 2.11).

the field but will not cover all occurrences. Even if you do not recognise the textural features of a particular hand specimen, remember that the golden rule is to *make accurate observations backed up by field sketches as appropriate*. More examples of these and other less frequently observed textures at the scale of hand specimens appear later in this field guide.

3.4 Mineral Identification

The most informative, yet also the most difficult, part of a field description of igneous rocks is the identification of the constituent minerals. Igneous rocks consist of abundant or *major* minerals, together with minor minerals that are much less abundant. Major and minor minerals may be termed *essential* and *accessory* minerals. Essential minerals are those that are necessary to the naming of the rock. They are usually major minerals, for example quartz in a granite, but are sometimes present in minor amounts, for example olivine, in an olivine basalt. Accessory minerals are those which are present in such small amounts that they are disregarded in the definition of the rock; for example Fe-Ti oxide in a gabbro, or biotite in a granite.

Even though most rocks contain only two or three essential minerals, together with a few accessory minerals, *it is rarely easy to identify all the essential minerals and often impossible to name the accessory minerals*. The problem is most acute for fine-grained mafic rocks in which many of the mafic minerals have a similar appearance. Even in coarse-grained leucocratic rocks, it can be difficult to distinguish among the different minerals (for example, between plagioclase and alkali feldspar). However, as you gain in experience of the properties of minerals and of the types that commonly occur together, the more confident your naming of rocks will become. It is worth remembering that most igneous rocks represent an assemblage of minerals that crystallised almost in chemical equilibrium with one another from a magma. So, for example the mafic minerals olivine and pyroxene *or* biotite and amphibole frequently occur together but other combinations, such as olivine and biotite, are much less common. The kinds of combinations that occur most frequently in igneous rocks are summarised in the mineral-rock classification diagrams in Section 3.5.

In the course of observing the colour and texture of a rock you will have formed an impression of the approximate number and some of properties of the individual minerals. The next stage is to identify some of those minerals with the aid of Table 3.4, which summarises mineral properties under five major headings: colour, cleavage, lustre, habit and hardness. Generally, it is best to start with the light-coloured felsic minerals, as these are often most easily identified. A few points about the use of Table 3.4:

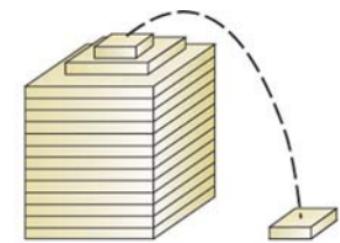
- Colour** is an obvious property of a mineral which, like that of a rock (Section 3.2), reflects its overall chemical composition. Generally, alkali/alkaline earth silicates (Na, K, Ca silicates) are light-coloured whereas silicate minerals rich in transition elements, particularly iron, are dark-coloured.
- Cleavage** is the tendency of minerals to split along well-defined planes that are related to weaknesses in their atomic structures. Cleavage surfaces are flat planes that reflect light evenly and are often responsible for lustre

(see below); for example freshly-broken feldspar crystals may show slightly jagged edges which, under a hand lens, are seen to be a composite of many flat cleavage surfaces intersecting at or near 90° (Figure 3.8). The perfect single cleavage of mica, which forms the brightly reflecting surfaces, often allows small flakes to be dislodged with a penknife or finger nail. In contrast, cleavage in pyroxenes and amphiboles is seen only occasionally as fine striations within the grains, whereas the characteristic cross-sections of euhedral crystals (eight-sided and six-sided lozenge shapes, respectively, for pyroxenes and amphiboles, Table 3.4) are much more diagnostic.

3. **Lustre** describes the reflective properties of a mineral and is assessed by turning the specimen until the mineral surfaces, particularly any cleavage surfaces, are caught by the light. Lustre varies from *dull* (non-reflecting), through *resinous* and *silky* to *bright*, which may be metallic, glassy or vitreous (which literally also means glassy but here is defined slightly differently) depending on the mineral concerned. A brightly coloured silicate mineral which is opaque to the transmission of light may have a *vitreous* lustre, like glazed pottery, such as clean crystals of amphibole and pyroxene. On the other hand, a similar translucent (weakly transparent) or transparent mineral may appear *glassy* (for example, quartz and fresh olivine).
4. **Habit** refers to the characteristic morphology of euhedral crystals. Of course, most crystals in igneous rocks are not euhedral and, given the random mixture of cross-sections through three-dimensional mineral grains represented in a hand-specimen, many grains may need to be examined before a characteristic shape is found (for example, pyroxene and amphibole habits are distinctive only when viewed close to one particular planar cross-section). Some minerals have quite well-defined characteristic shapes in most euhedral cross-sections, for example lath-shaped feldspars which are long and thin, but rectangular, dodecahedral or trapezohedral garnets and acicular (that is, needle-like) tourmalines. Others, such as quartz and olivine, are less commonly euhedral. *Twinning* results from crystal growth in different orientations on either side of a common atomic structural plane. It can be visible in large crystals of feldspar, where it provides the only practical way of distinguishing alkali feldspar (simple twins only) from plagioclase (simple and multiple twins) in hand specimen (see Table 3.4). Twinning is best observed by rotating cleavage or crystal faces in sunlight; individual twins are seen to have slightly different angles for optimum reflection and, at the hand specimen scale, are commonly seen only in alkali feldspars.
5. **Hardness**, a property related to the strength and uniformity of atomic bonding within a crystal, is described using a simple but non-linear scale of increasing hardness from 1 to 10 known as *Mohs' scale* (Figure 3.9). If the mineral grains are large and fresh enough (weathered grains are often softer than the original), then a series of scratching tests using a finger nail (hardness circa 2.5),

a copper coin (hardness circa 3.5) and a steel penknife blade or geological hammer (hardness circa 5.5) may prove diagnostic. For example, mica or calcite will scratch easily with a coin or penknife; feldspars may scratch a penknife blade with difficulty, but quartz will always scratch it easily.

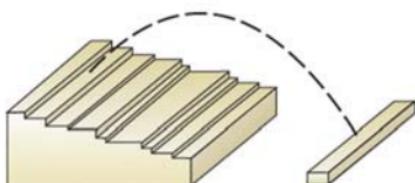
Table 3.4 contains sufficient information for good examples of the 10 most common minerals that occur in coarse-grained igneous rocks to be identified. Also included are eight of the more abundant accessory minerals whose



One direction - basal



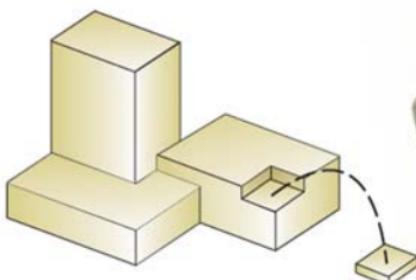
example - mica



Two directions - prismatic



example - calcite



Three directions - cubic



example - galena

Figure 3.8 Examples of different types of cleavage.

Table 3.4 Mineral properties in igneous rock hand specimens.

Mineral	Typical chemical formula	Colour	Cleavage	Lustre	Habit	Hardness
Felsic minerals						
Quartz	SiO_2	Colourless to pale grey when surrounded by dark minerals; transparent	None; irregular or curved fracture surfaces	Glossy, shiny	Rare trigonal pyramids but usually irregular, anhedral	 7
Alkali feldspar	$(\text{K}, \text{Na})\text{AlSi}_3\text{O}_8$	White or pink, sometimes orange or yellow	Two sets at 90° poorly visible	Usually dull, sometimes silky or vitreous	Tabular crystals; shiny cleavage surfaces may show simple twins. Elongate rectangular 'laths', lamellae, or irregular masses of plagioclase may be noted, in which case the crystal is termed a perthite	 6
Plagioclase feldspar	$\text{NaAlSi}_3\text{O}_8$ to $\text{CaAl}_2\text{Si}_2\text{O}_8$	White or green, rarely pink or black	Two sets almost at 90° , poorly visible	Usually dull, sometimes silky or vitreous	Lath-shaped crystals; shiny cleavage surfaces may show multiple, parallel twins	 6–6.5
Nepheline	NaAlSiO_4	White to pale grey	Two poor cleavages, one occasionally distinct	Greasy, vitreous	Usually occurs in micro-crystalline groundmass; occasional aggregates of crystals	5.5–6

IGNEOUS TEXTURES AND CLASSIFICATION

Mafic minerals		One excellent cleavage, cleaves into thin flexible sheets	Shiny, silver and pearly	Tabular crystals sometimes 6-sided, especially in pegmatites	2–2.5
Muscovite (mica)	$KAl_2(AlSi_3O_{10})(OH)_2$	Colourless to pale brown or green	Very poor, usually fractures	Glossy when fresh, vitreous	6–7
Olivine	$(Mg,Fe)_2SiO_4$	Olive green, yellow-green, sometimes brown	Two good sets meeting at 87°/93°	Vitreous when fresh, dull when altered	6
Pyroxene	(i) $(Mg, Fe, Ca)_2Si_2O_6$ (augite, etc.) (ii) $NaFeSi_2O_6$ (aegirine)	Black to dark green or brown Yellowish-green	Two good sets meeting at 56°/124°	Vitreous when fresh, dull when altered	5–6
Amphibole	(i) $Ca_2(Mg, Fe)_5Si_8O_{22}(OH)_2$ (e.g. tremolite) (ii) $Na_2Fe_3^{2+}Fe_2^{3+}Si_8O_{22}(OH)_2$ (riebeckite)	Black to brownish black or dark green dark blue			

(continued overleaf)

IGNEOUS TEXTURES AND CLASSIFICATION

Leucite	KAlSi_2O_6	White or grey	None	Vitreous or resinous	Often euhedral trapezohedral crystals in alkaline lavas	5.5–6
Hematite	Fe_2O_3	Red to red-brown, sometimes black	None	Dull	Usually fine and powdery, occasionally scaly or fibrous crystals	5.5–6
Magnetite, (a spinel mineral)	Fe_3O_4	Black, brownish-black	Poor	Metallic, dull	Small equidimensional granular crystals, occasional cubes or octahedra	5.5
(Other spinels are $\text{M}^{2+}\text{M}_2^{3+}\text{O}_4$ where M^{2+} is Fe, Mg, Mn, etc. and M^{3+} is Al, Fe, Cr, etc. e.g. dark brown to black <i>chromite</i> , FeCr_2O_4 , which occurs in some peridotites)						
Ilmenite	FeTiO_3	Black, brownish black or grey	None	Metallic or dull	Thin plates or scales usually elongate crystals, sometimes rod-like	5.6
Monazite	$(\text{Ce}, \text{La}, \text{Th})\text{PO}_4$	Pale yellow to dark brown	Moderate single cleavage	Resinous	Thick tabular crystals in granites and gneisses	5–5.5
Secondary minerals (see also Table 7.2)						
Calcite	CaCO_3	White, translucent	Three sets rhombohedral	Vitreous, rarely glassy	Usually granular or fibrous in igneous rocks, common in veins, cavities, and so on. NB: Reacts with dilute acid	3

(continued overleaf)

IGNEOUS TEXTURES AND CLASSIFICATION

Table 3.4 (continued)

Mineral	Typical chemical formula	Colour	Cleavage	Lustre	Habit	Hardness
Zeolite group	For example, $(\text{Na}_2\text{Ca})(\text{Al}_2\text{Si}_3\text{O}_{10})_n \text{H}_2\text{O}$ (<i>n</i> varies from 2 to 8)	White, pale yellow or pale green, rarely pink, red or blue	Variable according to mineral type	Usually vitreous or silky	Massive or granular crystals lining cavities, particularly, amygdales; radiating fibrous clusters or needles	5–6
Clay group	For example, $\text{Al}_4\text{Si}_4\text{O}_{10}(\text{OH})_8$ (kaolinite)	White to pale browns and greens	Good, but not visible in hand specimens	Dull	Fine powdery aggregates replacing mainly feldspar in igneous rocks	1
Epidote	$\text{CaFe}^{3+}\text{Al}_2\text{Si}_3\text{O}_{12}(\text{OH})$	Pale yellows and apple green, rarely brown or red	One good cleavage	Vitreous	Variable, often elongated crystals, needles and radiating groups, coarsely crystalline varieties in hydrothermal veins and vesicles	6–7

IGNEOUS TEXTURES AND CLASSIFICATION

Chlorite	$(\text{Mg},\text{Al},\text{Fe})_6(\text{Si},\text{Al})_8\text{O}_{20}(\text{OH})_4$	Mid-green to dark greenish-yellow	One good cleavage gives thin sheets	Dull to pearly and 'micaceous'	Usually aggregates of fine crystals, sometimes thin tabular flakes replacing mafic minerals in igneous rocks	2–3
Pyrite	FeS_2	Brassy yellow, occasionally brown or black	Poor	Metallic; iridescent tarnish	Often good cubic crystal faces, occasionally striated. Granular aggregates, particularly along veins in igneous rocks	6–6.5
Zircon	ZrSiO_4	Colourless pale brown	Two poor	Adamantine/glassy	Small prisms with terminal faces often euhedral	7.5

IGNEOUS TEXTURES AND CLASSIFICATION

MINERAL	MOH'S RELATIVE HARDNESS	COMMON OBJECTS
Talc	1	
Gypsum	2	
Calcite	3	Finger nail (2.5)
Flourite	4	Copper coin (3.5)
Apatite	5	
Orthoclase	6	Steel nail (5.5) Glass plate (6)
Quartz	7	
Topaz	8	Streak plate (7.5)
Corundum	9	
Diamond	10	

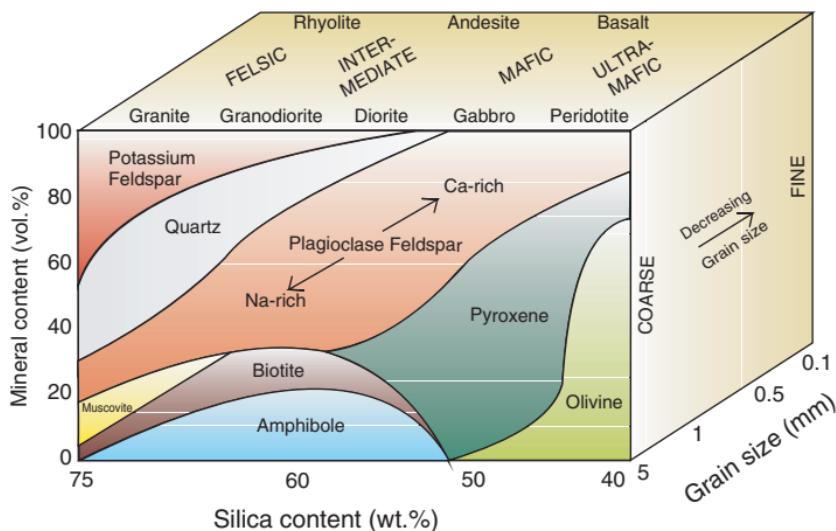
Figure 3.9 Moh's harness scale.

individual abundances rarely exceed 10% by volume of a rock. Whereas apatite, sphene (titanite), garnet and leucite sometimes form easily identified euhedral crystals, the three opaque iron-titanium oxide minerals, which often fill the interstices between major silicate minerals, have similar appearances in hand specimens. The list of secondary (that is, post-magmatic) minerals includes (i) those found as amygdale minerals in lavas (for example, calcite, zeolites); (ii) those found in hydrothermal veins cutting igneous rock outcrops (for example, calcite, chlorite, but also quartz) and (iii) those which replace primary silicates in rocks that have been subject to pervasive hydrothermal or metasomatic alteration and weathering reactions at submagmatic temperatures (for example, clay minerals and/or epidote replacing feldspars and mica; chlorite replacing mafic minerals).

3.5 Naming and Classification

When describing a medium- or coarse-grained field specimen you should estimate, as a percentage, the volume of the rock occupied by each identified mineral and, for those you are unable to identify, make a brief note of their properties for later reference. Observations of colour, texture and mineral content can now be

IGNEOUS TEXTURES AND CLASSIFICATION



Grain size	Rock Type			
	Felsic	Intermediate	Mafic	Ultramafic
Fine <0.25mm	Rhyolite	Andesite	Basalt	-
Medium 0.25–2mm	Microgranite	Microdiorite	Dolerite	-
Coarse >2mm	Granite	Diorite	Gabbro	Peridotite

Figure 3.10 Preliminary classification scheme for igneous rocks using colour index (front face shows approximate proportions of light and dark minerals that occur at different silica contents), common silicate minerals (where recognised in the field) and grain size (decreasing with depth into the diagram).

combined into a field name for the rock specimen using, initially, Figure 3.10. Earlier, it was noted that the natural chemical range of most igneous rocks from 40 to 75% SiO₂ is accompanied by a progressive increase in their content of felsic minerals and a corresponding decrease in their mafic constituents. This information appears on the front face of Figure 3.10 with the acid, silica-rich rocks on the left and the ultrabasic rocks on the extreme right.

Although this scheme for preliminary rock identification works well in the field for medium- and coarse-grained rocks, it is often possible to follow more closely the precise definitions of the IUGS Subcommission (appendix Figure A.1). This scheme classifies all rocks with a content of mafic minerals (plus Fe–Ti oxides, carbonates of igneous origin and accessory minerals) of

IGNEOUS TEXTURES AND CLASSIFICATION

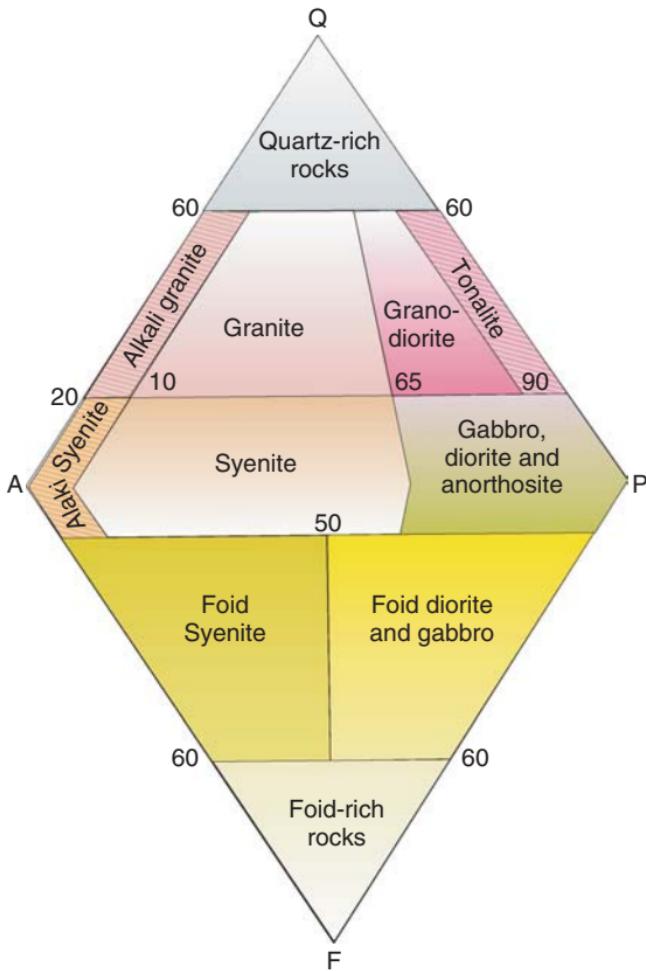


Figure 3.11 Simplified QAPF diagram for naming medium- and coarse-grained rocks in the field using their salic mineral content (see text and appendix Figure A.1 for further details). Most igneous rocks fall in the shaded area. Remember that foid is an abbreviation for feldspathoid.

less than 90%, primarily using the relative proportions of their different felsic minerals: Q = quartz, A = alkali feldspars, P = plagioclase feldspars, F = feldspathoid minerals (e.g. nepheline and leucite in Table 3.4, though there are other feldspathoids; these minerals have similar chemistry to feldspars but with less SiO₂ and are often collectively known by the abbreviation foid minerals). Figure 3.11 is a simplified version of the QAPF diagram which,

IGNEOUS TEXTURES AND CLASSIFICATION

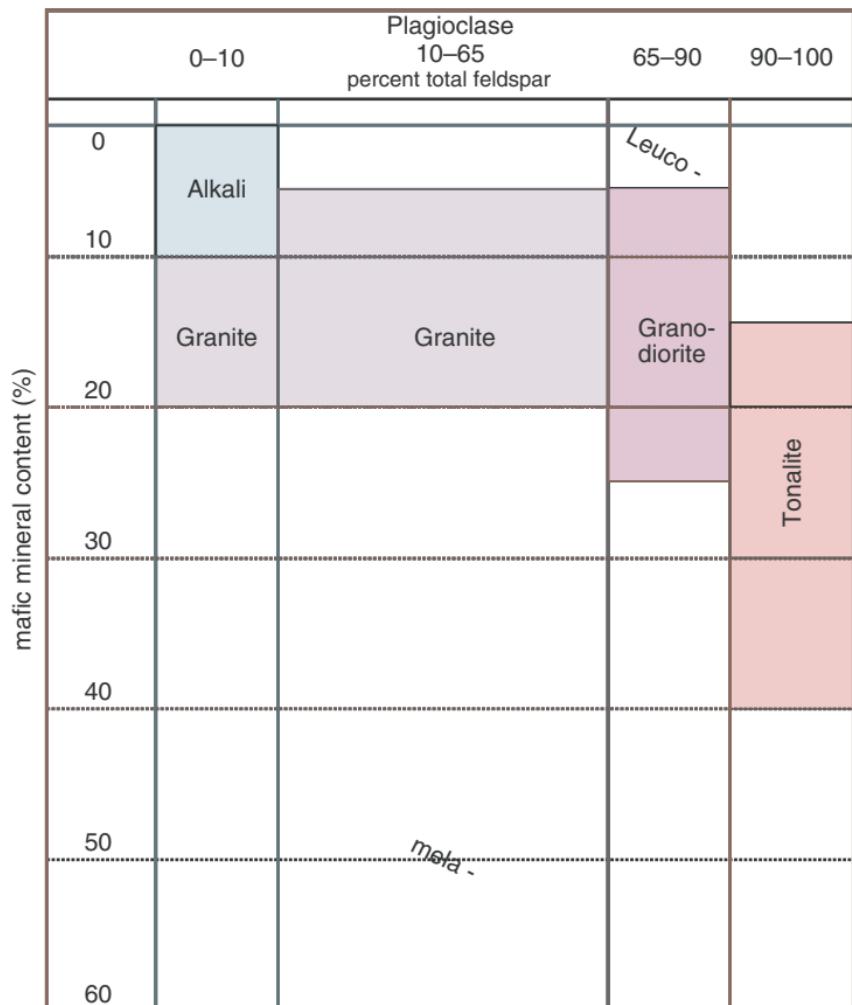


Figure 3.12 Plot of increasing content of plagioclase feldspar (left to right), as a percentage of total feldspar, against mafic mineral content for granitoid rocks, defined as those with quartz as between 20 and 60% of the salic minerals.

with experience, can be used in the field quite successfully, especially in conjunction with Figures 3.13–3.14.

Consider first those rocks where quartz is typically 20–60% of the felsic minerals: these can be named accurately as alkali granite, granite, granodiorite or tonalite, if the relative proportions of alkali feldspar and plagioclase are estimated (cf. Figure 3.12). Where this is not possible, any sample with

IGNEOUS TEXTURES AND CLASSIFICATION

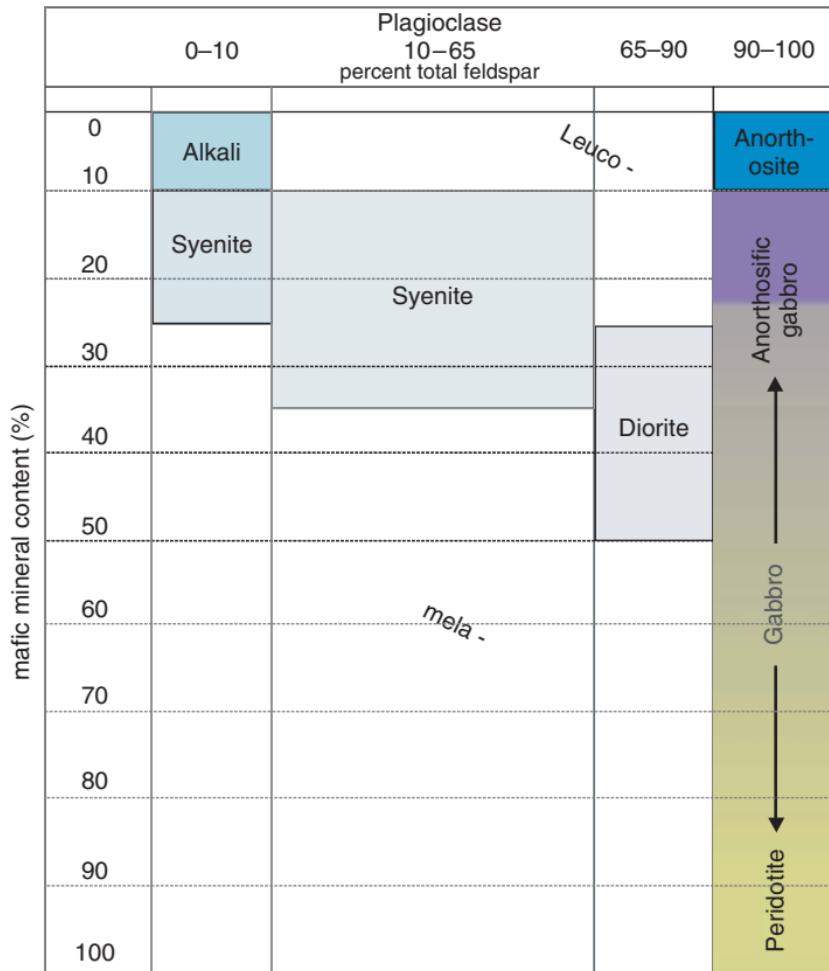


Figure 3.13 Plot of increasing content of plagioclase feldspar (left to right), as a percentage of total feldspar, against mafic mineral content for common rocks with less than 20% of the salic minerals as quartz. The difference between diorite and gabbro is discussed in the text and the names used here are defined more precisely in later chapters.

20–60% quartz may, provisionally, be termed ‘granitoid’. Figure 3.13 illustrates the usual range of mafic minerals present in granitoid rocks and the areas in which the prefixes leuco- and mela-would normally be applied. Similarly, the feldspars of rocks with lesser amounts of quartz, or no quartz and small amounts

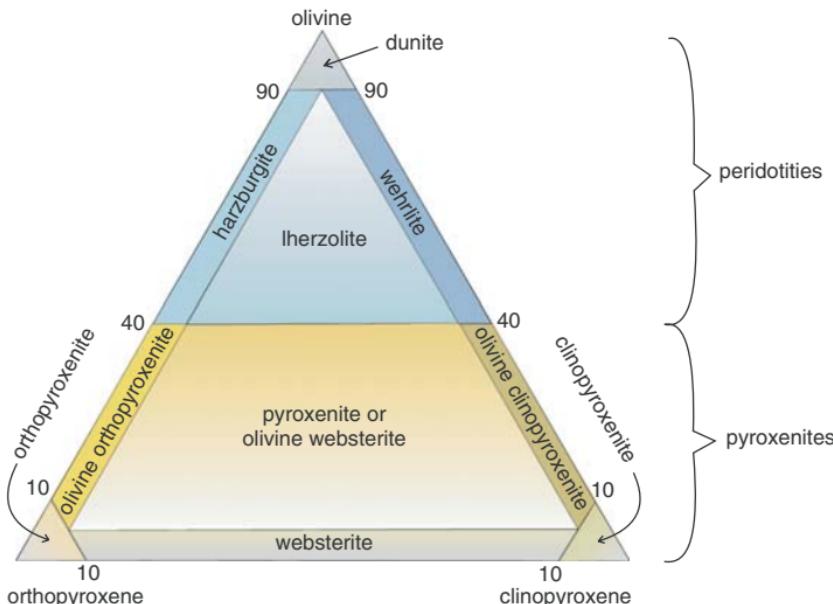


Figure 3.14 Triangular plot showing proportions of olivine, orthopyroxene and clinopyroxene for classification of ultramafic rocks (rocks with 90% or more of mafic minerals). The commonest ultramafic rocks are lherzolite and harzburgite.

of feldspathoid minerals may allow a distinction to be made between alkali syenite, syenite and an abundant group of plagioclase-rich rocks – the diorites, gabbros and a less common group discussed later, the anorthosites (compare Figures 3.11 and 3.13). Strictly, the distinction between diorite and gabbro lies in the composition of the plagioclase feldspar which must contain more than 50% of the albite ($\text{NaAlSi}_3\text{O}_8$) molecule in diorite and more than 50% of the anorthite ($\text{CaAl}_2\text{Si}_2\text{O}_8$) molecule in gabbro. As the compositions of plagioclase feldspars cannot be determined in the field, the use of a secondary characteristic is – the nature of the common mafic mineral (Figure 3.10) – to make a preliminary distinction between diorite (with biotite and amphibole) and gabbro (usually with pyroxene). It is emphasised once more that, wherever possible, these provisional field names should be re-examined in the light of subsequent microscope studies using a more detailed version of the QAPF diagram (Figure 3.11 and appendix Figure A.1).

Among the rock-types introduced so far are two major acid-intermediate-basic geochemical series, or associations, the occurrence of which tends to be mutually exclusive.

IGNEOUS TEXTURES AND CLASSIFICATION

1. The abundant granite-granodiorite (tonalite-diorite-gabbro series (cf. Figure 3.10), often known as the *calc-alkaline* association.
2. The less abundant alkali granite-alkali syenite-syenite-foid syenite (usually nepheline syenite) series, known as the *alkaline* association.

These associations have strict geochemical definitions based on chemical ratios, such as $\text{CaO}/(\text{Na}_2\text{O} + \text{K}_2\text{O})$ (see Gill, 2010), but, as will be clear from (Figure 3.11 and appendix Figure A.1), the geochemical differences are reflected in the nature of the feldspars and, less obviously, in the mafic mineral content. Rocks of the alkaline series have relatively low abundances of mafic minerals (for example, alkali syenite and syenite in Figure 3.13) and, instead, their low silica (basic) end-members extend closer to the feldspathoid apex in Figure 3.11 – nepheline is a typically abundant mineral. In contrast to the dark-browns and black colours of pyroxenes and amphiboles in diorites and gabbros, the smaller amounts of mafic minerals of alkali rocks are often distinctive, consisting of spectacular green or dark-blue alkali pyroxenes and amphiboles (for example, aegirine and riebeckite – compare formulae with common pyroxenes and amphiboles in Table 3.4). Thus a secondary characteristic distinguishing alkali granite from calc-alkaline granite, for example is often the presence of dark-blue amphibole needles in the former and of black prismatic amphiboles and/or micas in the latter.

In the low silica, ultrabasic rocks, there are two important groups to consider (cf. Figure 3.13):

1. A leucocratic group, the *anorthosites*.
2. A melanocratic group, the *peridotites* and *pyroxenites*.

Anorthosites are pale-brown or white monomineralic rocks dominated by plagioclase feldspar; varieties with more than circa 15% of mafic minerals are termed anorthositic gabbros. They are sometimes found in association with alkaline rocks such as syenites and nepheline syenites and sometimes with layered mafic-ultramafic rocks. In contrast with anorthosites, rocks that are virtually devoid of feldspar but have abundant mafic minerals (>90%), and range from dark-green or brown to black in colour, are known collectively as *ultramafic*. These are subdivided further, according to their particular combinations of mafic minerals, into pyroxenites (>60% pyroxene), peridotites (>40% olivine) and several other minor categories that are not easily recognised in the field (Figure 3.14). *Dunite*, an olive-green monomineralic rock with 90–100% olivine, is perhaps the most distinctive, but recognition of the other types depends on the separate identification of orthopyroxene and clinopyroxene which is sometimes difficult in hand specimens. Orthopyroxenes are generally of paler brown or green colours than clinopyroxenes and sometimes weather to a

red-brown colour whereas, with a few rare exceptions, clinopyroxenes are very dark brown, green or black. Green orthopyroxene is distinguished from olivine using crystal shape and cleavage (Table 3.4). However, a particularly distinctive clinopyroxene that occurs in some ultramafic rocks in bright green *chrome diopside* ($\text{CaMgSi}_2\text{O}_6$ containing a few per cent of the $\text{NaCrSi}_2\text{O}_6$ molecule).

The igneous rocks considered so far are all composed essentially of silicate minerals. However, a rather rare group of igneous rocks, known as *carbonatites*, contain more than 50% (by volume) of carbonate minerals and occur as intrusive dykes and plugs and more rarely as extrusive lavas. These are subdivided further using the major carbonate mineral present into calcite (CaCO_3)-carbonatites, known as *sövites*; dolomite ($\text{CaMg}(\text{CO}_3)_2$) carbonatites, known as *beforsites*, which tend to be cream or pale-yellow; and *ferrocarbonatites* where yellow-brown siderite (FeCO_3) is the main mineral. Although these igneous rocks are rare, they are associated with alkaline igneous rocks in many areas.

The terminology introduced so far in this section (cf. Figures 3.11–3.14) applies exclusively to coarse-grained rocks. Finer grained examples in which the requisite minerals can be identified in hand specimen should be given the same names with the prior adjectives (Table 3.3, but see also Table 3.6) ‘*medium-grained*’ or ‘*fine-grained*’ (for example, fine-grained granite). However, in the special case of fine- or medium-grained mafic intrusive rocks, the terms *dolerite* (or *diabase* in the USA) are in common use and are equally as acceptable as fine- or medium-grained gabbro.

Although most medium-grained rocks can be matched in mineral composition with the common rock-types described above, there is one group that occurs in minor intrusions which has distinctive mineralogy and texture: the *lamprophyres*. These are medium-grained mesocratic or melanocratic (occasionally ultramafic) porphyritic rocks with phenocrysts of biotite and/or amphibole in a groundmass containing feldspars and/or feldspathoid minerals. They may contain hydrothermal calcite and zeolites, and the primary igneous minerals are frequently extensively altered to these minerals. They are difficult to classify further due to the fine grain size and alteration of the groundmass. Biotite bearing examples that are found in association with feldspar are commonly termed *Minette*; Amphibole bearing examples with no feldspar *Monchiquite*. *Kimberlites* are also found in minor intrusions, pipe like bodies and erupted units (see Chapter 6) and are usually characterised by significant inclusions of mantle olivine. It is notoriously difficult to classify rocks like *Kimberlites* and *lamprophyres* in the field and often geochemical classification is required. Readers are directed to the recent detailed classification paper by Tappe *et al.* (2005) which outlines some of the specific issues associated with classifying *lamprophyres* and associated rock types.

Fine-grained rocks have an individual set of names, although for each coarse-grained rock type there is a fine-grained equivalent (Table 3.5). The distinctive

IGNEOUS TEXTURES AND CLASSIFICATION

Table 3.5 Fine-grained rock-types, their medium- and coarse-grained equivalents, and possible phenocryst minerals (see also Table 4.1).

Fine-grained rocks	Medium- and coarse-grained equivalents	Common phenocryst minerals
<i>Felsic (leucocratic)</i>		
Rhyolite	Granite	Quartz, alkali feldspar
Dacite	Granodiorite	Quartz, plagioclase feldspar
Trachyte	Syenite	Alkali feldspar, occasional mafic minerals
Phonolite	Nepheline Syenite	Alkali feldspar, nepheline or leucite, occasional green alkali pyroxene
<i>Mafic (meso- to melanocratic)</i>		
Andesite	Diorite	Plagioclase, pyroxene or amphibole, occasionally quartz or biotite
Basalt	Gabbro	Plagioclase, pyroxene, olivine

mineralogy of different porphyritic rocks in this category, together with their colour, often provides enough information for them to be named in the field; it is customary to use their phenocryst minerals as descriptive adjectives (see Table 3.5). Not all the phenocryst minerals listed in Table 3.5 need occur in each of the respective rock-types; indeed, most porphyritic rocks are named on the basis of the phenocrysts present, for example plagioclase, pyroxene-phyric andesite, olivine-phyric basalt. In cases of doubt, and particularly for aphyric specimens, the terms fine-grained felsic or mafic rock are recommended. The nomenclature of volcanic rocks is discussed in more detail in Chapter 5. Finally, Table 3.6 provides a checklist for easy reference when describing and naming igneous rocks in the field (see also Table 2.1).

Table 3.6 Checklist for describing igneous rock hand specimens in the field
(to be used in conjunction with Table 2.1 for outcrop description).

- 1 Examine weathered surface of rock outcrop, noting texture and colour: sometimes weathered surfaces provide a clue to the relative hardness of minerals and to their composition (e.g. the red-brown weathering residues of orthopyroxenes and olivines)
- 2 From the outcrop, collect and number representative sample(s) with fresh surfaces. When collecting for thin section or analytical work, trim off as much weathered material as possible in the field to save effort later and to lighten the load for carrying. Clean any waste chips away, especially from farmland, and do not leave the outcrop looking a mess
- 3 Record colour of the fresh surface and, where possible, estimate the colour index (% ferromagnesian minerals)
- 4 Record the colour of the weathered surface and any features that were not visible in the fresh surface
- 5 Examine the grains under a handlens:
 - (a) If the rock is aphyric, note any other textural features and record felsic or mafic composition (see Table 3.2)
 - (b) Record coarse, medium or fine grain size of the rock (see Table 3.3) and note textural relationship between minerals (e.g. Figures 3.5 and 3.6)
 - (c) If the rock is porphyritic, record grain size and textural relationship of phenocrysts to groundmass
- 6 Record the degree of homogeneity, the presence of layering, lamination, flow banding, vesicles and other special textural characteristics, such as the presence of inclusions
- 7 Estimate the number and proportions of the different minerals present and, for each, record where possible: colour, cleavage, lustre, habit, hardness. Use Table 3.4 for identification purposes
- 8 Use mineral associations to predict the possible presence of other minerals you can not immediately identify and in classifying the rock. Combine your observations to give the specimen a field name
- 9 Combine your observations to give the specimen a field name, using Figures 3.11–3.14 for medium- and coarse-grained rocks and Table 3.5 for porphyritic fine-grained rocks. If you are unable to be specific, at least use general/generic names, such as ‘mafic lava’

VOLCANICS 1 – LAVA FLOWS



Lava lake surface- ErtaAle, Ethiopia



Pahoehoe lava - Hawaii

Figure 4.1 Picture of modern lava flows (opposite cover) composite of sampling pahoehoe Hawaii and lava lake surface.

4

VOLCANICS 1 – LAVA FLOWS

When magma reaches the surface it erupts as volcanic material, either as effusive lava flows (see Figure 4.1) or explosively to form fragmental pyroclastic rocks. In the next two chapters we will focus on the products and deposits of volcanic activity. In this chapter we will look at the key structures and observations that can be made from lava flows. In Chapter 5, Volcanics 2; pyroclastic rocks will be the focus. It is important to remember that a mixture of lava flows and pyroclastics rocks are often found together in most volcanic settings, and there are some gradations between these two major divisions which we will touch on in both chapters.

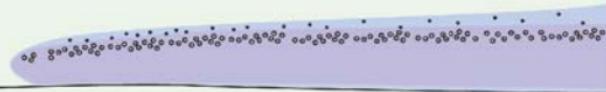
4.1 Lava Flow Emplacement Mechanisms

Before we focus on the key surface and internal structures found in lava flows, it is important to briefly consider the mechanisms that are involved in their emplacement as this will help to put some of your observations into context. Also it is important to realise that our understanding of how lava flows are emplaced has recently been much enhanced (in the late 1990s to early 2000s). The largest lava flows are predominantly basaltic in composition and form vast outpourings of lava known as flood basalts. These have similar compositions to modern lavas as on Iceland and Hawaii and it was observations on Hawaii that resulted in our understanding of the emplacement of lavas by inflation (Figure 4.2). Flows start as thin sheets of lava which quickly form an insulating crust. Further development of the lava involves inflation from inside of the flow core, which is protected from cooling by this insulating crust. By this mechanism lavas can travel long distances from source without significant cooling and crystallisation.

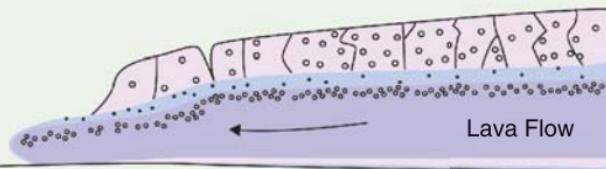
Other mechanisms for producing lava flows include the development of flows from fire fountains, where the initial fountain of hot lava droplets coalesce back into a coherent lava and flow away from source. Examples of this have been recorded from Hawaii and from Japan, but such a process may be difficult to identify in the field as the result will look like a normal lava flow. This is also true of examples of rheomorphic ignimbrites (see Chapter 5).

VOLCANICS 1 – LAVA FLOWS

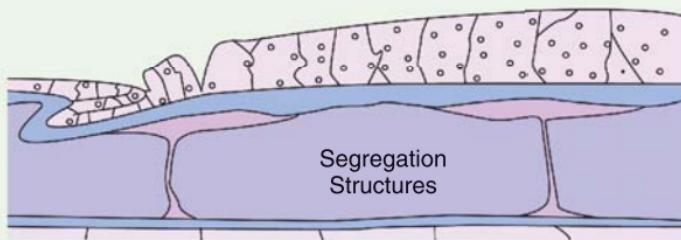
1. Breakouts (hours)



2. Inflation (days to months)



3. Stagnant lava flow (months to years)



4. Final Frozen lava flow

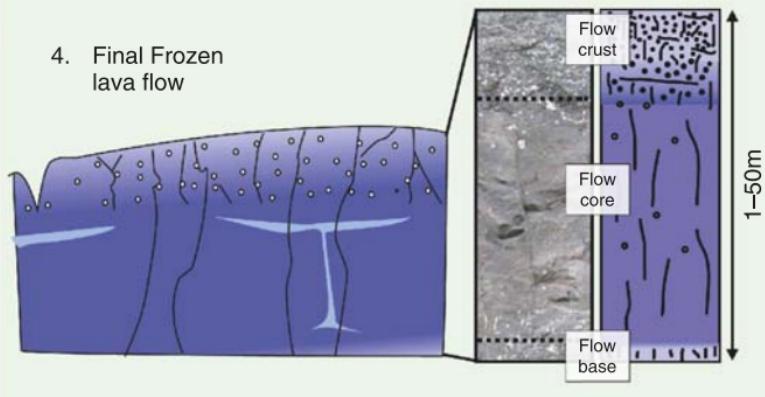


Figure 4.2 Inflation model for lava flow emplacement. The lava starts as a relatively thin flow and inflates from the inside once a crust has developed.

4.2 A Compositional Divide for Lava Flows

Although there is a continuum in composition from ultra-mafic through mafic, intermediate and felsic/silicic (acidic) compositions, two major types of lava flows are recognised readily in the field; mafic/basaltic flows and felsic/silicic flows. This reflects the nature of the predominance of one or other type of flow in a particular volcanic setting (for example, Chapter 1), and the major differences in viscosity from mafic/basaltic (low viscosity) to intermediate-felsic/silicic (high viscosity). In this book we will present the major lava flow structures under these two main sub-sections, but one should bear in mind that there is overlap in some instances.

When confronted with many lava flows in the field one can be somewhat stuck to work out the composition and therefore classification. If the lava is fine grained or a quenched glass with no crystals visible, a type of flow known as aphyric, the colour index (as described in Chapter 3) is not of use to qualify the composition (for example, Obsidian, a highly silicic glass is black in colour). In such cases the morphological features of the deposit described below in this chapter may help in determining an approximate ‘type’ for the lava. However, in most cases lava flows erupt carrying a crystal cargo (within the magma) known as phenocrysts, which usually represent the main phases crystallising from the melt. These lava flows, termed porphyritic lavas (for example, Figure 4.3), are very useful as the phenocryst population can be used to gauge the lava composition (for example, Table 4.1). In combination with features such as colour, field relationships and mode of occurrence, Table 4.1 is a useful guide to naming porphyritic rocks in the field. For example, olivine is most common in basalt and basaltic andesite lavas; abundant feldspar (plagioclase) with hornblende is characteristic of andesite; and quartz and feldspar (alkali feldspar) are diagnostic of rhyolite.

Basic rocks composed of natural glass are termed *tachylite* and these form on the skin of lava flows or along the marginal zones of basic sills or dykes. Volcanic glass of intermediate and acid composition is termed *obsidian* or *pitchstone*. Obsidian is a black or dark-coloured glass, with a bright glassy lustre and well-developed conchoidal fracture. Pitchstone has a dull, *resinous* or pitchy lustre, rather than the glassy lustre of obsidian. (This reflects the high content of water in pitchstone in comparison with obsidian.) Fragments of both types may occur as pyroclastic materials, or form minor intrusions or lava flows. Glassy volcanic rocks of intermediate and particularly acid composition commonly show flow banding (cf. Figure 2.9) as a result of the drawing out of small inhomogeneities during viscous flow.

Such natural glasses are unstable and start to hydrate and to crystallise or *devitrify* soon after formation. This process takes place progressively so that, in general, there are hardly any unhydrated natural glasses older than about 20 Ma, and, although glasses of Palaeozoic age are known, such glasses rarely exceed



Figure 4.3 Photo of porphyritic lava, Faroes. Large white/light tabular feldspar crystals in a fine grained basalt matrix. Note that tabular shapes show many long intersects in 2D cross section.

Table 4.1 Composition of major phenocryst phases in porphyritic lavas.

	Olivine basalt/ picrite	Basalt	Basaltic andesite	Andesite	Dacite	Rhyolite
Plagioclase	**	**	***	***	***	**
Olivine	***	**	**	*	—	—
Pyroxene	*	**	**	**	*	—
Hornblende	—	*	*	**	**	*
Biotite	—	—	—	*	**	**
Alkali feldspar	—	—	—	*	**	***
Quartz	—	—	—	—	**	***
Fe-Ti oxide	*	*	*	*	—	—

**Frequently present.

***Often present.

*Rarely present

—absent (or very rare)



Figure 4.4 Photo of spherules in a glassy pitchstone from the Etendeka, Namibia.

100 Ma in age. However, many ancient volcanic rocks must have originally contained natural glasses. Crystallisation occurs by spontaneous growth of tiny crystals or *crystallites*, commonly of quartz and feldspar. Usually, the individual crystals are not visible in hand specimens but devitrification crystals, in radiating clusters forming small circular structures resembling cod roe or oolites (termed *spherulites*), may be scattered throughout the devitrified rock (for example, Figure 4.4). Complete devitrification yields an aphyric microcrystalline rock, which may be light-coloured lava. (Such rocks may be termed *felsite*, but the term is also applied to any light-coloured fine-grained rock composed chiefly of quartz and feldspar.)

4.3 Mafic/Basaltic Lava Flows

As the most common occurrences of lava flows are mainly basaltic in composition we will concentrate on these in detail, before introducing examples from felsic/silicic flows, and eruptions of lava in water.

4.3.1 Surface textures

The surfaces of subaerial mafic lava flows belong to two major types: *Pahoehoe* and *Aa* (Figure 4.5). *Pahoehoe* flows (Figure 4.5a) have smooth, billowing ropy or entrail-like (*entail-pahoehoe*) crusts of quenched glass. Ropy crusts

VOLCANICS 1 – LAVA FLOWS

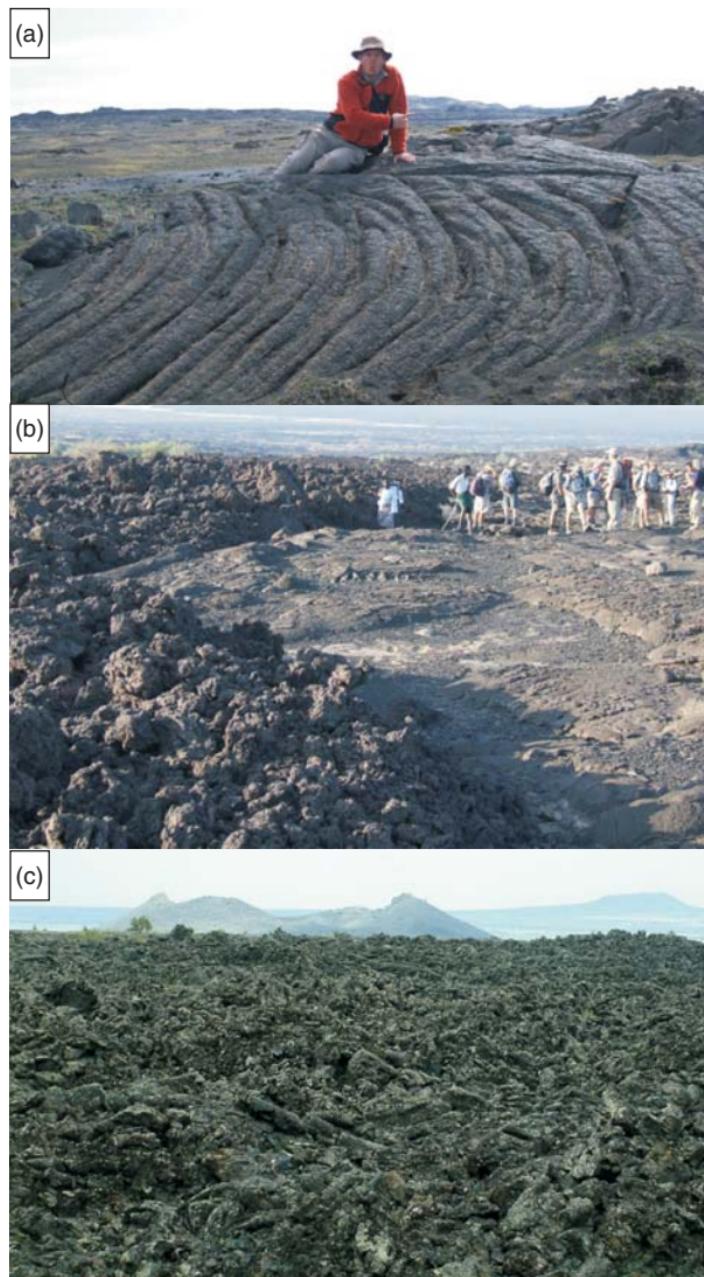


Figure 4.5 Photos of lava surface textures. (a) Pahoehoe, Iceland. (b) Aa over pahoehoe, Ethiopia. (c) Blocky Pahoehoe, China.

are sometimes considered characteristic of pahoehoe flows but are much less common than smooth crusts and small lobes which form from break-outs on the flow surface. Aa flows (Figure 4.5b) have rough clinkery and spinelike (spinose) surfaces reflecting tearing apart of fragments of the solidified crust on the top of lava, that is more viscous than that forming pahoehoe flows.

Both pahoehoe and aa lava surfaces may occur upon the same lava flow, changing from pahoehoe near to the vent to aa further downslope. Also variations between the two have been termed blocky pahoehoe. It is thought that subtle changes in the amount of microlites (small crystals) can dramatically change the viscosity of the lava moving it from a pahoehoe to aa type of flow.

4.3.2 Internal structures

It is particularly important to look at sections through lava sequences in the field. In thick sequences of basaltic flows one can recognise different volcanic facies based on their internal structures and the way in which the different lava units build up the volcanic stratigraphy (for example, Figure 4.6). In general subaerial lavas have an internal structure which varies from a lower to central core part and a minor lower and thicker upper crust region. These zones represent the well de-gassed inflated core region and the fractured and vesicular rich crust units which develop during emplacement (see Figure 4.2).

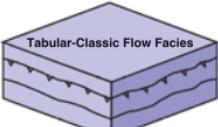
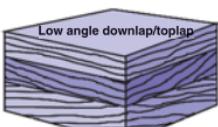
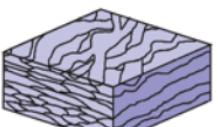
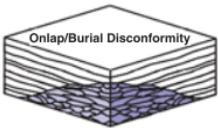
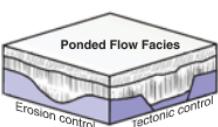
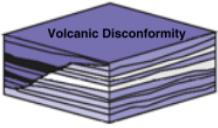
Examples of facies types	Facies types	Examples of facies associations
 Tabular-Classic Flow Facies	Tabular-classic: laterally extensive thick (>50m) flows, often well-developed columnar jointing	 Low angle downlap/toplap
 Compound-Braided Flow Facies	Compound-braided: Thin anastomosing pahoehoe flow sheets up to a few metres thick	 Onlap/Burial Disconformity
 Ponded Flow Facies	Dipping hyaloclastites: Dipping prograding foresets of hyaloclastites, up to tens of metres thick Ponded flows: Eruptions fill pre-existing topography, units can be 100 m thick Sills and dykes: Sills are often found at the base of the province, dykes are associated with igneous centres Facies associations	 Volcanic Disconformity

Figure 4.6 Examples of key volcanic facies found in flood basalts (thick sequences of predominantly basaltic lavas). Facies can be mapped out from flow to flow, but some of the facies associations may not be fully realised until the whole sequences has been mapped out in detail.



Figure 4.7 Photo of entablature and colonnade columnar joints, Iceland (photo: Tony Philpotts).

Some basaltic lavas, sills and dykes are characterised by a very well developed *columnar jointing* (Figure 4.7). These reflect contraction during cooling of solidified lava. These joints are generally orientated perpendicular to the upper and lower cooling surfaces of the flow, and split the lava into polygonal columns. The columns typically have five or six sides but some have three or four, and a few may have seven sides. The columns are mostly straight and parallel-sided, but some curve and vary in width when traced over a distance exceeding a few metres.

Columnar jointed lava flows often show a two- to threefold division, with regularly spaced units of straight-sided well-developed columns, termed *colonnades*, separated by zones composed of blocky, irregular columns arranged in a complex way, termed the *entablature*. This entablature-colonnade structure has been shown to be well developed in eruptions that have been significantly influenced during cooling by water (Figure 4.8).

Lava flows commonly show evidence of vesiculation. Vesicles may be perfect spheres, but this is rare; they are more usually ellipsoidal or almond-shaped, evidence for the direction of lava flow. After formation, gases and hot solutions circulating through the rock may deposit minerals within vesicles. Such deposition may continue by circulation of groundwater long after the lava is cold. The mineral infillings are termed *amygdales* and these rocks may then be termed *amygdaloidal* (Figure 4.9) (for example, *amygdaloidal basalt*). The commonest

Formation of entablature/collonade in lava flows

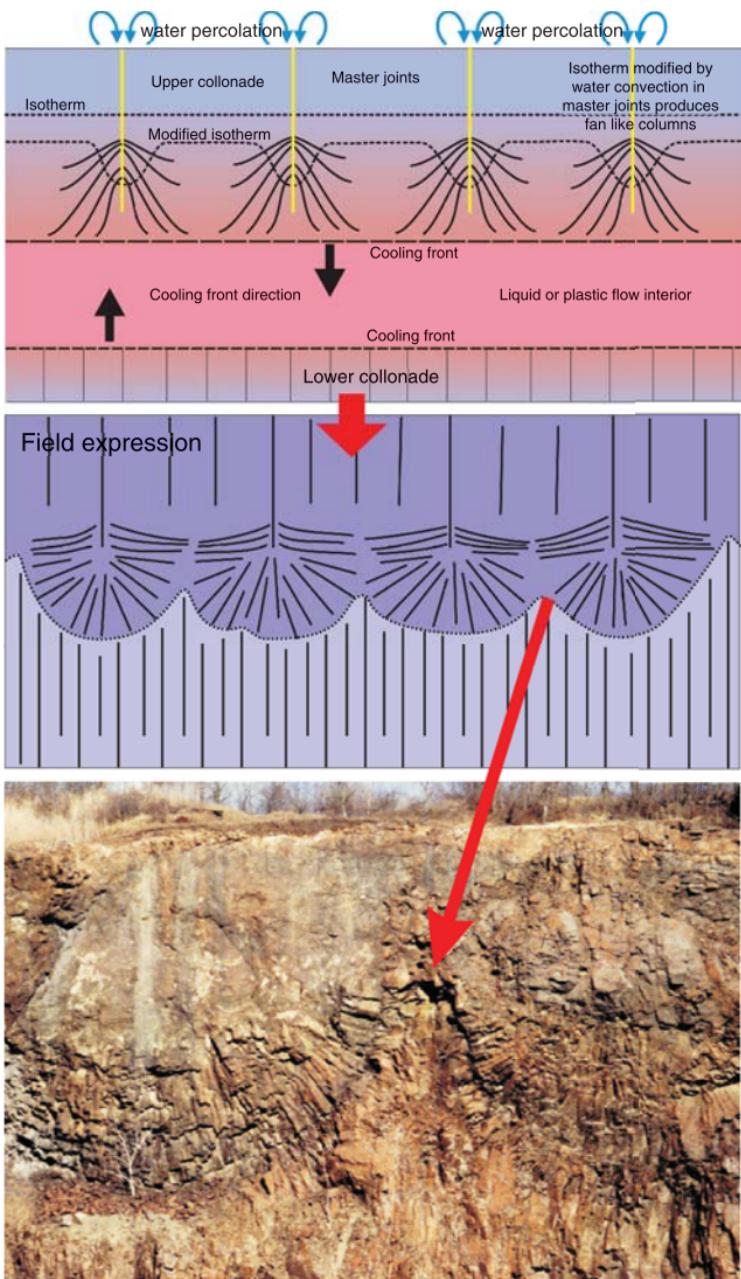


Figure 4.8 Cooling of columnar joints with aid of water (photo: Tony Philpotts).

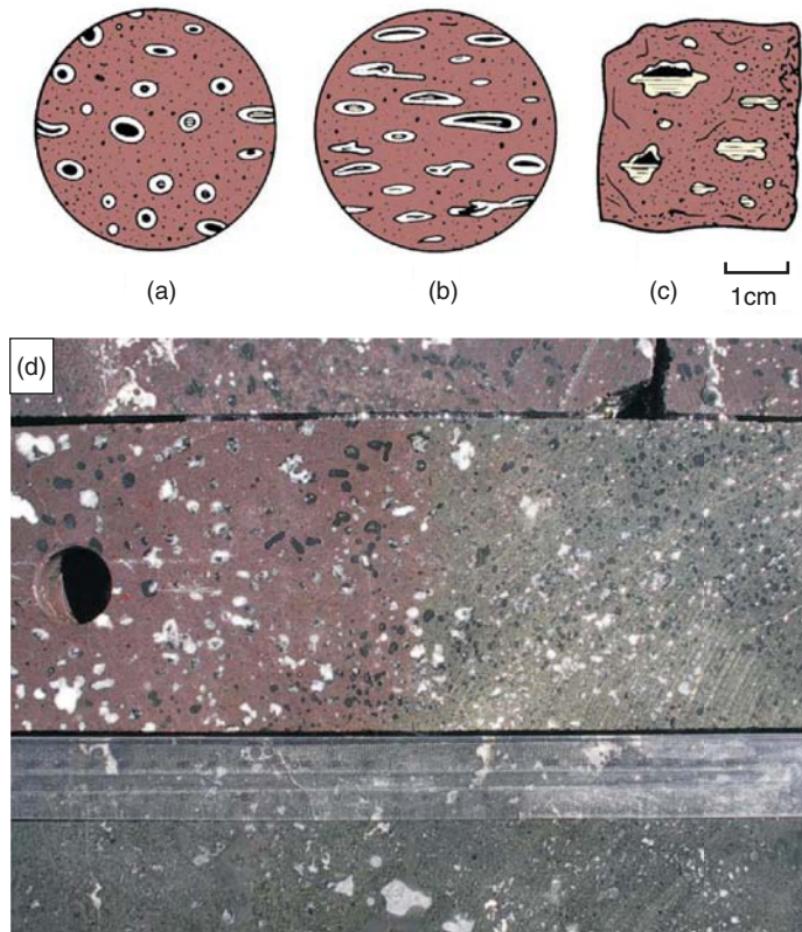


Figure 4.9 (a–c) Schematic amygdaloidal structure in lava. (a) and (b) Silica (white) was deposited on the walls of the original vesicles, and subsequently chlorite (lined) filled the remaining open spaces. The vesicles in which the amygdales of (b) were deposited were flattened and drawn out by the flowing of the lava. (c) Amygdales and larger cavities with lining and stratified fillings of silica indicating way-up. (d) Picture of amygdaloidal basalt from core samples, Namibia.

amygdales minerals include calcite, silica and zeolites; cf. Table 3.4. In some examples the amygdales are filled in a way that gives way-up indicators which can be of specific use in deformed lava sequences (for example, Figure 4.9c).

Such amygdaloidal fillings may be partial or complete and the filling may be zoned reflecting progressive deposition of hydrothermal minerals towards the centre. In older terranes flattening of amygdales might occur as a result of tectonic deformation accompanying folding. This may be distinguished from flow flattening, as the former should be parallel to regional tectonic trends on a medium (that is, outcrop) and larger scale.

Structures formed during movement and solidification of basic lavas can be used to determine the stratigraphic top, and the overall shape and direction of movement within a flow. Certain associations of structure may be so characteristic of a given flow that they may be used for tracing units and mapping contacts. An idealised section through a pahoehoe basalt flow is shown in Figure 4.10. Non-viscous basic lavas may show alignment of phenocryst minerals such as plagioclase. Such lavas generally contain vesicles which may occur within continuous lines or sheets, giving the rock a banded appearance or platy fracture, but such structures are *not* necessarily parallel to the overall flow direction. Larger vesicles are sometimes more abundant in the upper parts of flows and may be stretched out parallel to the last movements of the lava. Within the lower part of the flow, cavities formed by upward movement of gas have a tubular form and are termed *pipe vesicles* (or amygdales), which may join upwards giving additional way-up evidence. The occurrence of such vesicles may therefore be

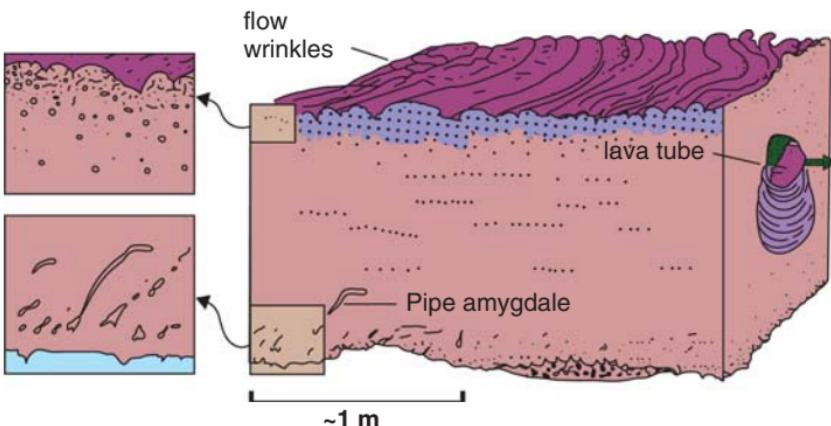


Figure 4.10 Idealised section through a pahoehoe basalt lava flow showing a partly filled lava tube, wrinkled and vesicular pahoehoe surface and pipe amygdales near to the base.

used to locate the base of the flow (way-up) and possible directions of final movement of the lava where bent (for example, Figure 4.10). Withdrawal of lava from the centre of a solidified lava flow may leave a hollow space within the flow which is termed a *lava tube*. These are common in pahoehoe flows and the occurrence of filling material on the sides or base of the tube may give evidence for the position of the top of the flow. The interiors of most aa flows are more massive than those described above, but some have a platy structure defined by sheets of small vesicles. Vesicles may be aligned in complex flow patterns but these are less well developed than in the case of pahoehoe lavas described above.

4.3.3 Scales of basaltic flows

One important thing to think about when mapping out lava sequences is their scale. Individual lava flows can range in scales over several orders of magnitude, with the largest flow units being greater than 1000 km^3 . More commonly you will be dealing with erupted units less than 1 km^3 in size, and it may be possible to map out lateral changes in thickness and volcanic facies on scales of a few outcrops. In some of the larger terrains, known as flood basalt provinces, the scales of flow unit and sheer volume of the whole sequence, require a good understanding of different scales of heterogeneity in the system and may involve a regional and even cross continent appreciation of the distribution of the volcanic facies (for example, flood basalts that have occurred during continental break-up such as North Atlantic Province or Paraná-Etendeka province Figure 4.11).

4.4 Felsic/Silicic Flows

The major difference in lava flows formed from more silicic compositions is the change in viscosity. This leads to lower aspect ratio flows, sometimes very little flow at all, and the development of joints and additional lava structures within the flow. A brief outline of some key features and example is given below.

4.4.1 Internal structures and morphology

Whereas aa surfaces occur on both basalt and basaltic andesite lavas, most andesite, dacite and rhyolite lavas are characterised by *blocky lava* surfaces. These surfaces have detached polyhedral blocks with planar or slightly curved faces and conspicuous dihedral angles. Intermediate and acid lavas sometimes show columnar jointing, but more commonly develop *platy* or *slabby joints*. The lower massive parts of andesite and dacite flows may be characterised by abundant close-spaced platy joints, ranging in dimensions from 1 to 50 cm,

VOLCANICS 1 – LAVA FLOWS

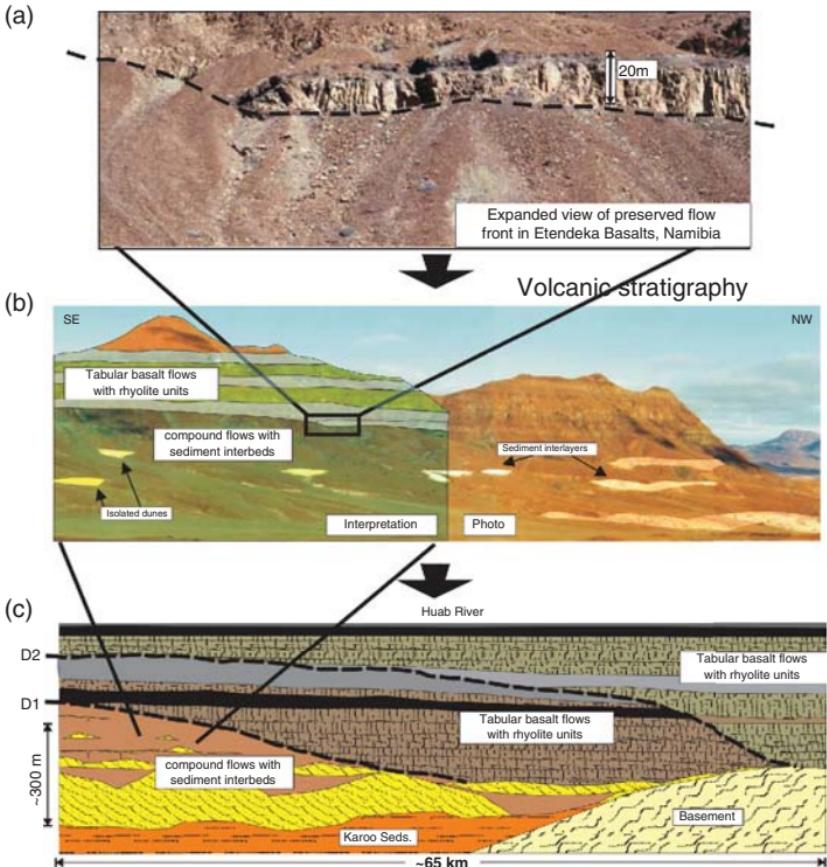


Figure 4.11 Scales of lava flood basalt heterogeneity in Etendeka flood basalts, Namibia. (a) Individual flows with their internal core and crust. (b) Lava flow stacking at large outcrop scale. (c) Mapped out lava stacking and lateral variations on a basin scale.

parallel to the underlying floor or curving down against the direction of movement, that is flat lying in the original lava. Traced upwards, such joint planes tend to become more widely spaced and slabby, and become more nearly vertical or even overturned near to the surface. Steep-sided platy joints on the surface of dacite or rhyolite extrusions may be termed *ramp-structure*. When recording the structures in silicic flows, it is useful to produce cross-sectional sketches to highlight the style of jointing top and bottom structures, distributions of glassy and breccia facies, and so on. Good examples of this are given in Figure 4.12.

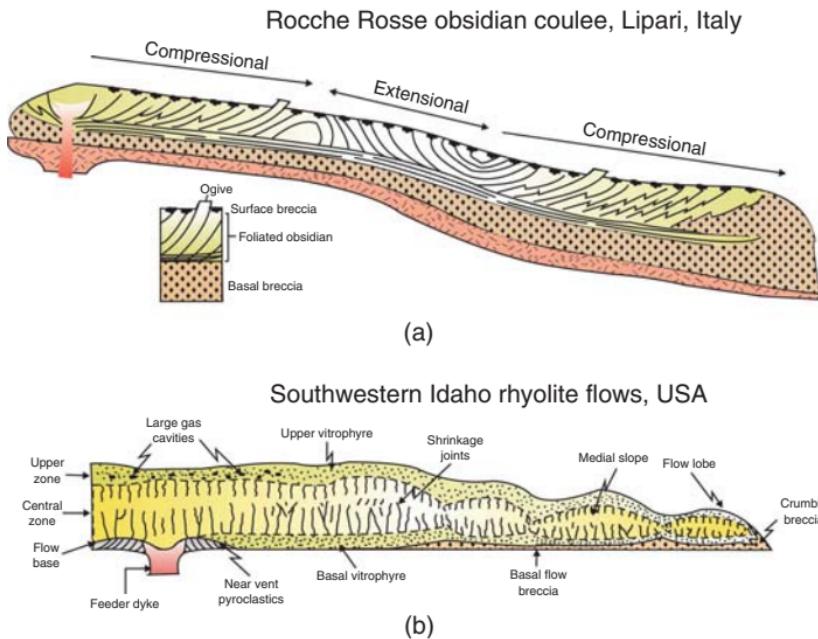


Figure 4.12 Examples of internal felsic flow structures.

4.4.2 Domes

Few dacite or rhyolite flows travel more than a few kilometres from the vent; these may reach thicknesses of up to several hundred metres. Extrusions of dacite or rhyolite lava that are circular in plan and have a height: diameter (h/d) ratio of 0.5–0.3 are termed *domes* (Figure 4.13). Domes commonly have heights of 50–150 m and diameters of 150–500 m. Domes form commonly in volcanoes in arc settings (for example, Chapter 1), and can collapse to form pyroclastic flows when they get over steepened during inflation. Dome growth can occur within the top cone of the volcano (for example Figure 4.14), or as a dome growth in the centre of a recently erupted and destroyed volcano as a *resurgent dome*, which starts the process of rebuilding the volcano (as in Mt St Helens, USA; Figure 4.15). Domes are most easily recognised in modern volcanoes and the recent geologic record.

4.5 Pillow Lavas and Hyaloclastites

Pillow lavas (Figure 4.16) are basalts that erupted below water and, since the upper part of the ocean crust is basaltic, such lavas may be the most abundant volcanic rocks of the Earth's surface (see also Chapter 8). However, it should be

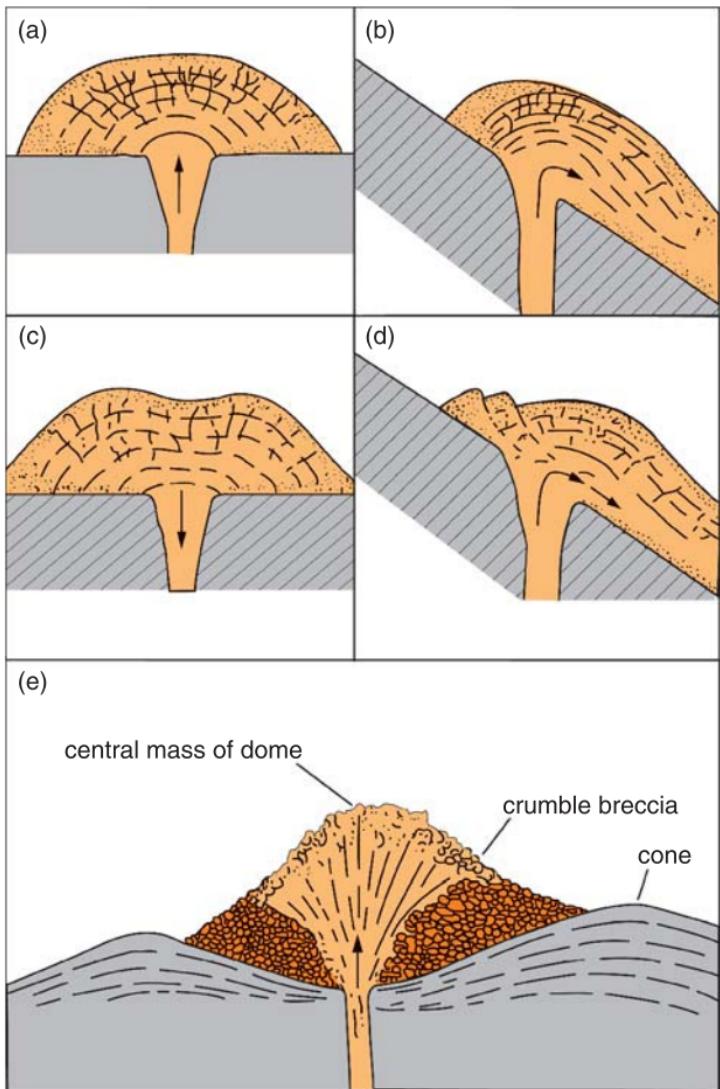


Figure 4.13 Schematic cross-sections of domes, showing flow lines (fine lines), joints (heavy lines) and brecciated margins (stippled). The arrows indicate the direction of lava movement. (a) Dome built on a nearly horizontal surface. (b) Dome built on a sloping surface. (c) The top of the dome has sagged due to drainage of magma back into the vent and to slight spreading of the dome. (d) Summit of the dome broken by faults due to down-slope movement of the main mass of the dome. (e) Cross-section of a dome built within a composite cone, showing steep fan structure developed within a confining wall of crumble breccia.



Figure 4.14 Photo of dome from Colima, Mexico (2009).

emphasised that pillows may form in any depth of water below a few metres, and even in mud and in melt water below a glacier. They are formed when lava is chilled rapidly by water: a chilled rind forms on the surface which is sufficiently flexible to move with the flow which remains molten inside. But as lava pressure increases, the rind breaks, causing formation of a propagating bulbous or tubular lobe. Such pillows vary from a few centimetres to several metres in their greatest dimension. The pillows have thin skins composed of basaltic glass, *tachylite* (reflecting rapid chilling of lava entering sea water), which in some cases has been altered by oxidation and absorption of water into a yellowish-brown waxy-looking substance, termed *palagonite*, and then to low-grade metamorphic minerals (for example, zeolite and chlorite). Palagonite may form subaerially or below water. The pillows are sometimes connected by slender necks and have tops that are usually convex upward, whereas the undersurfaces may be flattish, concave upward or project downwards into interstices between the underlying pillows (Figure 4.17a). These characteristics may be used to infer the way-up of the volcanic pile.

The interstices between the pillows are often filled with fragments of broken glassy lava skins, cherts, limestone or shale, which when compacted, are lithified and later converted to metamorphic minerals. Many pillows have a radiating columnar structure and may show concentric zonation of vesicular zones, produced as volcanic gases were episodically exsolved within the core.



Figure 4.15 Photo of Mt St Helens dome, USA (c. September 2001).



Figure 4.16 Photo of pillow lavas, Iceland.

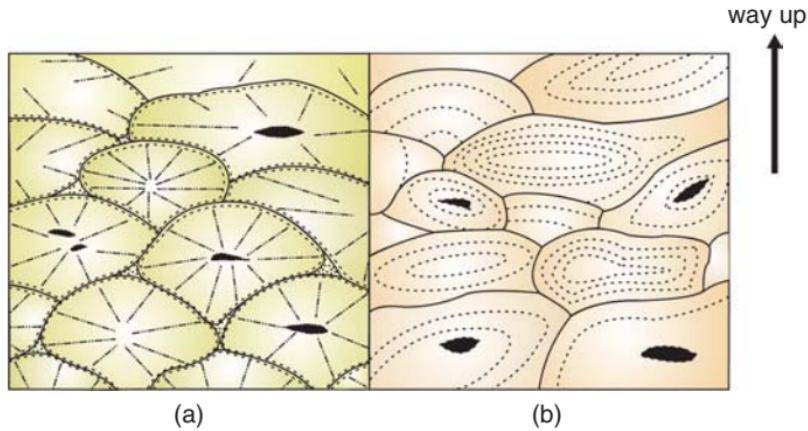


Figure 4.17 Diagrammatic cross-section showing the distinction between pillow lava (a) and lobes of pahoehoe lava (b). In (a) the short dashed lines represent the inner margin of the glassy skin on the pillows, the long dashed lines represent radial cooling joints and the stippled areas represent sedimentary material filling interstices between the pillows. In (b) the dotted lines reflect concentric variations resulting from cooling of the lava lobes. In (a) and (b) the black ornament represents spaces which would be open in young lavas, but would be filled with secondary material in older lavas.

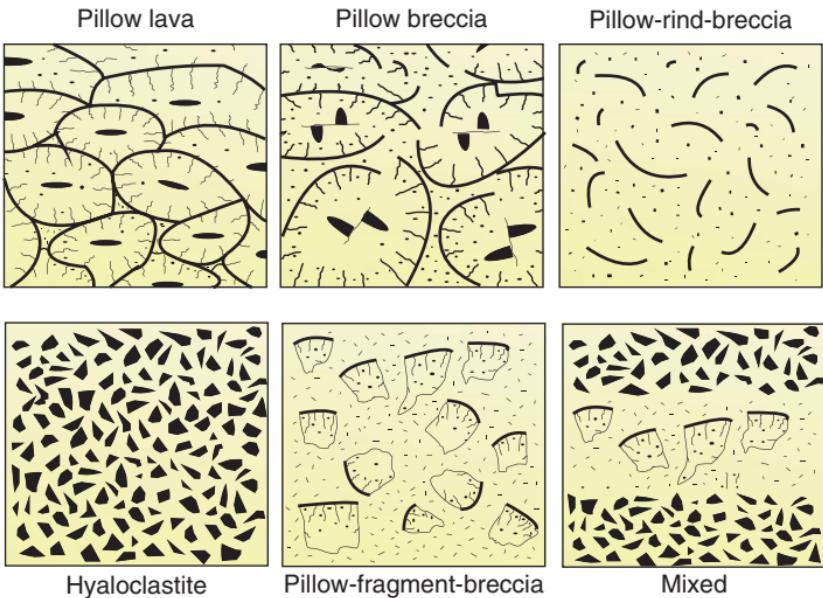


Figure 4.18 Different stages of fragmentation. Various mixtures of these facies may be found in close association to each other in any one example (For example, Figure 4.19).

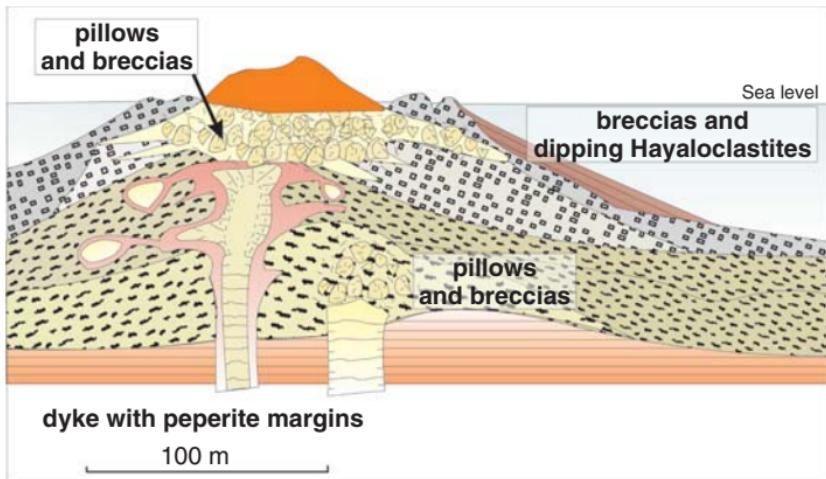


Figure 4.19 Schematic of emergent volcano showing various hyaloclastite breccias and pillow lavas.

VOLCANICS 1 – LAVA FLOWS

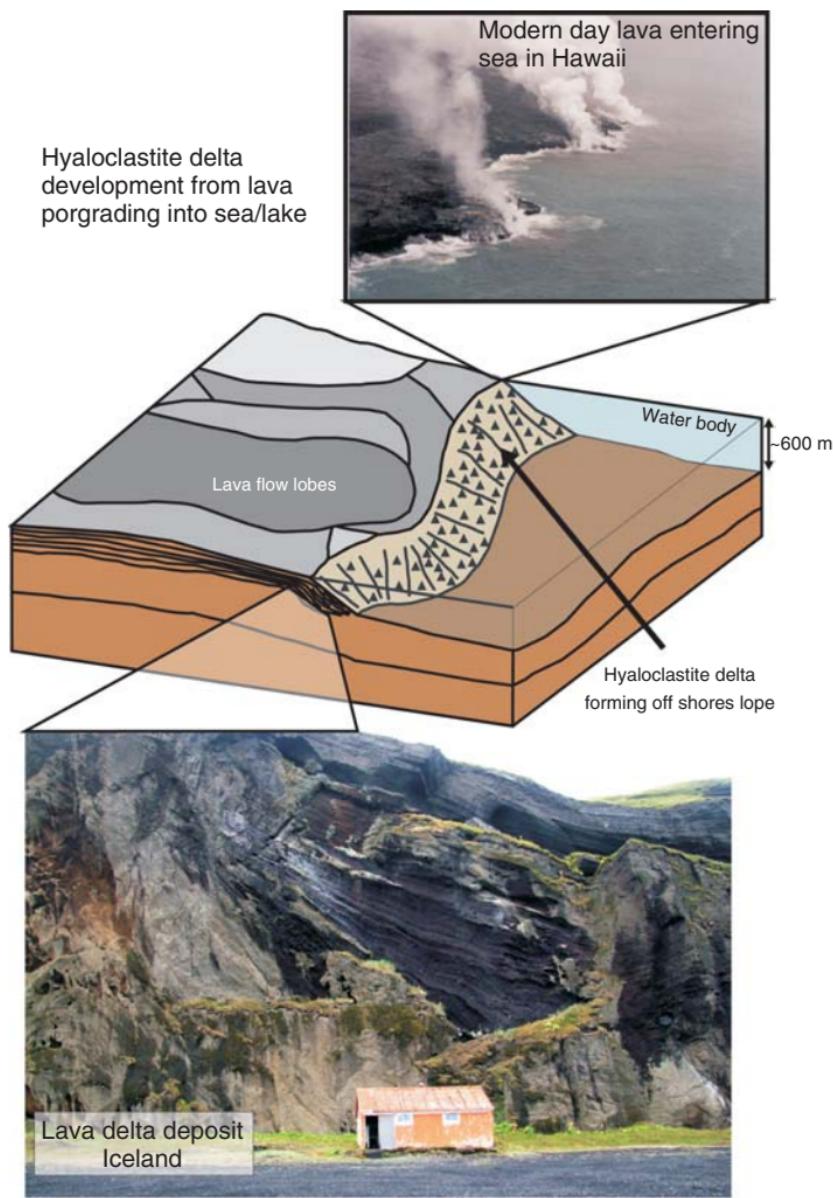


Figure 4.20 Figure of hyaloclastite delta showing modern day examples from Hawaii and a cross section through a sequence in Iceland.

However, since the pressure of sea water at a depth of 3000 m is sufficient to prevent exsolution of dissolved gases, the vesicularity of pillows varies from about 10–40% when they formed at shallow depths (less than 500 m) to below about 5% at 1000 m, and this may be used as a crude estimate of the depth of water into which the pillow lavas were erupted.

Pillows may be spheroidal, ellipsoidal or so flattened (even when undeformed tectonically) that they may be difficult to distinguish from other similar structures. For example, many ancient basaltic lavas contain pillow-like structures, such as lobes of (subaerial) pahoehoe lava, that might be confused with true pillow lavas (Figure 4.17b). In view of the palaeogeographic significance of pillow lavas (particularly in the identification of ophiolite complexes, see Chapter 8) and their value as ‘way-up’ indicators, you should examine the outcrop carefully for the features described above.

A further product of eruption of basaltic magma into water results from large-scale fragmentation of glassy lava (for example, pillow rinds and pillow lava fragments – Figure 4.18). Such fragmentation produces a deposit composed of glassy basaltic particles of sand size mixed with separated pillows. Such deposits are termed *hyaloclastite* or *palagonite-tuff* when the basaltic glass has been palagonised. Hyaloclastite commonly forms massive poorly-bedded deposits associated with subaerial lava, or submarine to emergent volcanoes (for example, Figure 4.19). In some cases hyaloclastite overlain by lava may be interpreted in terms of advance of (subaerial) lava from a shoreline into water to build a lava ‘delta’ that progrades the shoreline out into the water (Figure 4.20). Sometimes the hyaloclastite shows distinct bedding and this might then reflect reworking of the fine-grained glassy material by waves and currents, so that the deposits may contain sedimentary structures reminiscent of sand deposited in shallow water.

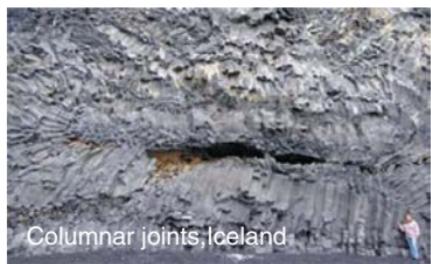
VOLCANICS 1 – LAVA FLOWS



Blocky pahoehoe crust, Tenerife



Flood basalts,
Deccan, India



Columnar joints, Iceland



Soil horizon,
Deccan, India



Pillow lava, Iceland



Giant spherule
Etendeka, Namibia



Pahoehoe, Iceland

Lava flows - Examples

VOLCANICS 2 – PYROCLASTIC ROCKS



Figure 5.1 Inspecting pyroclastic deposits on the side of Santorini Caldera, Santorini, Greece.

5

VOLCANICS 2 – PYROCLASTIC ROCKS

The second major category of volcanic rocks are those formed by the explosive eruption of magma at the surface resulting in pyroclastic rocks (Figure 5.1). These types of deposits can be found in almost all volcanic settings but are particularly common in convergent plate margins (Chapter 1). Here we will look briefly at the processes that form pyroclastic rocks, followed by some of the specific things that need to be recorded to help classify and quantify them. We will also consider some of the other types of volcaniclastic rocks which develop as the result of fragmentation of juvenile magma at the surface. The reader is reminded of the close association of pyroclastic/volcaniclastic rocks and rocks formed from lava flows described in Chapter 4.

5.1 Structures, Textures and Classification

Pyroclastic rocks are composed of materials fragmented by explosive volcanic activity. Three types of fragment may be found in any pyroclastic deposit.

1. *Fragments of new lava* which are termed *juvenile* and which may range from solid un-vesiculated material to fragments of highly vesiculated lava (for example, pumice and scoria) and shards of glass.
2. *Individual crystals*, yielded by liberation of phenocrysts present in the juvenile lava as a result of fragmentation.
3. *Lithic fragments* which may include any older rock within the deposit, but often comprising older lavas.

5.1.1 Primary structures

Bedding and associated structures that result from the depositional processes are wide and varied in pyroclastic rocks. This is due to the range of depositional environments, flow types and fragment types that make up the rocks/sediments constituents. It is important to make sure that you are able to recognise primary deposits from those which are re-worked (these will be dealt with under epiclastic deposits later in this chapter).

VOLCANICS 2 – PYROCLASTIC ROCKS

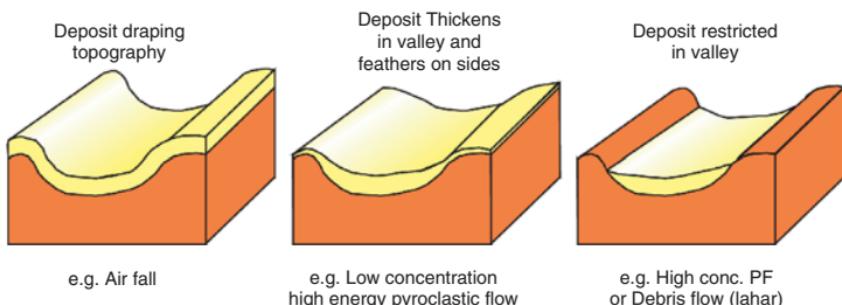


Figure 5.2 Deposit characteristics related to topography.

Pyroclastic sediments can be deposited by pyroclastic fall (fall out) and by pyroclastic flows and therefore a large range of currents and sorting mechanisms can affect the resultant deposits. Fall out deposits will blanket topography, whereas pyroclastic flows will commonly be restricted to pre-existing valley topography, so an understanding of the deposits lateral as well as vertical variation is critical in understanding the rocks (for example, Figure 5.2). Sedimentary bedding structures again will vary considerably depending on the type of density current (high or low concentration) which sediments are being deposited from and from fall out type layers. Examples of common sedimentary structures are given in Figure 5.3. For example, low concentration lateral blast type deposits (sometimes termed base surge) can blanket topography like fall out but are recognised by their more poorly sorted character and presence of tractional structures like cross bedding (see Figures 5.3 and 5.4).

Bedding structures vary from massive poorly sorted to highly sorted to graded beds. A range of the grading structures that can be found is given in Figure 5.5. With grading in primary deposits it is important to understand the different densities of the particles within the deposits. Lithic fragments of a smaller size will have the same density as pumice fragments (air filled) of a much larger size, and it can be very useful to record both lithic and pumice size variations separately throughout the deposit. This is of particular importance in fall out deposits.

5.1.2 Classification

Pyroclastic fragments are classified by the type of material and their grain-size limits. The grain size limits used in pyroclastics is given in Table 5.1 including reference to the phi units used in sedimentary rock classification (see Tucker). A simple classification of and nomenclature of tuffs and ashes based on their fragmental components is given in Figure 5.6. Note the term *tuff* (from the Italian ‘tufo’) is used for consolidated (ancient) rocks and *ash* used in unconsolidated

VOLCANICS 2 – PYROCLASTIC ROCKS

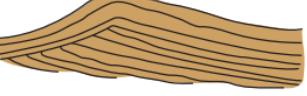
 direction of transport →	Gently sloped wave of generally long wavelength, low amplitude often grading laterally into planar beds, and found as faint laminations in fine-grained massive beds.
	Relatively symmetrical antidune shape, built up on the stoss side, elongated on the lee side, with one more marked inner unconformities.
	Festooned dunes; direction of transport is perpendicular to the plane of the paper.
	Cross laminations occurring in bedding sets 2 to 8 cm thick.
	Chute and pool structure with coarse-grained, steeply dipping stoss side.
	Symmetrical dunes with lee side accumulations of coarse material.
	Antidunes with rounded crest and internal unconformities.
	Sinusoidal ripple-drift laminations of short wavelength.

Figure 5.3 Morphologies of structures commonly found in low-density pyroclastic flow deposits.

modern deposits. Deposits are commonly poorly sorted, bi-modal and a classification using combination of different clast sizes is given in Figure 5.7.

Juvenile vesiculated fragments (Figure 5.8a,b) range from highly vesiculated pumice (density may be less than 1000 kg m^{-3}), through less well vesiculated juvenile fragments, sometimes termed ‘expanded andesite or dacite’ (density



Figure 5.4 Cross bedding tractional structures in the basal parts of the Minoan ignimbrite deposit, Santorini, Greece.

about $1000\text{--}2000\text{ kg m}^{-3}$) to poorly vesiculated or non-vesiculated rock fragments (density over 2000 kg m^{-3}). Pumice is generally taken to apply to fragments of intermediate or evolved (felsic) chemical composition of density about 1000 kg m^{-3} or less. Scoria is the term used for dense vesicular fragments of basic or intermediate composition. The term *glass shard* is applied to ash-sized fragments. (Table 5.1) resulting from the fragmentation of bubble or scoria walls, so tending to have Y-shaped or cuspatate shapes (Figure 5.8a).

Crystals may be derived by fragmentation of partially-crystallised porphyritic magmas and often may behave as a distinct population during transport and deposition within pyroclastic deposits (Figure 5.8c–e).

Lithic fragments (Figure 5.8f) are usually the densest components of a pyroclastic deposit and include non-vesiculated juvenile fragments which are genetically related to the volcanic host, termed *cognate lithics*, and accidental country rock fragments incorporated during eruption or transport and which are, respectively, termed *accessory* and *accidental lithics*.

Tephra (Greek, for ash) is a generic term for any airborne pyroclastic accumulation usually given to unconsolidated deposits. The use of *tephra* layers, which bear their own unique chemistry and character, as temporal marker horizons is known as *tephrochronology*.

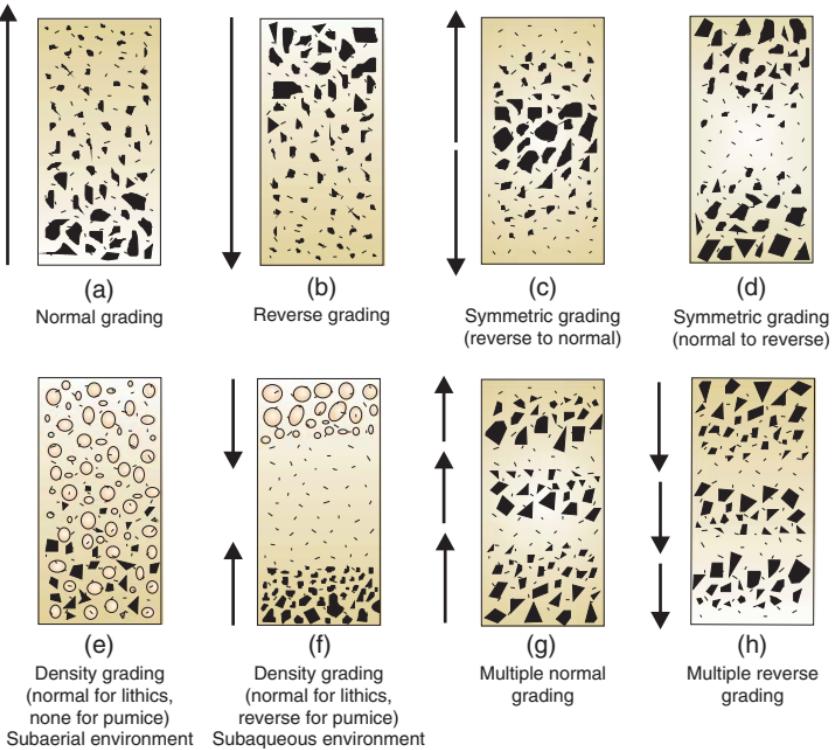


Figure 5.5 Different styles of grading as seen in ignimbrite deposits.

5.1.3 Textures

The textures that are found in pyroclastic rocks are wide and varied, due to the varied size range (huge blocks to very fine ash), density variations in the clast population (for example, pumice vs lith fragments) and the differing current regimes that can deposit the units (for example, fall out, high and low density pyroclastic flows, and so on). Many of the standard sedimentary structures and textures associated with the emplacement of turbidites-debris flows-grain flows are of importance (for example, tractional and grading structures – as above; see also Tucker, this book series), and it is important to note the specific rock texture in terms of the grain size (Table 5.1), grain size sorting (Figure 5.9) and the modal abundance of the key fragments. Additionally there are a number of specific textures that may be found in only pyroclastic units which include:

1. Accretionary lapilli (Figure 5.10), which are small spheres found in both fall and flow deposits, varying in size from 1 mm to several centimetres, but

VOLCANICS 2 – PYROCLASTIC ROCKS

commonly 2–10 mm in diameter. They may consist of a relatively thin shell of fine-grained ash around a core of coarse ash, or be composed wholly of concentric layers of fine-grained ash (as in oolites). Such lapilli are believed to be formed by accretion around solid nuclei in a medium of condensed moisture (cloud and/or rain) within an eruption column.

- Impact structures, commonly termed bomb-sags (Figure 5.11) in fall and flow deposits, which result from fall of bombs (or more commonly, blocks) into finely bedded, commonly water-saturated sediments. Such impact structures are characteristic of pyroclastic deposits erupted in wet conditions.

Table 5.1 Grain size limits for pyroclastic fragments.

Grain size ¹		Pyroclastic fragments		
(φ)	(mm)			
-8.0	256	Coarse	{	Blocks and bombs ²
-6.0	64	Fine		Lapilli
-1.0	2	Coarse	{	Ash
+4.0	1/16	Fine		

¹The grain size is expressed in two scales. The ‘phi’ (φ) scale has values which decrease arithmetically as grain size (expressed in mm) increases geometrically. The conversion from the mm scale to the phi scale is arithmetically very simple and is expressed by the equation $N = 2^{-\varphi}$, where $\varphi = 1, 2, 3$, and so on. and N is the grain size in mm.

For example, consider a grain size of 2 mm:

$$N = 2^{-\varphi} 2 = 2^{-\varphi}$$

and

$$2 = 1/2^{\varphi}$$

Therefore

$$2^{\varphi} = \frac{1}{2} = 2^{-1}$$

So, for a grain size of 2 mm, $\varphi = -1$.

Therefore, zero on the phi scale corresponds to a grain size of 1 mm, whereas $\varphi = +1, +2$ and $+3$ correspond, respectively, to $\frac{1}{2}, \frac{1}{4}$ and $\frac{1}{8}$ mm and $\varphi = -1, -2$ and -3 correspond, respectively, to 2, 4 and 8 mm, and so on.

²A block that takes on a distinctive form during fragmentation, flight or upon landing is termed a volcanic bomb. Such bombs are often given descriptive names such as ribbon bombs, spindle bombs, breadcrust bombs and cowdung bombs.

Sorting (σI): (Inclusive Graphic Standard Deviation) $\sigma_I = \frac{(\phi_{84} - \phi_{16})}{4} + \frac{(\phi_{95} - \phi_5)}{6.6} \phi_x$ is the grain size (phi units) at the $x\%$ value on the cumulative frequency curve (see Tucker, this book series).

3. Eutaxitic textures – these refer to textures formed in hot ignimbrites which start to weld back together (see below) and include streaked out pumice termed fiamme (Figure 5.12).
4. Fluidisation pipes – gas-escape structures, or fluidisation/fumarole pipes found more commonly in thick, massive ignimbrite deposits. These vertical columns or pipes contain lithic and coarse pumice fragments depleted of fine ash by gas streaming through these localised zones, which also often preferentially cements these horizons making them resistant to erosion (Figure 5.13).
5. Vapour-Phase Crystallisation – a post-depositional process, where crystallisation takes place in open spaces under the influence of a vapour phase. Hot vapours, are generally enriched in H_2O , CO_2 and SO_2 and Cooling of these element-rich phases may result in the crystallisation of a variety of minerals into open cavities. *Lithophysae* is a hollow, bubble-like structure found within the cavities of pyroclastic flows, and in some examples a well-cemented, white rock with little pore space known as *sillar* can develop and may be associated with fluidisation pipes.

Probably the most important skill to work on when recording pyroclastic deposits is the use of schematic and quantitative logs (for example, Chapter 2), accurate grain-size information and a keen eye for sedimentary structures (Figures 5.4 and 5.5). If you are working in such areas you can construct your own log templates, use quadrant statistics and often trowels and small shovels

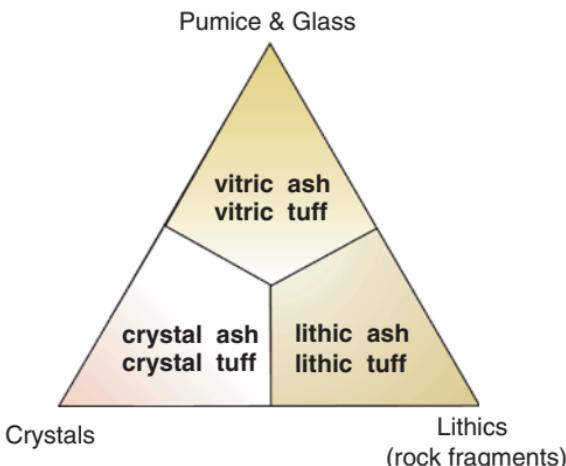


Figure 5.6 Classification based on the type of fragments present.

VOLCANICS 2 – PYROCLASTIC ROCKS

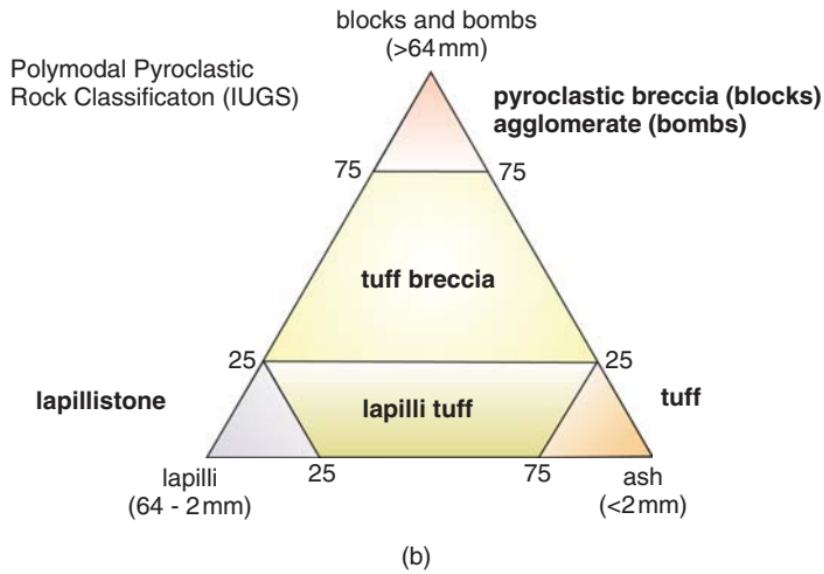
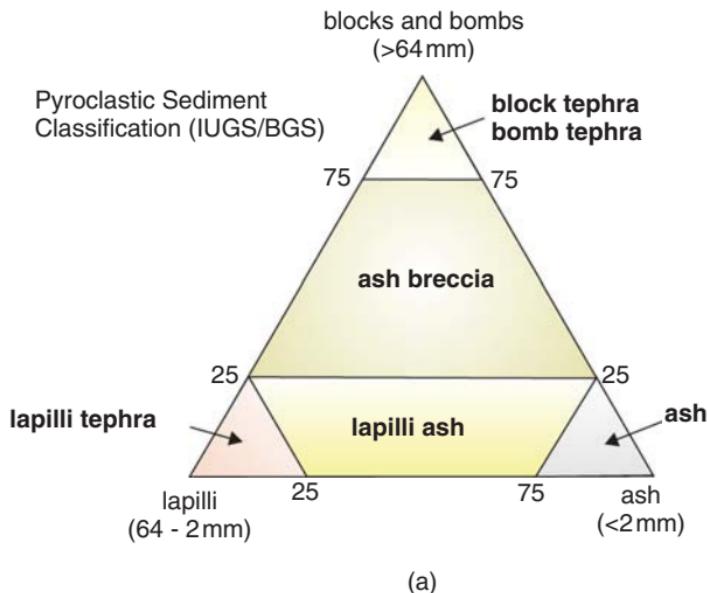


Figure 5.7 Classification system based on the sizes of material, separate for unconsolidated/modern (a) and rock record/ancient (b).

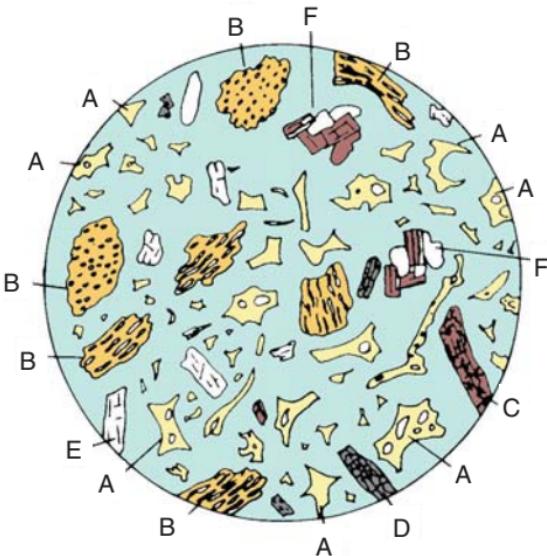


Figure 5.8 The components of a typical air-fall ash of dacite composition. The diameter of the field is about 2 mm. The ash has a fine matrix which has been omitted and consists of angular glass shards, (a) some showing cuspat forms; pumice fragments, (b) showing vesicles that are nearly round in cross section in one direction and much stretched in the direction right angles to the first; and crystals of hornblende, (c) biotite (d) and feldspar (e) with a little quartz and lithic fragments (f).

are useful in poorly or un-consolidated deposits (for additional instruction on recording sedimentary structures refer to Tucker, this book series).

5.2 Pyroclastic Flows and Ignimbrites

As with lava sequences (Section 4.1), it is very useful to have some conceptual idea of the processes which are involved in the deposition of pyroclastic rocks, to help understand how deposits form. Here we will briefly describe some of the recent developments in our understanding of *pyroclastic flows* which will help you to properly describe and record their deposits; *ignimbrites* and *block and ash flows*.

5.2.1 Understanding pyroclastic processes

Ignimbrite deposits are formed from explosive eruptions that result in turbulent mixtures of fragmented rocks, hot magma, crystals and ash which flow down

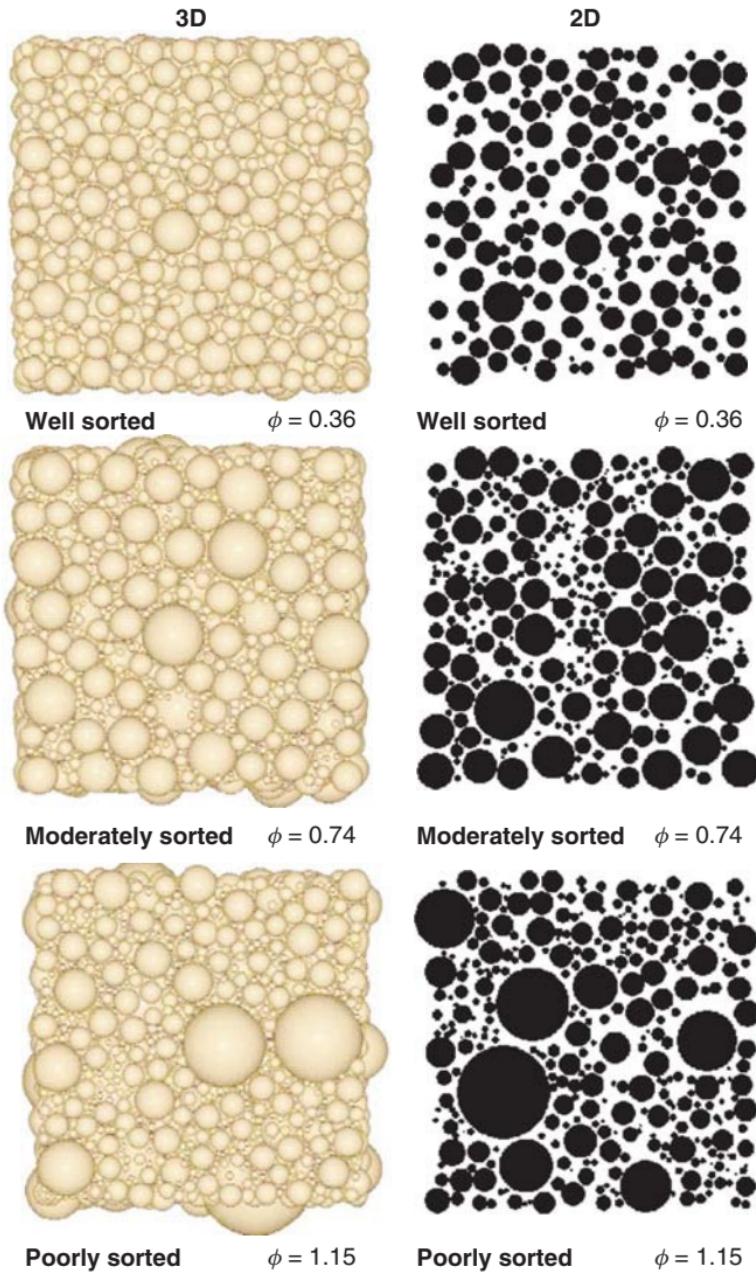


Figure 5.9 Models for estimating grain size sorting in 2D and 3D (sorting is worked out using a cumulative frequency curve of the grain size – see Table 5.1).



Figure 5.10 Accretionary lapilli in valley fill ignimbrite deposit, Tenerife. Note concentric rings round smaller pumice fragments (see also examples at end of chapter).



Figure 5.11 Bomb sag in tuff ring deposit, Tenerife.

VOLCANICS 2 – PYROCLASTIC ROCKS



Figure 5.12 Streaked out light and dark fiamme from welded ignimbrite, Tenerife. Light and dark examples show co-existence of two juvenile magmas.



Figure 5.13 Fluidisation/fumarole pipes resistant to erosion from Crater Lake, Oregon, USA.

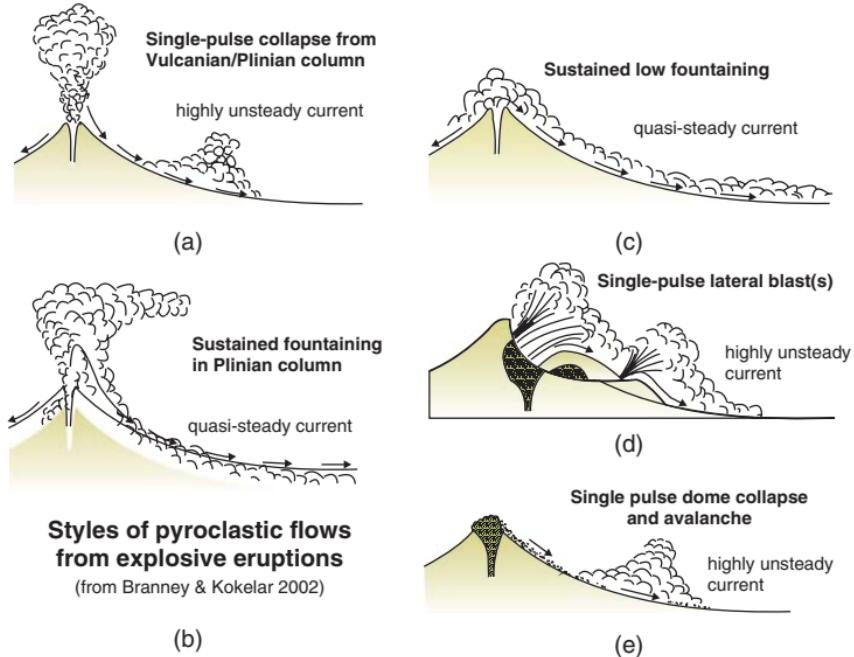


Figure 5.14 Schematic of the styles of pyroclastic eruptions (from Branney and Kokelar, 2002).

slope as density currents known as pyroclastic flows (previous terms include Nuee ardente – firey cloud, a term now more commonly associated with bock and ash flows). It is important to recognise that such mixtures of fragmented material can form in a number of ways during an eruption which will result in different currents and concentrations of material, for example Figure 5.14.

Another important thing to remember is that an ignimbrite deposit may well have formed over a protracted time, during which different currents and concentrations of material are involved in the deposition. You must think of the deposit building up through time (which could be minutes-hours even days in duration) from the base up, with the deposit being created at a boundary zone (the progressive aggradation model in Figure 5.15). As the flow velocity increases (waxes) and decreases (wanes), the style of the resultant deposit will reflect this.

5.2.2 Common ignimbrite deposit characteristics

Non welded ignimbrites can be quite un-consolidated and are susceptible to weathering very quickly after their deposition. Sometimes units can be lithified early as the constituents form a natural cement type mixture so may be preserved

VOLCANICS 2 – PYROCLASTIC ROCKS

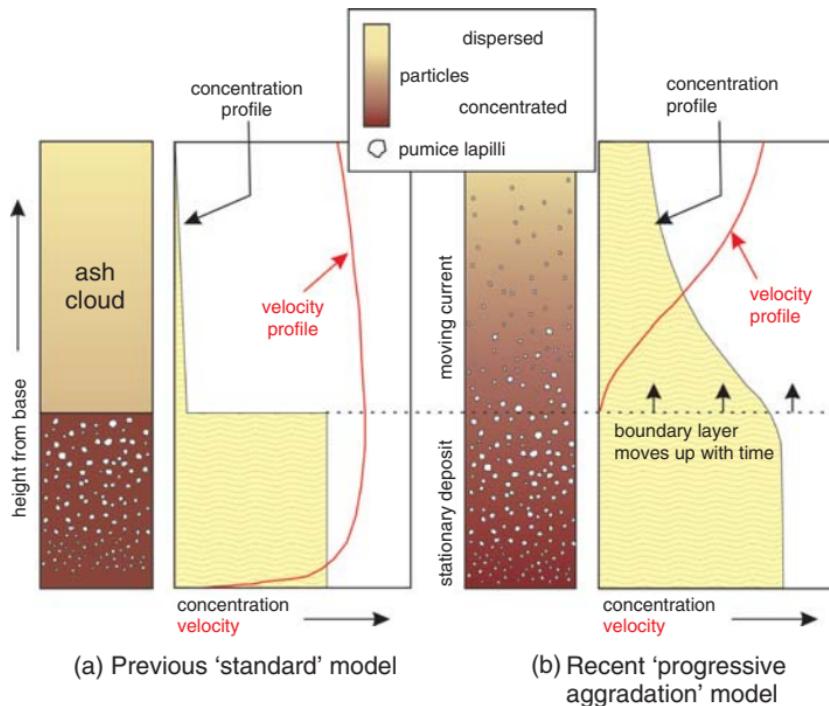


Figure 5.15 Schematic showing the previous and more recent models of ignimbrite emplacement from pyroclastic flows.

by early cementation/vapour phase alteration. A typical section through an ignimbrite is given in the schematic log in Figure 5.16. Here a basal ground layer commonly of fall out is followed by a basal deposit from a low concentration lateral blast (may also be termed surge in some literature), followed by the main ignimbrite deposit, a poorly-sorted mixture of lithics and pumice which may show an increase in pumice concentration and size towards the top of the deposit, which is normally capped by an ash sequence of varying size, sorting and structure, known as the co-ignimbrite ash. A field example of a non welded ignimbrite is given in Figure 5.17.

Welded ignimbrites are characterised by having a wholly or partly welded part of the deposit, which results from the deposit being very hot during deposition. In extreme circumstances the whole deposit may be so hot that it fuses back together to the characteristics of a lava of similar composition. Such ignimbrites are known as *rheomorphic*. More commonly the central/lower part of the flow becomes indurated and welded/intensely welded, displaying a *eutaxitic* texture known as *fiamme* (flattened pumice) in the densely welded parts. In extreme

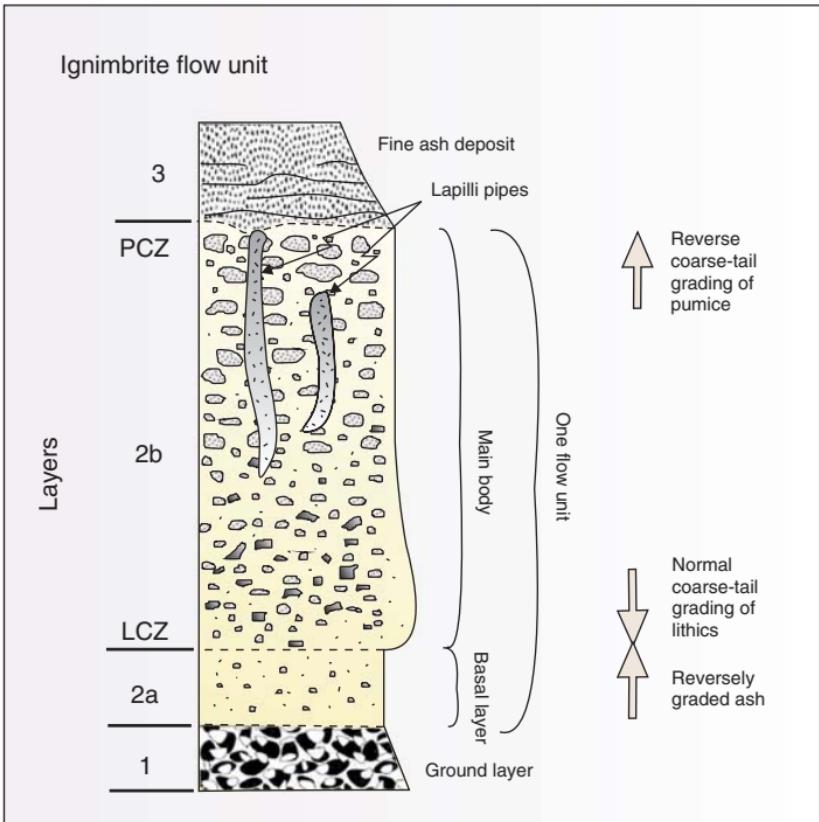


Figure 5.16 Idealised flow unit from a ponded ignimbrite. Layer 1 can be composed of lithic and/or pumice rich material, thought to deposit at the front of the pyroclastic flow, they can show tractional structures consistent with low concentration pyroclastic flows. Layer 2 can show a variety of structures and grading and can also vary from lithic to pumice concentrated zones (LCZ, PCZ). Layer 3 shows transition into low concentrated flows and fall deposits.

welding streaked out sheath folds can be present indicating extreme shear during rheomorphism. As with unconsolidated ignimbrites, welded examples are best characterised by logging the vertical and horizontal variations recording the major textures, grainsize and lithological variations, and so on (for example, Figure 5.18). An example of the welded part of an ignimbrite is shown in Figure 5.19.

Debris avalanches occur due to the collapse of volcanic material under gravity. These can be from dome collapse or from sector collapse of a volcano and



Figure 5.17 Thick non-welded ignimbrite deposit from the Minoan eruption, Santorini, Greece.

result in block and ash flows. In some instances they can be hot (i.e. with dome collapse and examples where hot injecting magmas causes slope failure), but also they can form from cold fragmented material. They tend to not travel as far as pyroclastic flows, and in the case of sector collapse result in a characteristic hummocky topography.

5.3 Scoria Cones

Scoria cones are very common small explosive eruption features that occur in most volcanic provinces. These are conical in morphology commonly with a crater at the top, and due to their unconsolidated nature the internal structures can be exposed by erosion and/or collapse. They usually represent single eruptions of relatively small volumes of magma and can commonly occur as fields of cones, or on the side of larger volcanoes where they are termed parasitic cones. Figure 5.20 highlights the stages of a scoria cone development, which commonly involve an initial build up phase with an original volcanic rim, which becomes reworked as the slopes get oversteepened and slide/collapse to form scoria talus (for example, Figure 5.20c,d). Around the apron of the cone you will often find bombs and other balististics. Examples of scoria cones are given in Figure 5.20e.

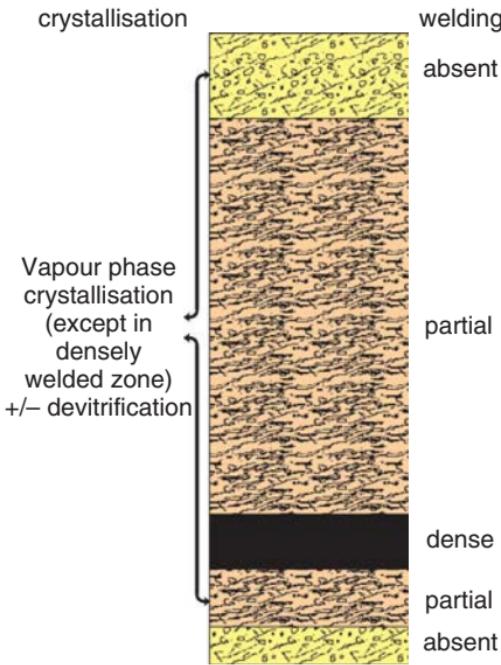


Figure 5.18 Variation in degree of welding within a single ignimbrite unit. The section is about 70–100 m in thickness.

5.4 Water/Magma and Sediment/Magma Interactions

5.4.1 Hydroclastic deposits

When eruptions occur under water or into ice then dynamic fragmentation of the magma can result in *phreatomagmatic* eruptions. Eruption of magma into, or through, water results in another suite of characteristic volcanic products which include the *hyaloclastites* (described in Chapter 4). The more violent pyroclastic eruptions in shallow water form small low-profile cones ($h/d = 0.1–0.2$) composed of fine-grained ash-grade pyroclastic material termed *ash- or tuff-rings* (Figure 5.11). Similar pyroclastic material is formed by eruption from larger composite volcanoes having crater lakes. In comparison with dry subaerial pyroclastic fall deposits, such material is generally finer-grained (M_d generally below 2 mm) and less well-sorted ($\sigma\varphi$, 1.5>). The individual layers are generally thin, commonly 1 mm to 1 cm, in comparison with much coarser subaerial pyroclastic layers (layers about 1–5 cm). Characteristic features of pyroclastic material formed by interaction with water are *accretionary lapilli* and *impact structures* (for example, Figures 5.10 and 5.11).



Figure 5.19 Example of welded ignimbrite structure, Bishop Tuff, Owens Valley, California. Note streaked out (dark) fiamme, with two lithic fragments maintaining their shape.

Eruptions under ice are characterised by the formation of flat-topped volcanic landforms termed *Tuyas* (table mountains) from examples in Iceland. Here the ice restricts the eruptive products leading to restricted and chaotic breccias (Figure 5.21a), and in some cases a meltwater lake will develop resulting in the formation of more typical bedded hyaloclastites as the volcano emerges and fills the accommodation space (for example, Figure 5.21b).

5.4.2 Peperites

The dynamic interaction of unconsolidated (commonly wet) sediments and juvenile magma result in a rock type known as *peperite*. These are relatively localised occurrences at the base of lava flows have flowed over and/or where intrusions have occurred into unconsolidated sediments. When sediments are wet the interaction is fuelled by the explosive fragmentation and *autobreciation* of the magma, and characteristic angular-irregular glassy fragments interspersed with indurated and fused sediment results (for example, Figure 5.22), commonly with sediment-filled dyke injections back into the host magma, and chilled contacts. Care should be taken not to invoke the role of water in all peperite formation as examples of dynamic interactions between juvenile magma and dry

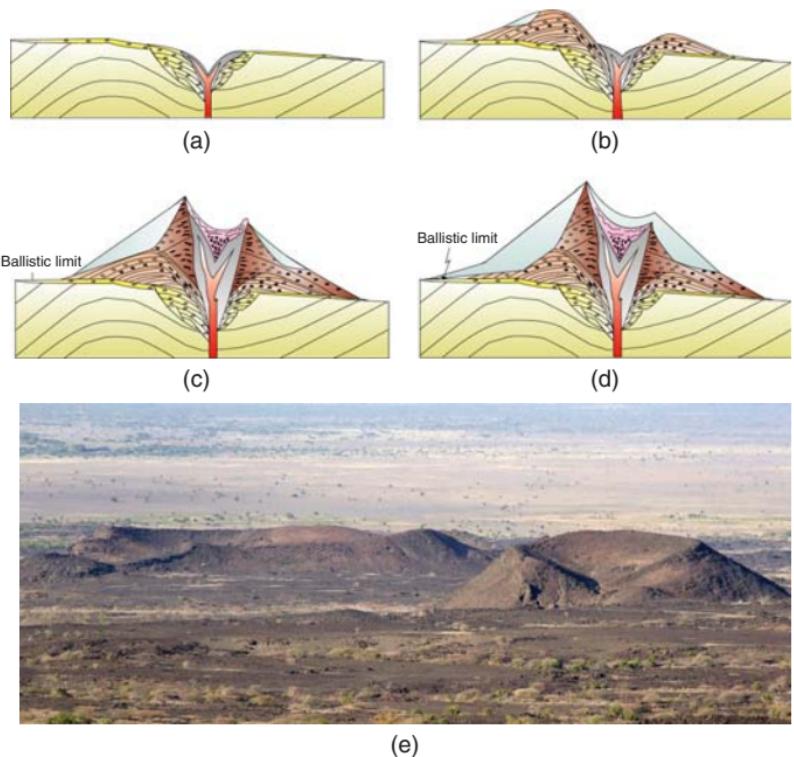


Figure 5.20 (a–d) Stages of scoria cone development from initial phreatomagmatic eruption (a), through phases of strombolian/phreatomagmatic intercalations (b), a final strombolian cone building phase (c) and a reworked talus slopes (d). (e) Examples of scoria cones from Afar, Ethiopia.

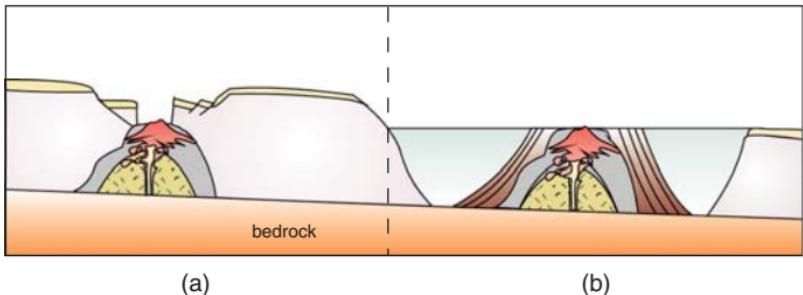


Figure 5.21 Schematic diagrams of eruption under ice. (a) Example with limited melt water and restricted eruption. (b) Melt water lake develops during eruption and emergence of the volcano leads to bedded hyaloclastite deposits in late stages.



Figure 5.22 Peperite example showing glassy highly angular lava fragments, some showing sand injections and jigsaw-like fit, SE Spain.

unconsolidated sands have been reported (for example, Figure 5.23) resulting in typical peperite textures. A close inspection of the fragmented magma and sediment type is needed when faced with candidate peperite rock types, and you must be able to convince yourself that (i) the magma involved is hot juvenile and molten and (ii) the sediment is unconsolidated, at the time of the formation of the brecciated texture.

5.5 Epiclastic Deposits

These refer to the secondary re-working of volcanic material by normal agents of weathering and deposition/transport; for example by gravity flows, water currents, waves and wind action (for example, Figure 5.24). They range from conglomerates through to sandstones and mudstones, typical of the common sedimentary types (see Tucker, this book series), but are characterised by being composed almost entirely of volcanic clasts. An epiclastic deposit is defined as having less than 25% of primary volcanic clasts. These sediments can be identified by the clast population being modified during transport, for example rounding of clasts, and by the reduction in primary glass and pumice clasts which weather and erode easily. One may also look for the mixture of multiple sources of volcanic type and other indicators such as fossils.



Figure 5.23 Peperite example from intermixed aeolian sand and lava fragments. Glassy (now weathering orange/red) bulbous and irregular feather like contacts, with the sand highly indurated (baked), Etendeka, Namibia.

A key sub-type of epiclastic deposit are *lahars*. These are formed by hot-cold debris flows which can occur with the reworking of hot mixtures during or very shortly after eruption, or by reworking of cold poorly-sorted material, commonly by water sourced by rain or melted ice. A simple classification of volcaniclastic mass flows is given in Figure 5.25, showing the continuum between debris avalanche and lahar type deposits. The deposits frequently consist of large volcanic fragments in a finer ashy matrix which is commonly coarser than mud-grade (Table 5.1) (Figure 5.26). The Indonesian term *lahar* is preferable for such water-deposited poorly-sorted deposits of volcanic materials. Lahars are also variable in terms of the proportion of juvenile and non-juvenile volcanic material. Fragmented volcanic materials deposited from a cool water-rich flow are also sometimes referred to as *mudflows*. Eruptions of volcanic material through lakes, below ice or during heavy tropical rain may generate lahars composed largely or entirely of juvenile volcanic material. Conversely, non-juvenile material may be erupted and there may, therefore, be a transition from hot (pyroclastic) lahars to water-rich *debris flows* formed on the slopes of volcanoes. These are especially common where deposits form in shallow-water conditions

VOLCANICS 2 – PYROCLASTIC ROCKS



Figure 5.24 Example of re-worked epiclastic boulder bed below a lava flow, from Iceland. Rounded clasts of coarse pebble/boulders are interbedded with coarse epiclastic sandstones.

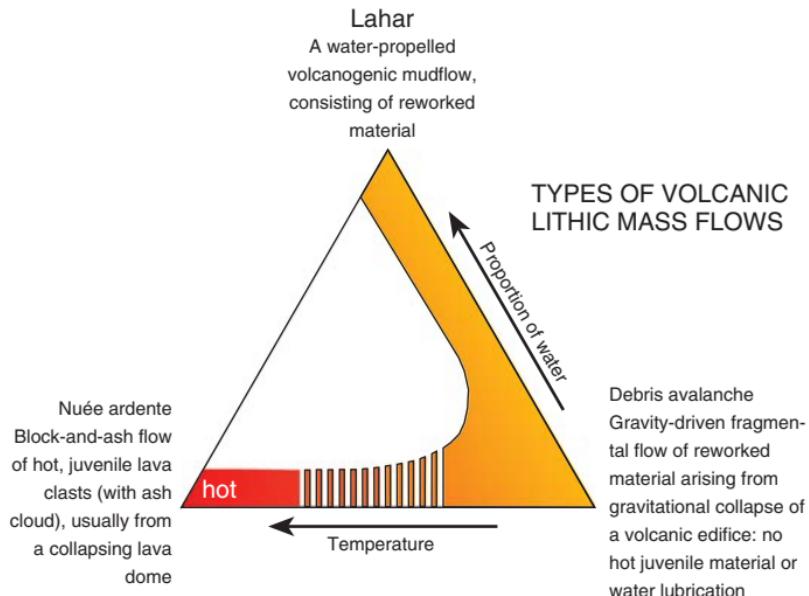


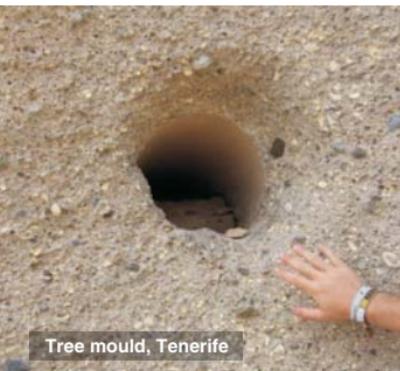
Figure 5.25 Continuum between debris avalanche and lahar. Hotter dome collapse resulting in block and ash flows (nuée ardente) is also indicated.



Figure 5.26 Debris flow/lahar type deposit of sub rounded polymict (mixed) clasts supported by a finer reddened matrix of coarse to fine epiclastic sand and mud, Arran, Scotland.

subject to volcanic instability and seismic activity. These include the cold lahars described above, which commonly contain angular blocks of partially lithified fragments of a uniform pyroclastic rock type. Such deposits do *not* result directly from volcanic activity but reflect disruption and flow of partially lithified pyroclastic material. Such material may not be strictly volcanic in origin and hence may be termed sedimentary-slump deposits (cf. Tucker, this book series).

VOLCANICS 2 – PYROCLASTIC ROCKS



pyroclastic volcanics - Examples

SHALLOW-LEVEL INTRUSIONS

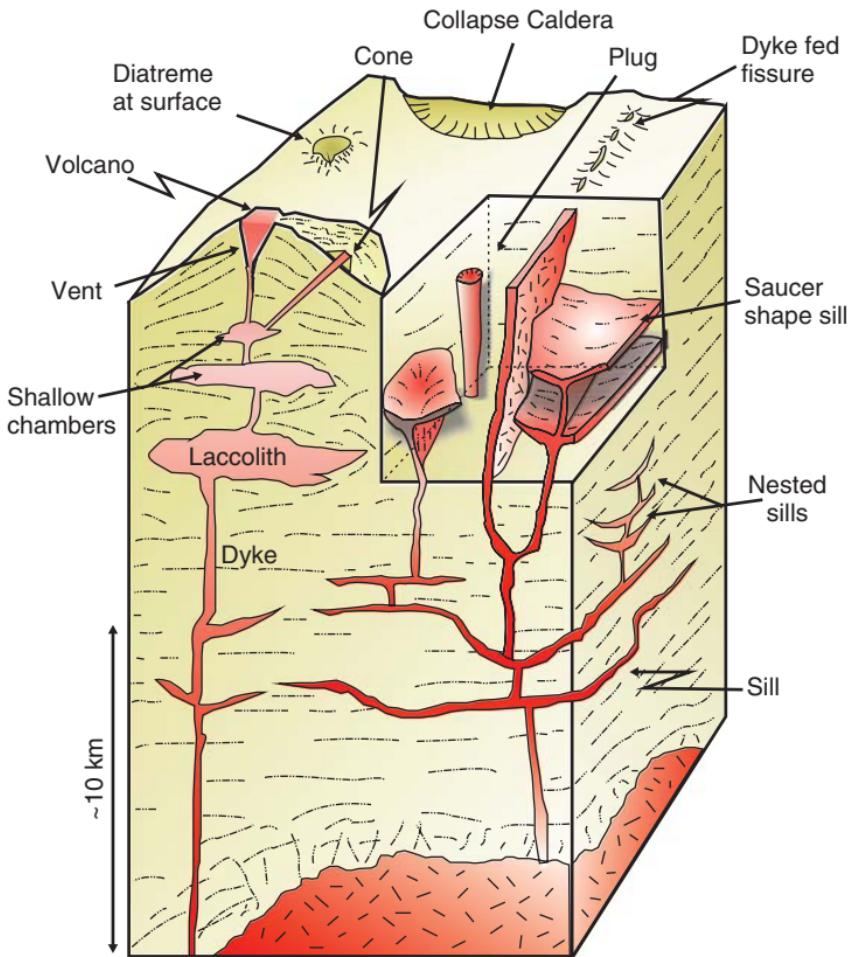


Figure 6.1 Schematic diagram of magma plumbing systems beneath volcanoes. Intricate networks of dykes and sills make their way from depth to the surface. Shallow chambers form laccoliths beneath volcanoes where magmas can mix and be recycled. A variety of small plugs, sheets and vent facies occur very near the surface.

6

SHALLOW-LEVEL INTRUSIONS

As magma makes its journey from its source (for example, mantle melting) to the surface it must travel through a pathway we may consider as the ‘magma plumbing system’ (Figure 6.1). This pathway may be simple or complex and at any stage in the field you may be looking at a particular aspect of such a system depending on the depth of erosion and/or the relationships of the igneous rocks you are considering.

When faced with intrusions in the field you will need to record some key valuable observations to help interpret their relationships:

- Contact relationships with country rocks; is this sharp-irregular, is there a chilled margin, are we intruding through sediments-metamorphic or even igneous host rocks?
- Contact relationships with other intrusions; cross-cutting relative age information, contacts (again sharp or irregular)?
- Internal textures; are there crystals and matrix. If there is a clear crystal population what are its textural features (crystal size, shape, internal zoning, and so on), what is the size of the crystalline matrix, how do textures vary from contacts into the intrusions, is there any evidence for vesiculation, and so on. Are there more than one magma present, for example magma mingling (see Chapter 9).

Shallow level intrusions also take on many forms which will become apparent as you map them out (see Figure 6.1). The most important forms are; Sills, Dykes; Laccoliths and plugs.

In this section we will consider the main types of shallow level intrusions and the key field relationships you will need to record in order to interpret them in the context of the magmatic system.

6.1 Sill and Dykes

The commonest form of minor intrusions are termed *sills* and *dykes* (Figures 6.1 and 6.2). A dyke (US, dike) is a tabular intrusion that cuts across horizontal or gently-dipping planar structures, such as stratification, in the surrounding rock.

SHALLOW-LEVEL INTRUSIONS

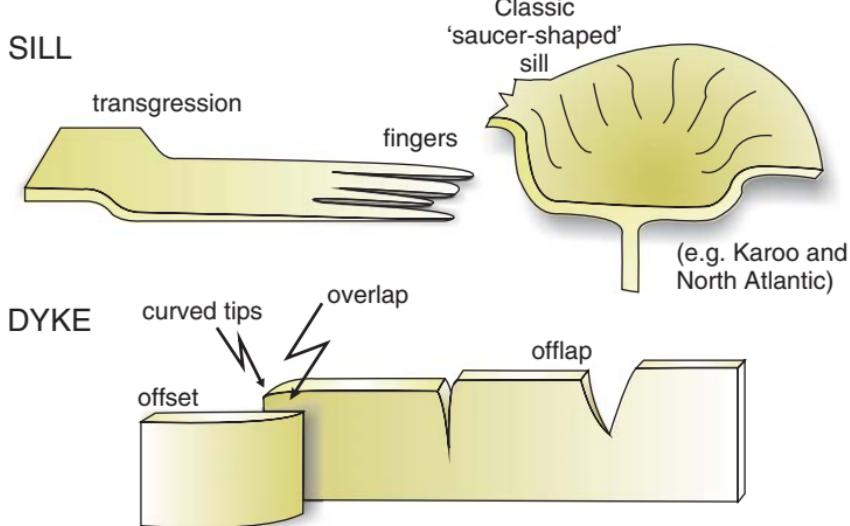


Figure 6.2 Idealised three-dimensional forms of dykes and sills. See text for discussion.

It therefore has the same form as a vein (described in Chapter 2). A sill is a tabular intrusion that parallels such horizontal or gently-dipping planar structures. Dykes are therefore discordant to structures in the country rock whereas sills are generally concordant between beds of layered rock but may show small side transgressions and/or larger dyke-like transgressions (Figure 6.2).

These definitions raise complications; in deformed terrains it may be difficult or impossible to determine whether a given tabular intrusion was emplaced as a dyke or as a sill (however, extensive sills are commonly emplaced only within well-stratified sedimentary sequences). Many extinct and eroded volcanic areas are characterised by the occurrence of an intricate arrangement of dykes and sills and, in such cases, these minor intrusions form the plumbing system to the volcano channelling the magmas that were erupted. The complex plumbing underneath many volcanic systems allows the separate evolution and mixing/recycling of magma batches, which can be recorded in the eruptive products, in crystal populations within the intrusions and by the juxtaposition of different magma intrusions.

Minor intrusions have mineral compositions corresponding to most other igneous rock types. They range from basalt through andesite and dacite to rhyolite in composition and may be named after the volcanic equivalents (cf. Tables 3.5 and 4.1). Thus a fine- or medium-grained minor intrusion of

SHALLOW-LEVEL INTRUSIONS

mafic mineral composition may be termed a *basalt*, for example *basaltic dyke*, although the terms *dolerite* and *diabase* are also used. Diabase implies a particular intergrown textural relationship between plagioclase and pyroxene (often observable only in thin section and termed *ophitic* texture) and is synonymous in US usage; however, in the UK the term diabase is rarely used, but may be sometimes used for a dolerite in which the feldspars and mafic minerals are highly altered. Intermediate and acid dykes are termed finegrained diorite, granodiorite and granite, and, if they are aphanitic, may be termed andesite, dacite and rhyolite as described for lava flows. Glassy aphanitic minor intrusions may be termed pitchstone or obsidian as defined in Section 4.2. Although most minor intrusions may be matched in composition with common rock types described earlier, a rather uncommon group, the lamprophyres (see also Chapter 3) which are distinct in mineral composition and texture from normal plutonic and volcanic rocks, also form minor intrusions.

Sills (for example, Figure 6.3), range from a few centimetres to hundreds of metres in thickness, although the commonest sills are between 1 and 20 m in thickness. Commonly sills occur as multiple intrusions forming ‘nested’ sill complexes (for example, Figure 6.1). They may show attenuated extremities which separate into elongate ‘fingers’ (Figure 6.2). Sills may be easily confused with basaltic lava flows (especially where both occur in the same sequence). However, close inspection will reveal chilled margins on the top and the bottom of sills and they will lack the surface characteristics of lavas (as described in Chapter 4).

Dykes (for example, Figure 6.4) like sills, usually vary from a centimetre to many metres, or even hundreds of metres, in width but large elongated discordant



Figure 6.3 Thick sills (dark brown) in Karoo sediments, Dry Valleys, Antarctica.



Figure 6.4 Basic dyke cutting through sediments in Arran, Scotland. Note how dyke steps, influenced by the joint/faulting pattern of the rock it is intruding through. Blown up picture shows flow alignment of vesicles in the centre of the dyke (cf. Figure 6.10b).

mafic intrusions with width measured in terms of kilometres may also be termed dykes. Dykes that radiate away from individual igneous centres are termed radial dykes, and, where such dykes (and those that cannot be related to such centres) are concentrated parallel to a regional tectonic trend, they are termed dyke swarms. When traced along strike, an individual dyke is rarely continuous. It may show discontinuity at outcrop, termed offlap, and may locally be offset in such a way that parts of the dyke show overlap (Figure 6.2).

Although most dykes and sills are broadly linear features, some are related to central igneous complexes characterised by sheet-like intrusions that may be circular in outcrop plan. These are termed *ring intrusions* and have been classified into two types according to the attitude of the contacts (Figure 6.5). *Ring-dykes* have vertical and outward dipping contacts which are usually taken to indicate emplacement by *subsidence* of the central cylindrical block (Figure 6.5a) in a tensional stress field. Ring-dykes may have thicknesses of up to 1–2 km but

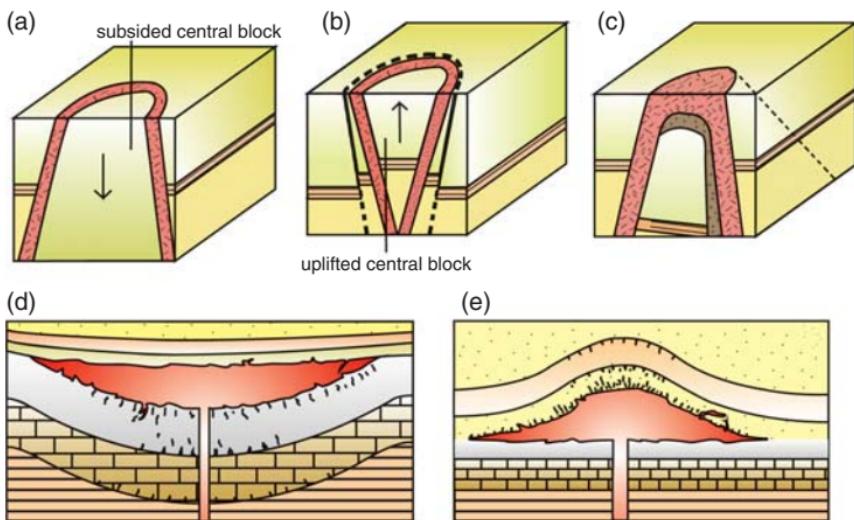


Figure 6.5 Schematic diagrams of circular shaped shallow intrusions. (a) A single ring-dyke; (b) a set of cone-sheets, idealised to show the features of shape, outcrop and relationship to horizontally – bedded country rock; (c) block diagram of two minor intrusions emplaced as a result of subsidence of a central block; (d) cross-section through lopolith; and (e) cross-section through laccolith.

are more commonly in the order of several hundred metres. They may occur singly or in multiple units, and the overall ring complex may be 5–10 km in overall diameter. Cone-sheets occur as single small intrusions, with outcrops separated by curved sheets termed ‘screens’ of country rock. By contrast to *ring-dykes*, the form of cone-sheets must indicate emplacement by *uplift* of the central conical block and this indicates uplift of the roof of the chamber during intrusion. Ring-dykes are much more abundant than cone-sheets, but the latter have been originally mapped out as common features of the Scottish Tertiary volcanic province. Because the mechanism of emplacement of ring-dykes involves subsidence of a central block (Figure 6.5a), such ring intrusions are sometimes associated with relatively flat-lying tabular minor intrusions emplaced over the block (Figure 6.5c). Ring-dykes and cone-sheets may be medium- or coarse-grained in texture and are more commonly of mafic rather than felsic mineral composition. It should be noted that recently many of the circular bodies attributed to be either ring dykes or cone sheets have been shown to be more like lopolith/laccolith type intrusions (for example, Figure 6.5d,e, see also Section 6.4), where the tips of the intrusion have been exposed at the surface, and careful mapping of circular type intrusions is required to fully assess the type of intrusion present. Where country rocks have been blistered up by the

SHALLOW-LEVEL INTRUSIONS

inflation of a shallow intrusion a laccolith results (Figure 6.5e), which again are commonly circular in plain section.

6.2 Working Out Emplacement History

Many of the sheet-like intrusions have been emplaced by the amalgamation of fingers as the intrusion propagates into the country rock (Figure 6.6). In section,

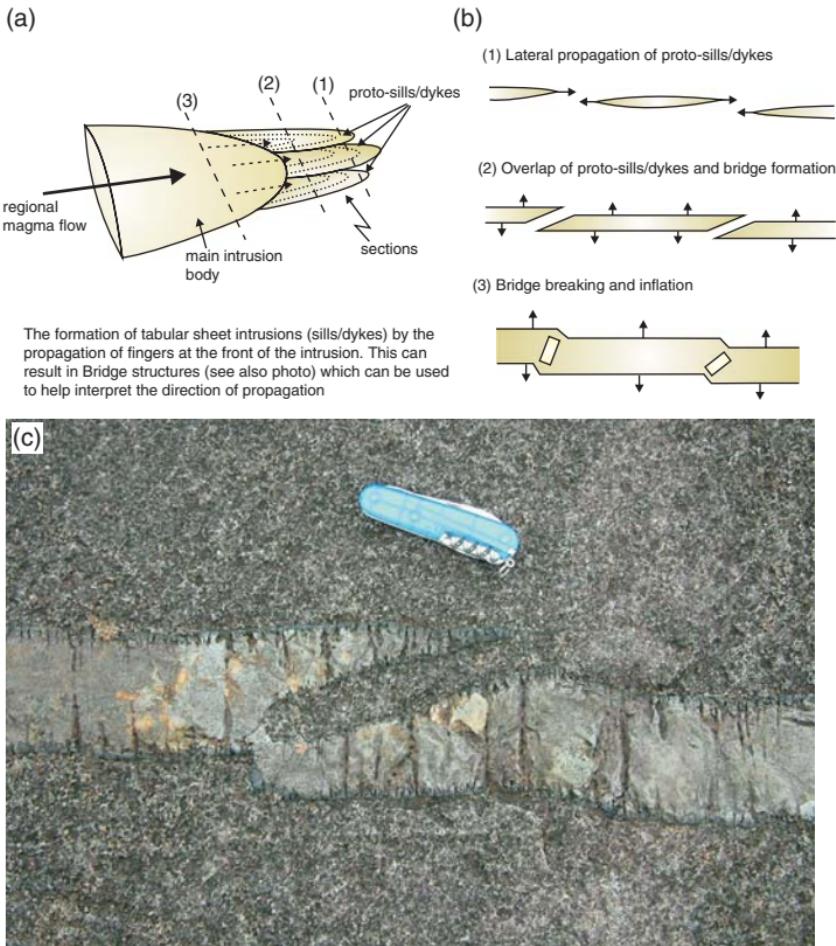


Figure 6.6 Emplacement of igneous sheets. (a) Formation of tabular sheets by propagation of fingers. (b) Stages of linking up of fingers in section. (c) Thin basalt dyke cutting into gabbro with bridge structure, Isle of Skye, Scotland (Note chilled margin and conjugate tips).

SHALLOW-LEVEL INTRUSIONS

you may be able to find bridge structures which will give a sense of the emplacement direction of the intrusion (for example, Figure 6.6b,c). For minor intrusions, it is important to study the igneous margins and the contact with the country rock. Contacts of minor intrusions are usually chilled and hence are fine-grained or glassy in character (for example, Figure 6.6c). They may provide a clue to the state and composition of the magma emplaced to form the minor intrusions; for example the volume and composition of the phenocryst phases indicates the proportion and type of crystals present within the magma at the time of intrusion. The grain-size variation across the intrusion may provide information concerning the cooling of the intrusion; a large variation in grain-size implies emplacement of hotter magma into cooler country rocks than does a lesser degree of variation. A margin that is not chilled implies that the intrusion was emplaced into warm or hot country rocks. The country rocks around minor intrusions frequently show the effect of thermal metamorphism. For most minor intrusions, this is limited to within a metre, or at most a few metres, of the contact. The metamorphic recrystallisation and the minerals in the contact rocks should be noted; for example shales may be baked and hardened, limestone may be recrystallised to contact metamorphic marble and rocks rich in organic material may be carbonised. Details of such metamorphism are given in Chapter 7, Table 7.2.

The relative ages of different groups of minor intrusions may often be determined from contact relationships. Thus, if more than one kind of minor intrusion occurs, for example as discerned by composition and/or geographic trends, then the outcrops should be studied for intersections, as shown in Figure 6.7. In such cases several episodes of injection by the same kind of minor intrusion may be demonstrated (cf. Figure 6.7d). However, care should be taken not to confuse intersecting dykes with branching dykes (Figure 6.7b). Where contacts are not exposed, the relative ages of minor intrusions may sometimes be determined from identification of xenoliths of one type within another. Also, minor intrusions can be related to regional tectonic activity, such as folding and faulting, and the relative ages should also be determined (for example, Figures 6.7 and 6.8). The emplacement of sills into well-bedded strata may also lead to complex field relationships which may easily be misinterpreted. For example, it is sometimes easy to confuse transgressions by sills with faulting (Figure 6.8). The absence of offsets in the beds above and below the sill (Figure 6.8b) confirms that this represents a simple transgression, whereas Figure 6.8c shows a sill (and the enclosing rocks) offset by faulting. More complex situations result, for example from intrusion of a sill (X) followed by intrusion of a second sill (Y) (Figure 6.8d).

Although most minor intrusions result from injection of a single pulse of magma, successive intrusion of magma into the same intrusion may lead to formation of more complex minor intrusions. If new magma of similar composition

SHALLOW-LEVEL INTRUSIONS

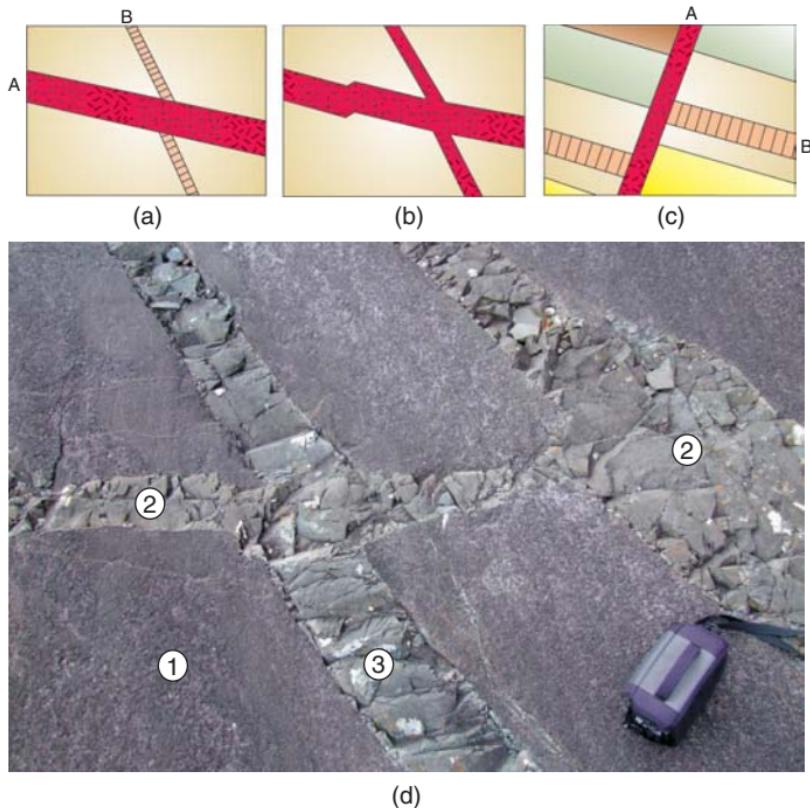


Figure 6.7 (a) Intersecting dykes, A being younger than B. Note that the apparent displacement of B results only from dilation by emplacement of A, and from lateral movement within the younger dyke, (b) Branching dykes, of the same age, (c) Section of dyke, A, which has entered a fault. The fault intersects a series of strata containing a sill, B. (d) Filed example of cross cutting dykes through a gabbro intrusion, numbered to indicate relative ages, Isle of Skye, Scotland.

is emplaced into the still-molten interior between the chilled margins of an earlier intrusion, this is termed a multiple intrusion. The formation of multiple dykes may be favoured by emplacement at sufficient depth for cooling to be sufficiently slow for dykes to be partially molten at the time of emplacement of the next intrusion. A particular association of multiple dykes, termed a sheeted complex, forms part of ophiolite complexes, and these are described in detail in Chapter 8. In rarer cases, a multiple intrusion might be composed of two contrasted magma types. These are termed composite dykes or sills. An example

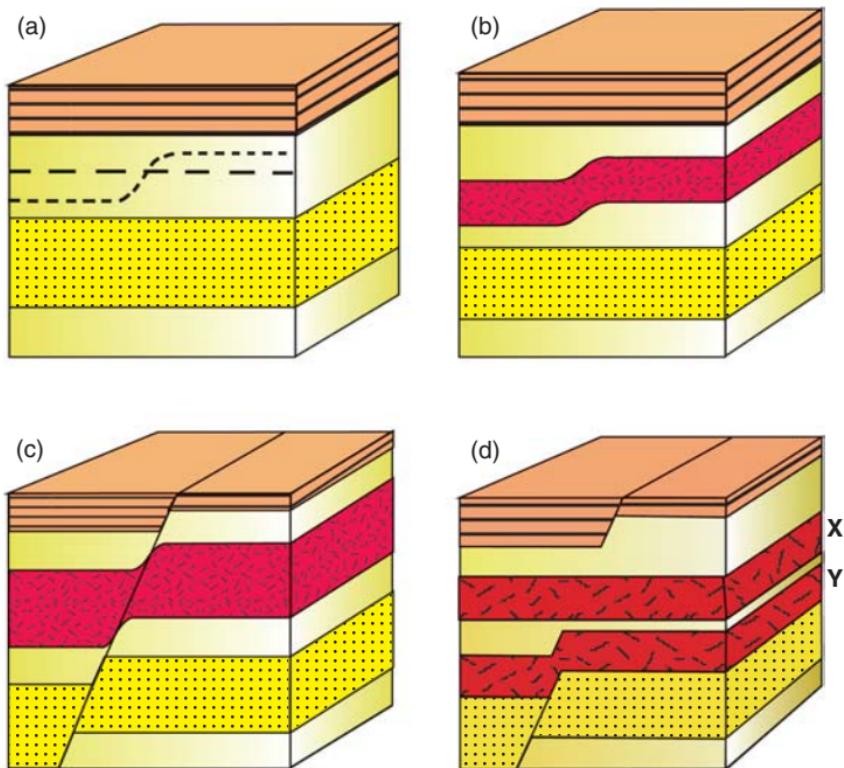


Figure 6.8 A sequence of strata (a) showing the effect of intrusion of a locally transgressive sill (b) or a completely concordant sill which is later displaced by faulting (c). The position of the transgressive sill (b) is shown by the short dashed lines in (a). In each case, the final disposition of the sill is similar (compare (b) and (c)) but the displacement of strata above and below the sill and the possible identification of deformation (e.g. fault brecciation) along the fault plane enables (c) to be distinguished from (b). (d) Sections of sills that have been emplaced into faulted strata. X is a sill that has been emplaced across the fault plane, whereas Y is a sill emplaced at a particular stratigraphic horizon and which exhibits transgression at the fault plane.

of an early intrusion of basic magma that has been followed by an intrusion of acid magma is given in Figure 6.9; the contact between the two types varies between very sharp and completely diffuse. In composite intrusions, the basic and acid components might contain xenoliths or xenocrysts of each other. Homogeneous minor intrusions might contain inclusions of contrasted composition, for

SHALLOW-LEVEL INTRUSIONS

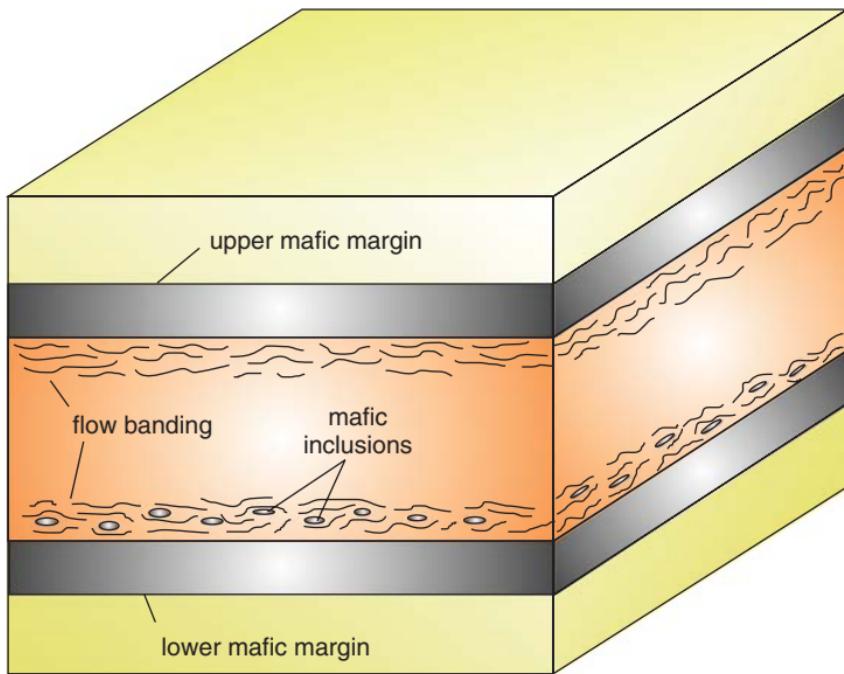


Figure 6.9 Diagrammatic section of a composite intrusion (grey = mafic, pink = felsic).

example basalt or andesite inclusions within dacite or rhyolite, which indicate magma mixing (see also Chapter 9).

In addition to compositional variation in the main rock-type(s) within minor intrusions, variations in the type and distribution of phenocryst phases can be important. For example, phenocrysts may be concentrated in the centre of a minor intrusion which has fine-grained aphyric margins (Figure 6.10a), reflecting flowage differentiation in which phenocrysts concentrated in the centre of the intrusion reflect more rapid magma flow compared with that at the margins. However, such cases might also reflect composite intrusion of aphyric and porphyritic magma (Figure 6.10c), so it is important to examine carefully the texture of possible contact zones in such heterogeneous intrusions in order to distinguish between these possibilities. Evidence for a sharp (or chilled) contact between the inner (porphyritic) intrusion would favour interpretation by multiple intrusion, although the magmas may have been related by fractional crystallisation below the depth of emplacement. Where vesicles existed in shallow emplacement, their deformation can also be used to help work out emplacement direction (for example, Figure 6.10b, cf. Figure 6.4).

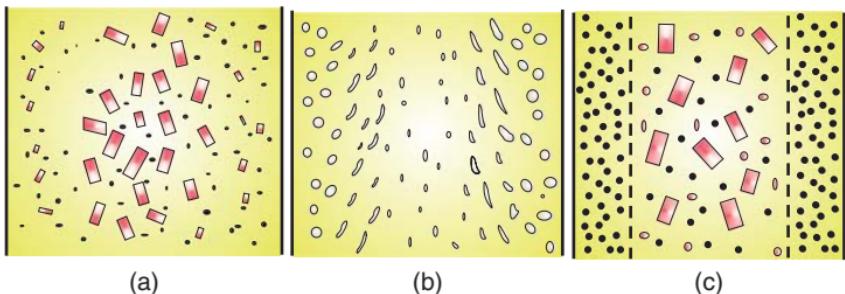


Figure 6.10 (a) Concentration of phenocrysts in the centre of a dyke as a result of flowage differentiation. (b) Vesicle deformation with flow (cf. Figure 6.4). (c) A composite dyke with an aphyric margin sharply separated from a porphyritic centre.

Besides the variations described above, sills, in particular, show other types of mineralogical variation as a result of their horizontal tabular form. These include variation as a result of crystal settling (cf. Chapters 2 and 8) which occurs to a noticeable extent within some thick sills, generally exceeding 50 m in thickness. Such sills may have a layer or layers rich in dense minerals such as olivine, pyroxene and Fe-Ti oxides near to the base (cf. Chapter 8). These layers overlie the fine-grained lower marginal layer and may be complemented by layers relatively poor in these minerals towards the top of the sill, below the upper chilled margin. However, it may be extremely difficult to distinguish between such mineralogical differentiation of a single magma, and that resulting from multiple or composite intrusion. For example, fine-grained olivine and/or pyroxene-rich dolerite may intrude aphyric dolerite. Distinction of such origins may require detailed chemical and mineralogical study of the sill. In other examples, the intrusion of two or more distinct magmas can result in quenched (blocky) textures of the hotter mafic magma against the cooler felsic (for example, Figure 6.11). Sometimes one finds that a new generation of magma, perhaps due to renewed magmatism at the site of an old intrusion, has cut slightly older but related igneous rocks. A particular example of this might be where a granitic magma has invaded a rectangular joint set in a previous mafic (that is, generally dark-coloured) igneous body. This is known as *net-veining*. As mafic magmas generally crystallise at higher temperatures than felsic magmas, a stage in the crystallisation of a mafic magma body may be reached at which it causes local melting of the host rocks to produce granitic liquid. This liquid may then penetrate joints within the cooling mafic body and its surrounding, a process known as *back-veining*.

As a result of fractional crystallisation and volatile concentration during crystallisation, many sills are characterised by segregations of coarser grained and/or more felsic material in the upper part. These include sharp-sided veins, irregular



Figure 6.11 Intrusion breccia of more mafic (angular blocks) into a more felsic host, formed by magma mingling in shallow sheet intrusions Ardnamurcan, Scotland.

segregations and irregular diffuse streaks termed *schlieren*. They vary from fine-grained aplites through medium-grained (including granophyre) to coarse-grained texture and may have large crystals (up to 10–20 cm long), particularly of pyroxene, amphibole and Fe-Ti oxide. They are then termed *pegmatites/pegmatitic schlieren*. The composition of segregations in sills generally reflects the composition of the crystallising magma. Felsic segregations in quartz-bearing basalt/dolerite sills are fine-grained diorite, granodiorite and granophyre whereas those in quartz-free, olivine basalt/dolerite sills are often quartz-free gabbro or diorite, containing zeolites and/or foid minerals. Such pegmatites can be very important to help age date such intrusions.

6.3 Volcanic Plugs and Diatremes

Erosion of composite volcanoes may reveal subcircular exposures of fine-grained rocks that may be interpreted as *volcanic plugs*. Such plugs are also termed volcanic pipes, and are generally a few hundred metres in diameter, but may extend up to a kilometre. The three-dimensional form of volcanic plugs is generally cylindrical. Most plugs are composed of shallow intrusions/lavas, but a

SHALLOW-LEVEL INTRUSIONS

small proportion consist of pyroclastic material; these are described separately below. The lava within plugs may be of various grain sizes and compositions; it often shows alteration and deposition of secondary minerals as a result of hydrothermal activity concentrated around the cooling volcanic vent. Most plug rocks show features such as brecciation (that is, an unsorted mixture of angular fragments) as a result of the passage of gases and hydrothermal solutions through them. Such brecciated rocks are termed *volcanic agglomerate* or *brecchia*. Many small pyroclastic volcanoes are underlain by plugs or pipes composed of pyroclastic material. These are termed *diatremes*, and include *breccia-pipes*, *tuff pipes* and *kimberlites*. Diatremes are narrow, steep-sided cylindrical or funnel-shaped pipes which have intruded through the crust, and consist of solid fragments of crustal and mantle rocks with or without solid fragments of juvenile rock, apparently emplaced at a low temperature. Diatremes may be composed mainly of comminuted crustal rock or juvenile material. Some diatremes have erupted at the Earth's surface as ash- or tuff cones, others can be shown not to have reached the Earth's surface but to have been revealed subsequently by erosion and are then termed 'blind pipes'. The crustal fragments in diatremes range from fine ash to blocks several metres in size. Juvenile fragments are generally glassy ash and pumice but may include crystalline material.

In natural exposures of diatremes, the walls of the vent are usually visible over only 100–200 m; most diatremes seem to converge downwards like steep-walled funnels. The rocks forming the diatreme may be bedded as a result of infall of products of successive eruptions and frequently they can be matched with rocks in the surrounding walls. Where diatremes were emplaced within a well-defined stratigraphic sequence, matching of such fragments may indicate their upward or downward movement. Commonly the fragments appear to have subsided with the diatreme with layers of tuff. Such subsidence may be up to several hundred metres (or, rarely, 1500 m) in vertical extent. The tuff and breccias comprising such diatremes are termed *tuffisite* and *intrusive breccia*, respectively, to distinguish them from extrusive rocks.

A distinctive rock that occurs in diatremes and occasional dykes is kimberlite (Figure 6.12), which is of economic interest since many kimberlites contain diamonds. Kimberlite is a serpentinised and carbonated mica-peridotite of porphyritic texture, containing nodules (that is, xenoliths) of ultrabasic rock-types characterised by high-pressure minerals such as pyrope (garnet) and jadeitic diopside (pyroxene). They are dark-green to dark-blue grey, and are often extremely brecciated with large crystal or rock fragments within a fine-grained matrix. Kimberlite pipes include an extraordinarily varied assortment of xenolithic inclusions; for example angular blocks of the country rock, rounded blocks of high-grade metamorphic rocks, such as might form the lower continental crust, and rounded blocks of ultramafic rocks composed of varying combinations of olivine, pyroxene, garnet (pyrope), spinel, Fe-Ti oxide and

SHALLOW-LEVEL INTRUSIONS

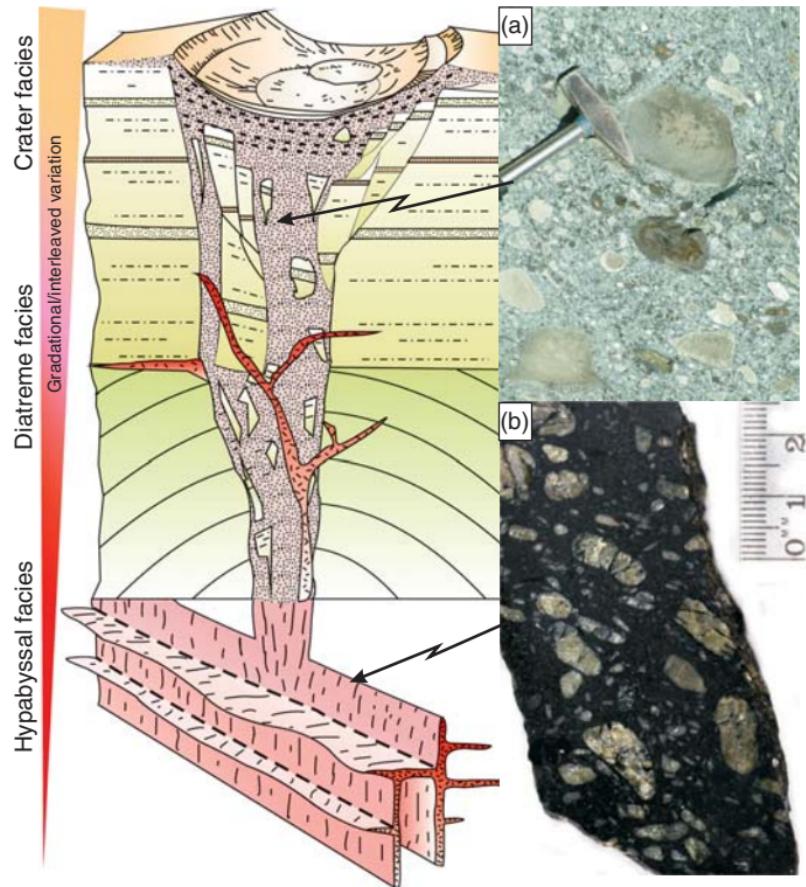


Figure 6.12 Idealised section through a kimberlite pipe, showing proposed facies and real examples: (a) A highly altered kimberlite breccia from Letseng diamond mine, Lesotho, Africa. (b) Hypabyssal/dike example from Bellshank, South Africa (photo courtesy of Graham Pearson).

mica (phlogopite). These rocks may be named according to the scheme shown in Figure 3.14 and are believed to represent samples of the upper mantle.

It has been suggested that there are key facies in kimberlite pipes (*Hypabyssal*, *Diatreme*, *Crater*, for example, Figure 6.12), and indeed the nature and form of kimberlites has in the past been driven by the earliest examples found which were pipe like bodies. It should be noted, however, that more recent discoveries of kimberlites show a much more varied form of intrusive and extrusive varieties, and careful mapping has shown that the different facies types can be mixed together at different levels within a single kimberlite.

SHALLOW-LEVEL INTRUSIONS

6.4 High-Level Subvolcanic Intrusions

Although some minor intrusions, and many plutonic intrusions, show no clear geological relationship with surface volcanic activity, many eroded volcanic areas reveal small intrusions of fine-grained diorite, granodiorite and granite composition that were probably subvolcanic in character and which may be termed high-level intrusions. Such intrusions may be approximately circular or elliptical in outcrop with a diameter of 1–4 km, or might form arcuate outcrops that may have been emplaced by subsidence of an overlying block as part of a ring intrusion. A classic example of a sub-volcanic system is exposed on the Ilse of Elba, Italy (Figure 6.13) where a range of laccoliths and small plutons of various (mainly granitic) composition are found in an inter-connected system.

High-level subvolcanic intrusions are characterised by association with contemporaneous volcanic rocks, sharp chilled contacts and a rarity of features such as pegmatites and aplites that are generally associated with intrusions at deeper levels. They may also have small cavities up to several millimetres in size lined by inward projecting euhedral crystals of the host intrusion, or of hydrothermally

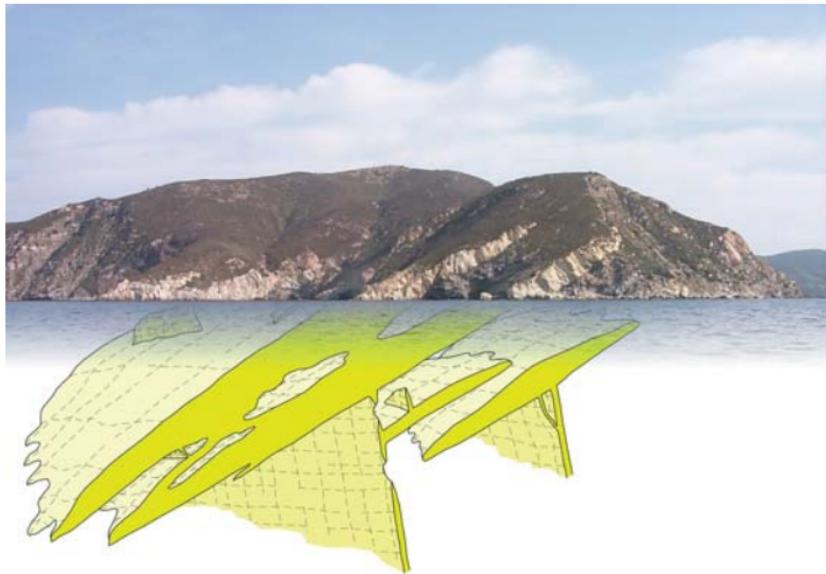


Figure 6.13 Coastline exposure and interpreted sub-volcanic system beneath Elba, Italy. Nested shallow intrusions like these are typical of sub-volcanic feeders (courtesy of Sergio Rocchi).

SHALLOW-LEVEL INTRUSIONS

deposited crystals such as quartz, feldspar and zeolites. These are interpreted as vapour cavities, analogous to amygdales and are termed *druses* or *drusy cavities-miarolitic cavities*. Rocks containing such cavities are termed miarolitic rocks. The presence of drusy cavities indicates exsolution of a fluid phase under low confining pressure and hence provides evidence of crystallisation shallower than a depth of about 1–2 km.

High-level intrusions may contain the range of xenoliths and inclusions described in Section 2.4.3 and these are sometimes more conspicuous near the upper margins than in deeper-level intrusions. In particular, high-level intrusions often contain sharp-sided angular xenoliths which are absent in deep-level plutonic intrusions. It is often difficult to establish the type and degree of connection of high-level intrusions with the surrounding volcanic rocks, and therefore careful study should be made of geological features such as xenoliths, inclusions and contact characteristics that might contribute to the establishment of such relationships. Finally it is possible to characterise the overall shape of the intrusions on a log-log length vs thickness diagram (Figure 6.14) which shows a characteristic S shape.

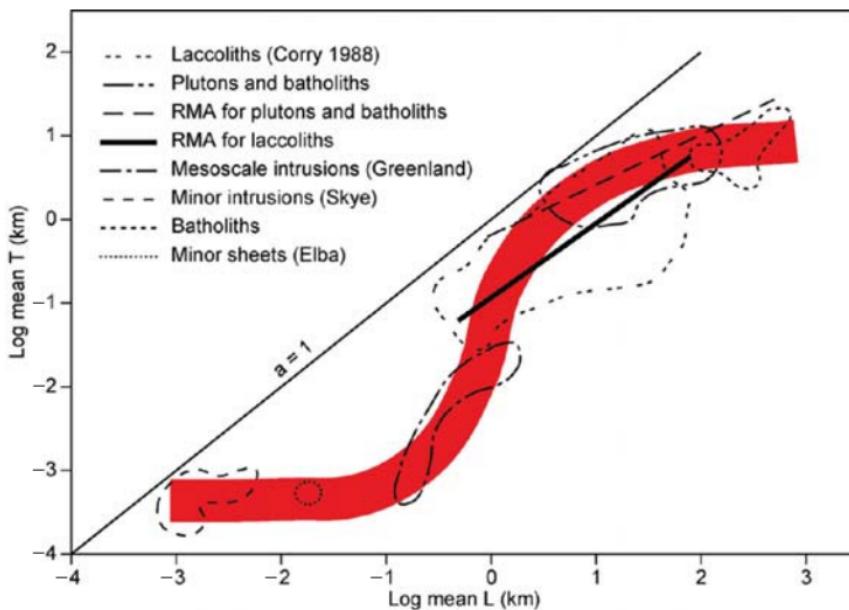


Figure 6.14 Relationship of length with thickness for different scales of intrusion. Deviations from the $a = 1$ line show asymmetric shapes of the intrusions (i.e. not balloon shape).

SHALLOW-LEVEL INTRUSIONS



Dyke, Archaean gneiss, Scotland



sill/laccolith Elba, Italy



Dykes through pillow lavas, Cyprus



Sill, 'Kilt Rock' Skye, Scotland



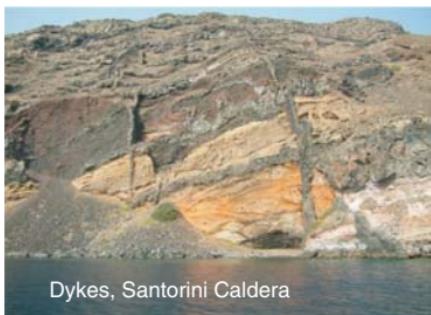
Glassy Chilled Dyke, Cyprus



Fine scale intrusions, Elba, Italy



Sills, Antarctica



Dykes, Santorini Caldera

Shallow-level Intrusions-Examples

GRANITIC COMPLEXES

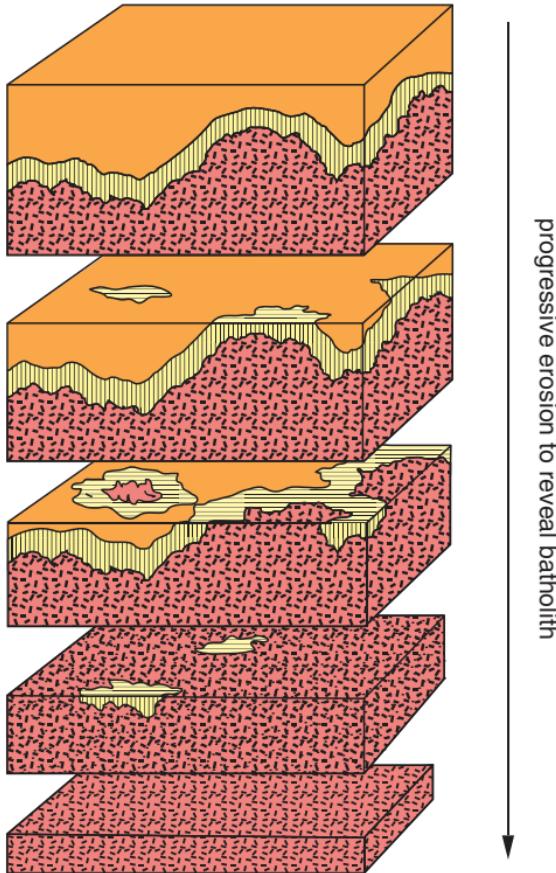


Figure 7.1 Successive stages in the exposure of batholith rocks (dash pattern) revealed by erosion of country rocks (plain) and the metamorphic aureole (straight lines).

7

GRANITIC COMPLEXES

7.1 Introduction

Granitic rocks are key components of the continental crust, and have over geological time helped keep the landmasses of the earth stable and elevated above mean sea level. Also referred to more generally as *granitoids*, they form part of a compositional spectrum ranging from basaltic (SiO_2 circa 50 wt%) to highly acidic ($\text{SiO}_2 > 70$ wt%) medium- to coarse-grained rocks (cf. Chapter 3). With increasing SiO_2 content, the plutonic rocks of this spectrum progress from gabbro-tonalite-granodiorite to granite (*sensu stricto*) and are emplaced entirely beneath the surface. Their volcanic (extrusive) equivalents form the complementary association basalt-andesite-dacite-rhyolite, sometimes referred to as BARD. Together these rock types define the *calc-alkaline* association, a distinctive petrological and geochemical typology formed at destructive plate margins (Chapter 1).

As with all plutonic rocks of the calc-alkaline association, the parent magmas cool and crystallise at varying depths inside pre-existing crust. For granites this crust is exclusively continental and this simple fact distinguishes them from other intrusive members of the calc-alkaline association, notably gabbros, which occur also within oceanic crust. One major goal in understanding how plutonic magmas were emplaced is to work out as accurately as possible their intrusion depth. Coarser grain size is one indicator of depth, but may not always be reliable as other factors including pressure quenching and water content of the magma can result in sudden rapid growth of large crystals. The relative amounts of intrusive rock and country rock available for study depend strongly on the level of erosion (Figure 7.1).

7.2 General Features and Occurrence

Most of the rocks in the calc-alkaline association are *granitoids*, that is they are granites, granodiorites and tonalites all containing $>20\%$ quartz, with feldspar, mica and amphibole in various combinations (Chapter 3, Figures 3.10 and 3.11). They occur together with varying amounts of diorite and gabbro. Over

several decades different groups of igneous geologists have sought to classify calc-alkaline granitoids into tectonically-related groups using a range of field, petrographic and geochemical criteria. The result has been a proliferation of classification schemes using letters as shorthand for defining a particular typology the most famous and widely known being the *I-S* classification of Chappell and White. One area of outstanding contention remains the degree to which, once classified, the granite type reflects its original magma source region. It has proved possible to distinguish, using field criteria, between *I-type* (igneous source) biotite – or hornblende-biotite granitoids and *S-type* (sedimentary source) muscovite-biotite, or two-mica, granites. Other members of the granite ‘alphabet soup’ include A-types (alkaline) and M-Types. One important sub-set of the granitoid family that does stand alone on age and/or compositional grounds are the trondhjemite-tonalite-granodiorite (TTG) suites found mainly in ancient, archaean crust. Trondhjemite is a name used for leucotonalites (cf. Figures 3.11 and 3.12, Chapter 3) dominated by quartz and sodium-rich plagioclase feldspar in which alkali feldspar is less than 10% of the total feldspar and the colour index is less than 10. There are however phanerozoic examples from modern subduction settings. The origin of these analogous, high sodium plutonic rocks, whose associated extrusive magmas are commonly referred to as adakites, originate via partial melting of amphibolite or garnet-amphibolite protoliths. Although some granitoids have transitional characteristics, and others may be difficult to designate because of hydrothermal alteration, the information in Table 7.1 provides a basis for preliminary classification in the field.

The importance of calc-alkaline granitoids in continental margin arcs (e.g. those of the circum-Pacific region) implies a magma-genetic link with subduction processes at destructive plate margins (cf. Chapter 1). About 100 000 km² of imperfectly exposed Mesozoic and Tertiary batholith forms the mountain chains of the western Americas from Alaska to south Chile. Although much of this batholith is unstudied, detailed fieldwork in some areas has revealed considerable internal variation. Mapping of the Peru coastal batholith based on careful tracing of internal intrusive-intrusive rock contacts within the batholith over large distances, in this case made possible in the vertical dimension by the deeply-dissected nature of the Andes has led to the map shown in Figure 7.2. This is a geological map of a 120 × 70 km area showing some of the major lithotypes (or units) comprising the batholith. The succession of emplacement is from basic to acid meaning the plutons comprising the batholith were constructed piecemeal from relatively small discrete magma batches.

Also in this area the batholith is cut by several major high-level ring complexes and other minor shallow intrusions (cf. Chapter 6), comprising granites, tonalites, diorites and gabbros that together confirm the link between intrusive and extrusive activity during major batholith construction. At high altitudes, the roof contacts of the batholith with overlying volcanics have been studied,

GRANITIC COMPLEXES

Table 7.1 General characteristics and possible origin of I-S-A-TTG-type granitoids.

Classification	I-type	A-type	TTG	S-type
Overall petrology	Part of a broad spectrum from diorite or tonalite through granodiorite to granite	Spectrum including peralkaline granites, alkali syenites and fayalite granites	Trondhjemite-Tonalite-Granodiorite (Dacite as volcanic equivalent)	Restricted to leucocratic granites
Distinctive minerals/ chemistry	Hornblende, biotite, magnetite, sphene	Alkali feldspar dominant, riebeckite, fayalite, rapakivi structure, fluorine-rich	Amphibole, plagioclase, high Na, high ratios of Sr/Y, La/Yb, low Y	Muscovite, biotite, cordierite, monazite, garnet, ilmenite
Enclaves	Mainly magmatic enclaves	May contain all enclave types	Mafic enclaves rare in trondhjemites	Mainly metasedimentary xenoliths
Associated metalliferous mineral deposits (see Chapter 10)	Porphyry copper and molybdenite deposits; lead-zinc deposits at lower temperatures	None distinctive	'Adakite' plutons world class porphyry copper; gold	Tin-tungsten deposits
Origin	Produced by partial melting of igneous material	Partial melting of mafic underplate or product of extreme differentiation	Partial melting of mafic protolith with garnet or amphibole stable, plagioclase absent	Produced by partial melting of meta-sedimentary material
Tectonic setting	Island arcs and continental margin subduction-related batholiths	Anorogenic/post orogenic setting (cratonic), continental rift zones	Early Archean crust, phanerozoic subduction zones	Continental collision zones with overthrust terrain

GRANITIC COMPLEXES

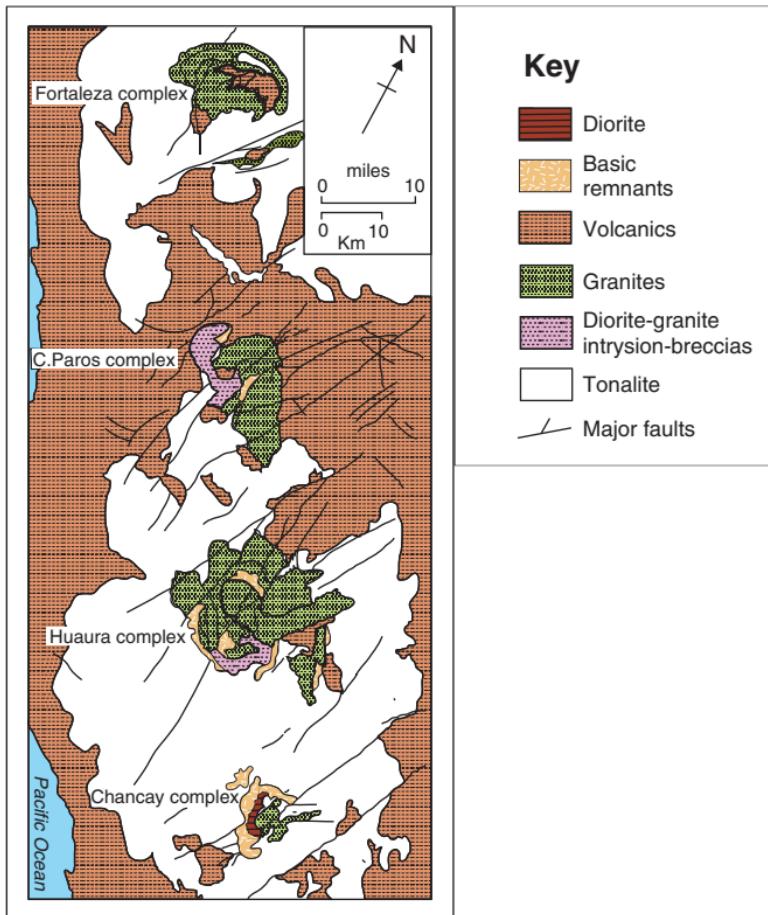


Figure 7.2 Geological map of part of the Coastal Batholith in central Peru showing the distribution of four major subvolcanic ring complexes.

whilst, at lower elevations, the different intrusive phases of the batholith have been mapped in detail, revealing a complex internal structure (see Section 7.4) but one dominated by sharp regular contacts with surrounding country rock (Figures 7.3 and 7.4). In Peru and elsewhere, individual plutons are sometimes separated by thin screens (average 10–100 m thickness) of metamorphic rocks, early mafic igneous rocks and or late aplite dykes (Section 7.5.3). Relative ages in igneous terraines of this type are determined using cross-cutting relationships such as truncated structures at contacts, enclaves of an older rock in a younger one and dykes or veins of a younger rock in an older. This complexity can result in a wide range of igneous contacts on a range of scales that need careful

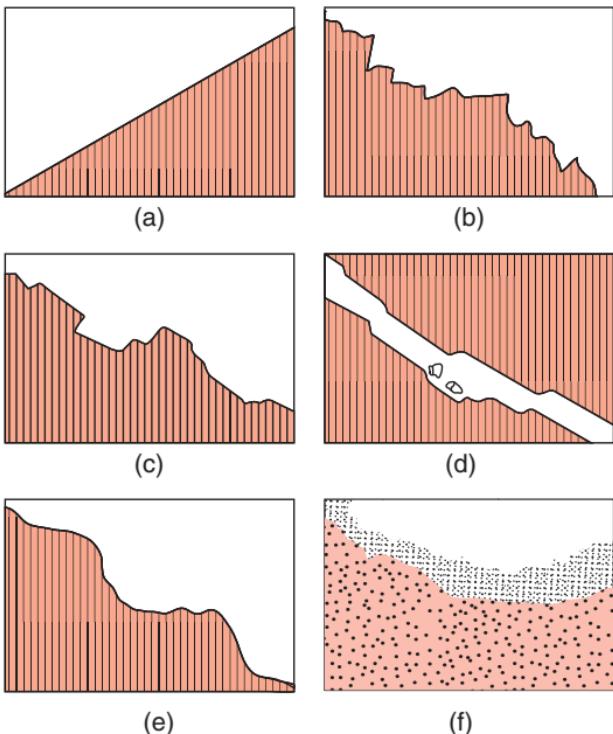


Figure 7.3 Types of igneous contact, as seen in cross-section. The intrusive rock is blank and the country rock is lined or stippled (f). The contact is straight in (a), jagged in (b), blocky in (c), blocky and matched in (d) and sinuous in (e). (f) shows a diffuse contact (dark stipple) between an igneous intrusion (blank) and the country rock (light stipple). These contact relationships may be observed on the centimetre to kilometre scale (Figures 7.4 and 7.13).

examination as they often yield important clues not just about timing but also magma rheology (see Chapter 9).

7.2.1 Pluton geometry

It was thought that the plutons comprising the Peruvian batholith (and by extension others elsewhere) extended to great depths. However, in Peru recent gravity data have shown that each pluton is a relatively thin sheet, extending downwards to depths of a few kilometres at most (Figure 7.5). Other studies from around the world have provided additional evidence in support for plutons (and by definition the batholiths they comprise) as relatively thin, broadly tabular structures as opposed to the more traditional textbook view of deeply penetrating blobs.

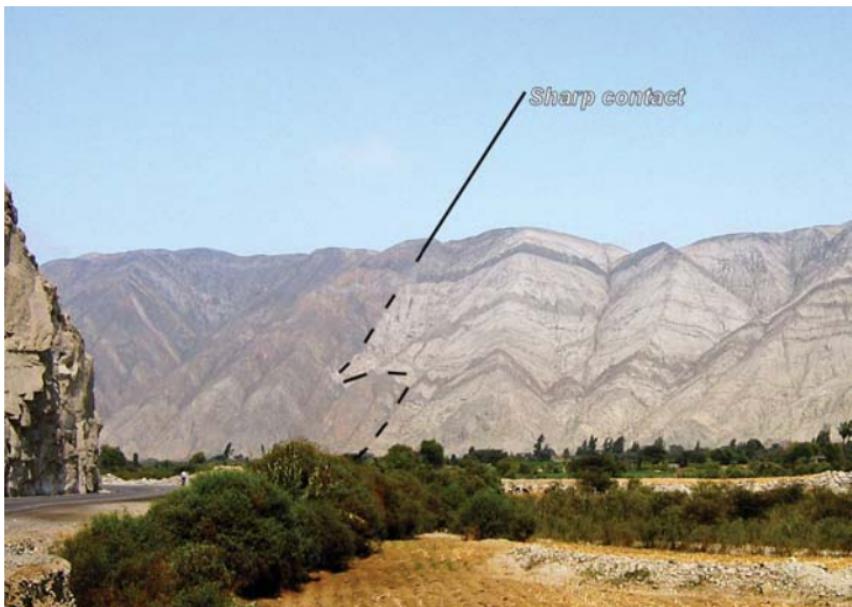


Figure 7.4 Contact between western margin of the Santa Rosa Tonalite, Coastal Batholith Peru and country rock comprising marine basalts and basaltic andesites. The contact is knife sharp and lacking in significant internal deformation, suggesting a passive mode of emplacement. The tabular intrusion is cut by syn-magmatic mafic sills and dykes that extend into the country rock.

7.3 Zoned Plutons

Zonation is seen in batholithic complexes worldwide. In the Coastal Batholith of Peru the overall composition of different plutons varies from diorite to leucogranite (generally low colour index granites with high silica content circa 70% wt. %). Although abrupt changes in rock type are usually confined to the often knife-sharp contacts between plutons, there are also less prominent variations, particularly within plutons. One excellent example of pluton zonation that has been mapped in detail is the circa 90 Ma old Tuolumne Meadows pluton in the Sierra Nevada batholith, USA (Figure 7.6). Here an outer facies of tonalite/granodiorite grades inwards into leucogranite and granite porphyry (Section 7.7.1).

Where, as in the case of Tuolumne Meadows, the composition of the pluton changes from mafic at its margins to more acidic in the core, the zonation is referred to as normal. The reverse, although less common, may also occur (e.g. Smartville Complex, California).

GRANITIC COMPLEXES

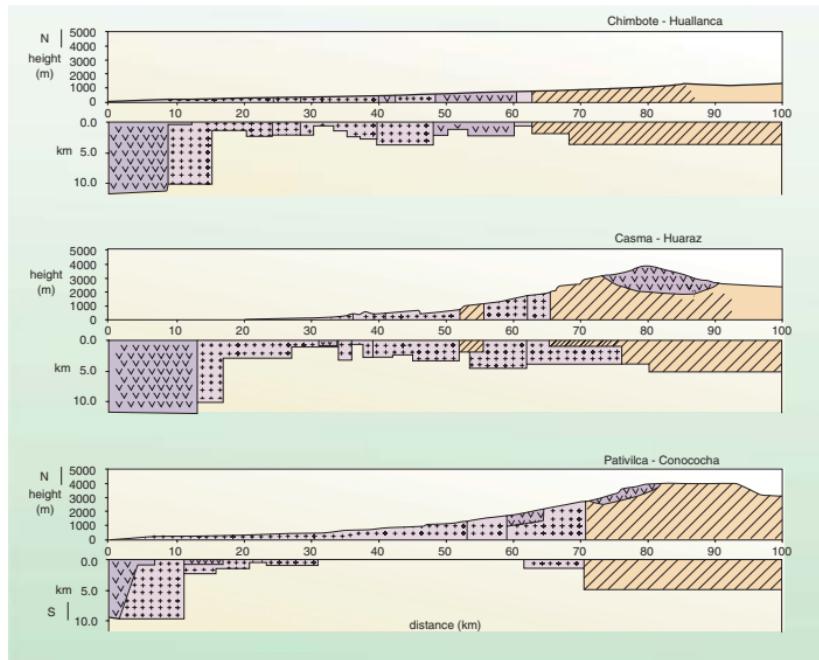


Figure 7.5 Geophysical cross section across part of the Peruvian Coastal Batholith based on gravity surveys. The plutons comprising the batholith (e.g. Santa Rosa Tonalite, Figure 7.4) are revealed as thin (circa 1–4 km thick) sheets, making the overall geometry of the batholith broadly tabular in form. The thicker subsurface zone in contact with basaltic volcanic rocks (v) is interpreted as a vertical magma feeder zone.

Normal zonation may result from *fractional crystallisation* processes whereby, during crystallisation, early-formed high-temperature minerals (for example, Ca-rich feldspars, Mg-rich pyroxenes or amphiboles) accumulated in the solidifying walls of the intrusion, leaving residual liquids towards the centre of the intrusion enriched in silica and alkali elements. In this way, a primary granodiorite magma, for example may fractionate to form a range of rock-types, from diorite and tonalite intrusion margins to granite at the centre, and the variation in rock-types across such intrusions may be completely gradational. As highlighted in the introduction, the identification of the intrusive sequence depends on careful field mapping of the interiors of plutons, for if an intrusion of diorite were invaded by a younger granite magma, then the best way of reaching this interpretation in the field would be by finding internal intrusive contacts (Section 7.5).

GRANITIC COMPLEXES

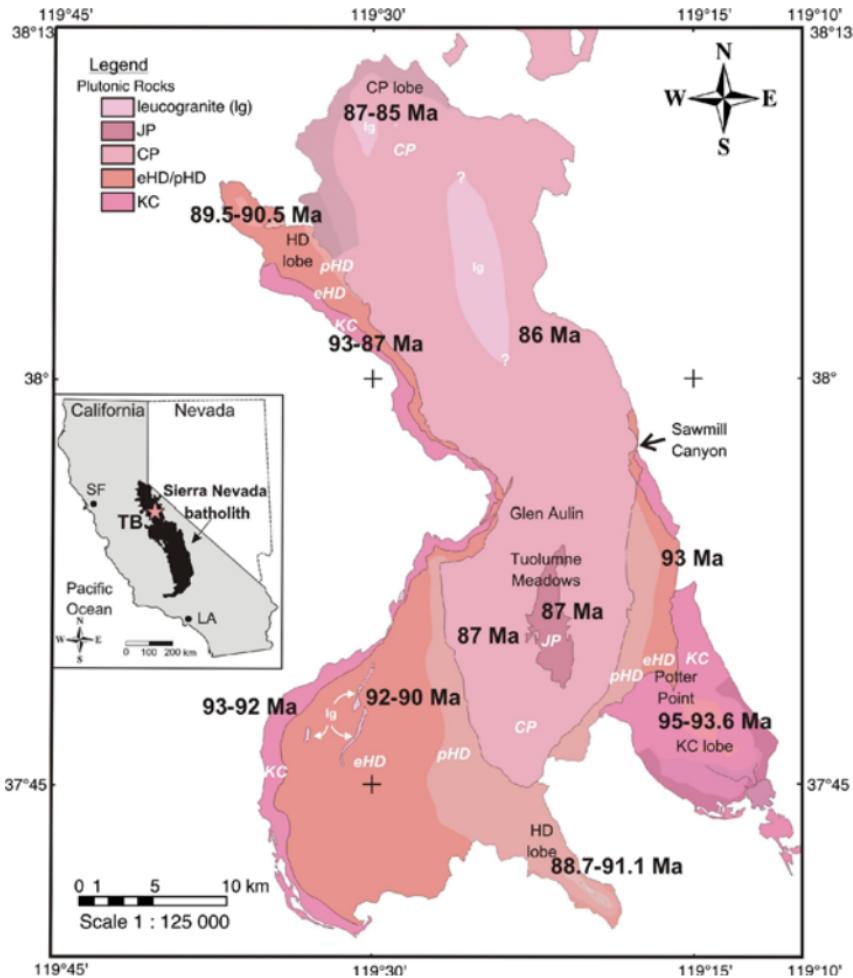


Figure 7.6 Map showing the normally zoned Tuolumne Batholith, Sierra Nevada USA. Facies types are: KC = Kuna Crest – hornblende-biotite granodiorite to tonalite, eHD = equigranular Half Dome = large hornblende + biotite granodiorite, pHD = porphyritic Half Dome = above plus K-feldspar megacrysts, CP = Cathedral Peak = biotite (only rarely hornblende) granodiorite to granite with large K-feldspar megacrysts, JP = Johnson Granite porphyry. Isotopic age dates (bold) that date emplacement of individual magma pulses give ranges from circa 95 to 85 Ma. (Courtesy Scott Paterson).

Where such contacts exist they can be complex if the earlier intrusion was incompletely crystallised, as in this state the intermingling of two magmatic fluids can result in magma mingling and where compositions are similar magma mixing or hybridisation across a broad (perhaps tens of metres) contact zone. Crystallised hybrid magmas may contain early-formed crystals derived from two compositionally-different magmas, each of which may be available for individual study on either side of the hybrid zone. If the earlier intrusion was solid but still warm enough to deform plastically, then diffuse contact zones, often with interpenetrating or lobate structures, would be found. Also, inclusions of earlier crystallised rocks within the later intrusion may provide clues about the state of the earlier magma when the later intrusion was emplaced. These alternative interpretations of compositional changes may be difficult to distinguish in the field and, in many cases, may only be resolved by laboratory work (see also Chapter 9).

7.4 Internal Structures and Textures

The mineral fabric (see also Chapter 2) and texture seen in granitic plutons can be more complex than that of extrusive associations. Often it is irregular in form, but regular linear or planar fabrics are often also well developed defined by alignment of phenocrysts, megacrysts or enclaves (Figure 7.7).

Identification of fabrics in plutonic rocks is a *major goal* of any field study as their measurement (both lineation and foliation) can provide useful information on mode, style and shear-sense during emplacement. For example, some alignment fabrics may reflect exclusively internal processes such as convective flow, layering or shearing during magma emplacement. This can result in localised but sometimes spectacular structures including ladder dykes, cross-layering and pipes (Figure 7.8).

7.4.1 Magmatic planar fabrics

Rotation and translation of crystals in cooling magma during emplacement either by local flow or regional deformation can lead to the development of a *magmatic planar* fabric. Identification in the field is an important but not always easy task. This is because magmatic fabrics often develop locally as non-pervasive foliations, either at the contact between different magma pulses, or in the vicinity of enclave swarms (Figure 7.9). They may also result from wider-scale deformation related to magma emplacement and in this case their orientations may correspond to the regional structural grain. Magmatic fabrics are unique in that by definition they form during crystallisation from a viscous fluid. However, their presence means that magma can also behave as an elastic solid, even though it may contain locally up to 50% melt. For this reason they are also referred to as *pre-full crystallisation fabrics*.

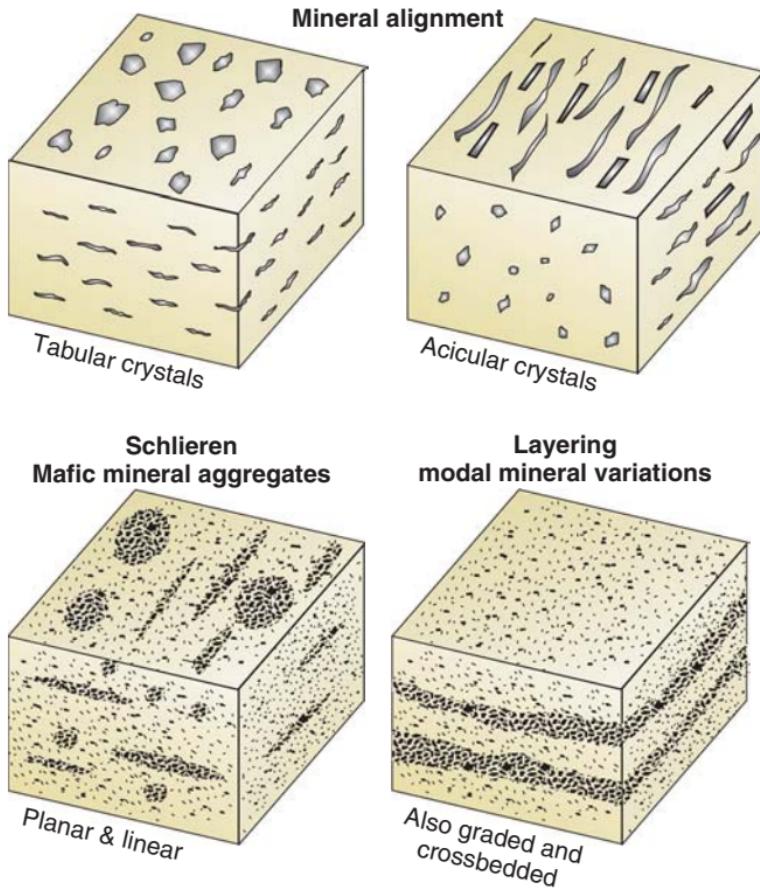


Figure 7.7 Schematic diagram highlighting some of the common crystal and crystal aggregate planar and linear fabrics.

7.4.2 Crystal plastic strain fabrics

Otherwise superficially quite similar planar fabrics can also be developed when plutonic rocks are subjected to regional deformation after the magma has crystallised (but not necessarily cooled to ambient temperatures). These are commonly S-C type fabrics located near the margins of plutons, or close to internal fault zones or contacts between different magma types. S-C fabrics comprise two contemporaneous planar foliations, an S-plane defined by the grain shape fabric, and a crenulation or shear plane (C). When observed correctly (that is normal to foliation in the X/Z plane of the strain ellipse and parallel to lineation), then the angle of intersection between the S and C

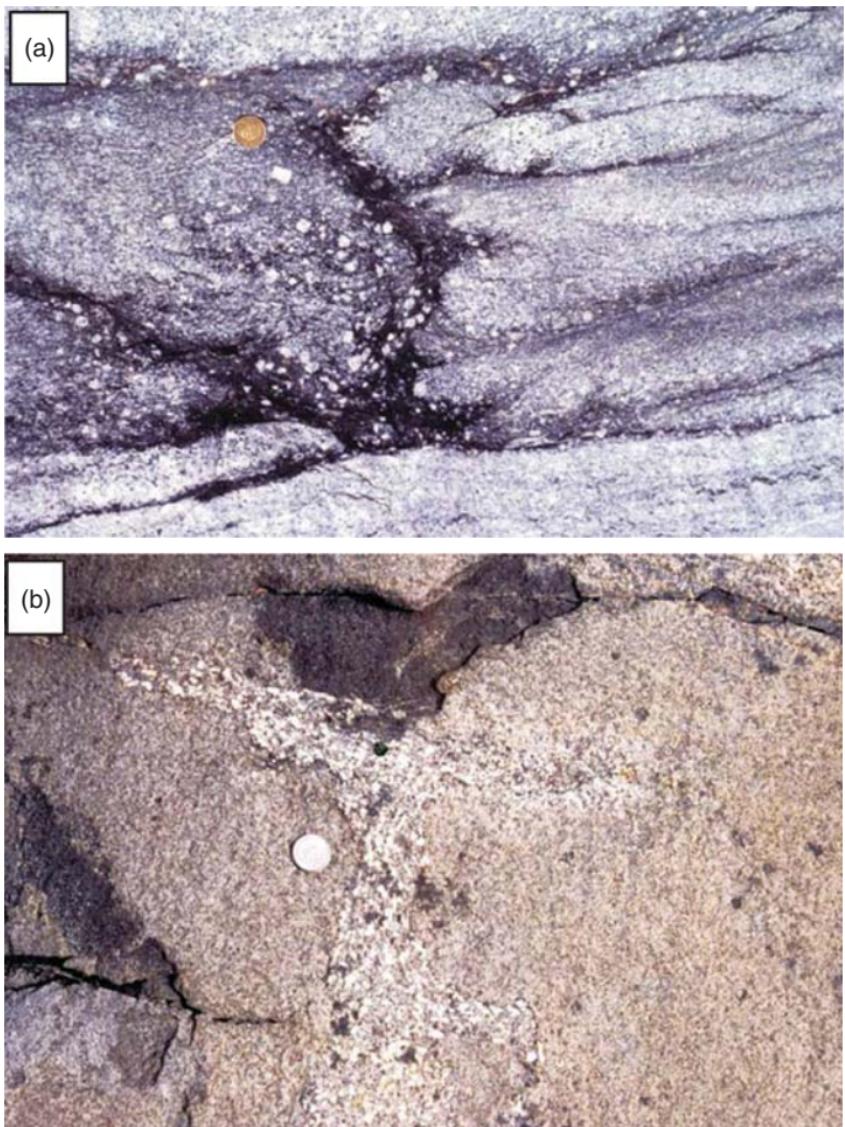


Figure 7.8 (a) Flow fabric defined by layers of biotite schlieren with local accumulation of alkali feldspar phenocrysts, Peru (b) mushy pipe of crystal-rich magma in granitic host, Maine, USA.

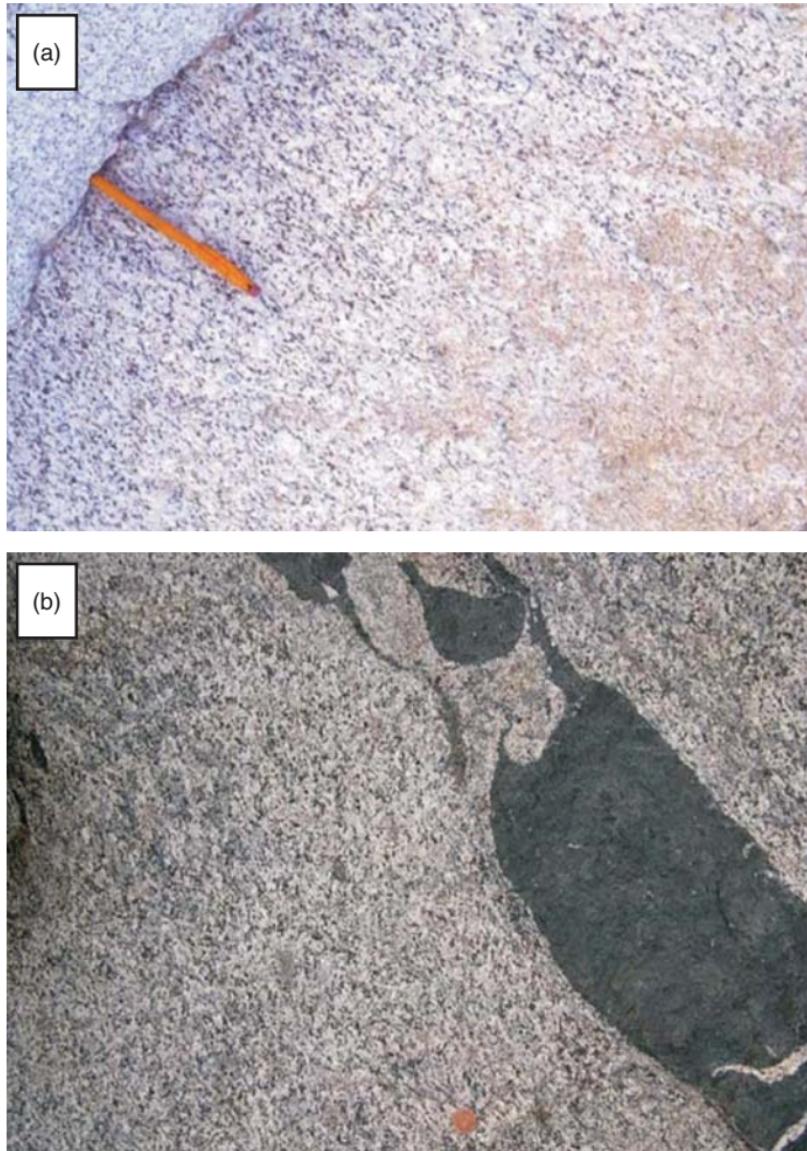


Figure 7.9 (a) Moderately dipping pre full crystallisation fabric in leucogranite defined by alignment of feldspar and biotite crystals (pencil marks fabric orientation). The fabric is subtle and shows no evidence for plastic deformation of the grain shape fabric (e.g. kinking of biotite, grain size reduction in the plane of the foliation). (b) Locally developed magmatic fabric in granite running parallel to the long axis of a sheared mafic enclave.

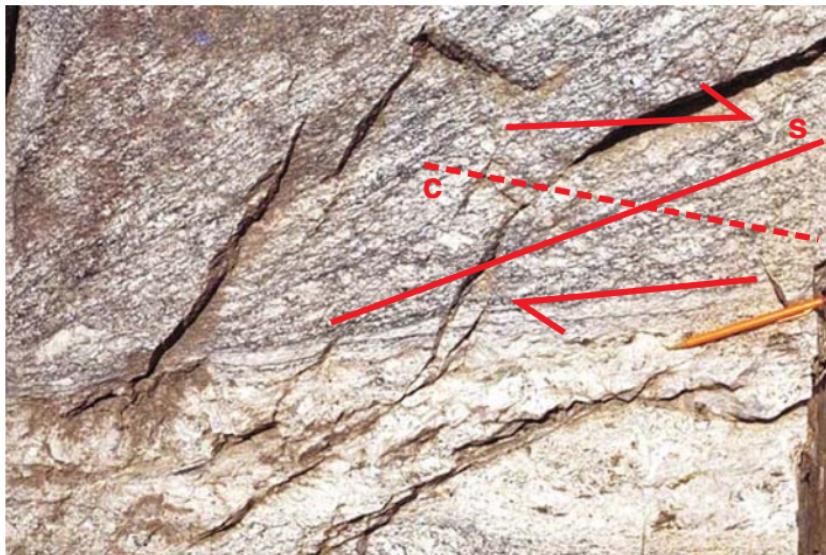


Figure 7.10 Deformed granite showing general orientation of a planar S-C fabric. Shear sense is clockwise (dextral). Shear strain is often amplified in deformed plutons at lithological boundaries, in this case a deformed pegmatite vein.

planes can reveal information about the degree of strain and also the sense of shear (dextral or sinistral) that caused the deformation (Figure 7.10). Detailed mapping of S-C fabrics as kinematics indicators and markers of strain intensity are very helpful in developing models of magma emplacement. At higher pressure-temperature conditions, metamorphic fabrics are more diffuse and the fabric may be difficult to interpret in terms of purely magmatic or purely metamorphic conditions. Under the highest pressure-temperature metamorphic conditions, plutonic rocks can partially melt and recrystallise as metamorphic rocks, such as gneisses.

It is important that a distinction between fabrics formed in the magmatic state (pre-full crystallisation fabrics) and those when the magma has solidified (crystal plastic strain fabrics) is made in the field. Diagnostic features of both are summarised in Figure 7.11. Measurement of both fabric types should be carried out just as you would when encountering fabric orientations in any deformed rock type, that is record carefully the dip and strike of the foliation and dip and plunge of any lineation or alignment of enclaves (see also structural guide McClay, 1991).

Finally, prolonged deformation in plutonic rocks can lead to development of a range of structures common in tectonically deformed rocks that overprint earlier magmatic fabrics. Thus, *shear zones*, *mylonites* and *cataclasites* are observed in

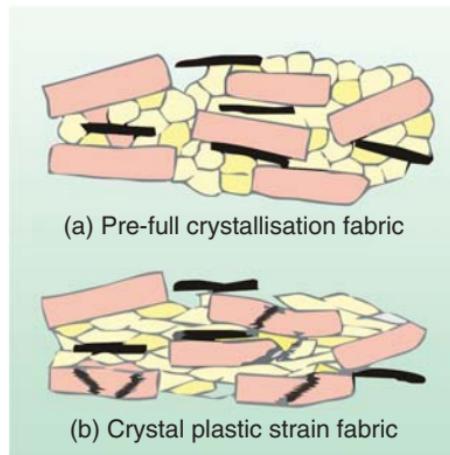


Figure 7.11 (a) Pre-full crystallisation fabrics are defined by passive rotation and parallel to sub parallel alignment of tabular and platy minerals. (b) Crystal plastic strain fabrics (formed very late stages of crystallisation or subsequently after complete solidification) show evidence of straining and change in particle shape (necking or rounding of feldspars, kinking and warping of mafic minerals).

granitic rocks and other plutons that have been exposed to progressive episodes of deformation that become more brittle-like with time (Figure 7.12). A good example of structural development in a granitic pluton is seen in the Miocene Cordillera Blanca batholith, Peru. Here the granitic magma was emplaced along a major crustal lineament and the deformation fabrics preserve a history of magmatic, post magmatic and overprinted tectonic deformation. The relationship between the internal fabric and structure development inside the pluton and its bounding faulted margin is shown in Figure 7.13.

7.5 Internal Contacts

It is now known that many large granitic bodies, once thought to be made of a giant lump of broadly homogeneous material are in fact comprised of sheets defined by changes in grain-size or composition. Where the temperature contrast between two cross-cutting magmas is large, their contact will often appear sharp or brittle-like in nature. Chilled margins may also develop. Where, however, interaction comes about in magmas with similar temperatures then their contact may be more diffuse. Here, contact margins may be gradational and mineralogical variations between sheets can be subtle. For example, a clear distinction between layers of tonalite alternating with granodiorite may be difficult

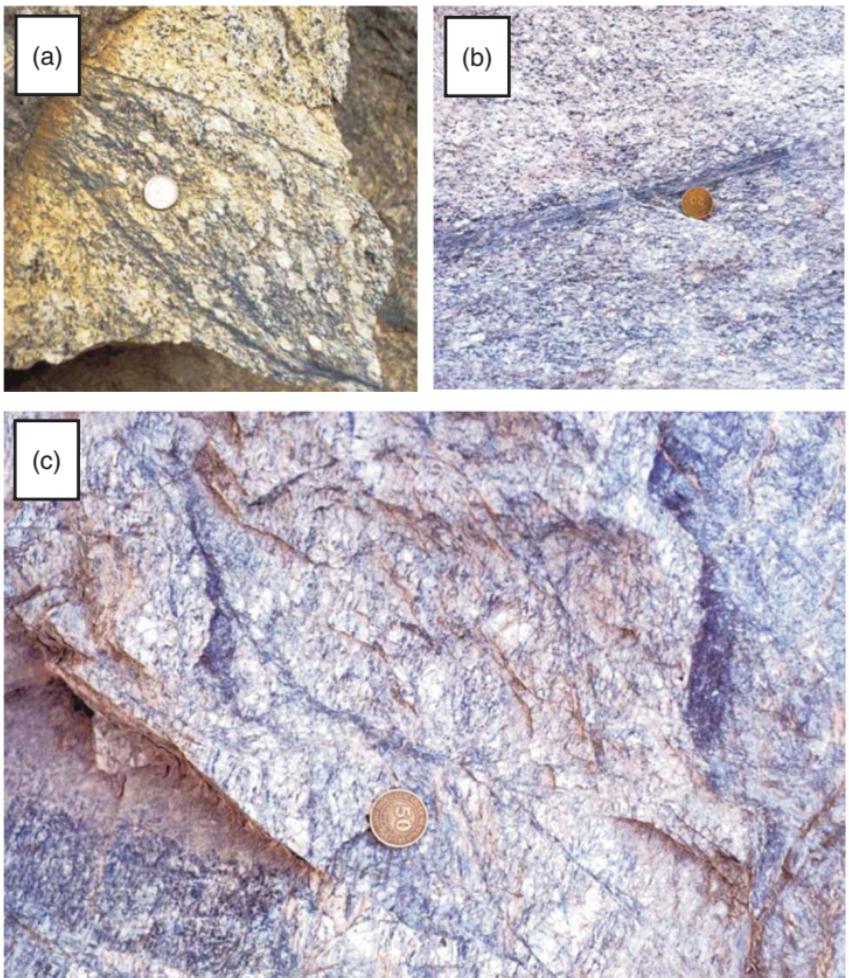


Figure 7.12 (a) Shear zone in granite (b) narrow band of dark mylonite (dark colour due to fine grain size) (c) irregular bands of cataclasite (brittle deformation) cutting earlier S-C fabric, Cordillera Blanca, Peru.

in the field, especially if the sheets are wide and their margins are diffuse. The main Donegal Granite in Ireland is a good example of this, where close inspection has shown the pluton was constructed via accumulation of a large number of smaller sheets (Figure 7.14). The nature of igneous contacts, either internally or with country rocks, needs to be recorded accurately, especially where such relationships are cross-cutting and hence age related (Figure 7.15).

GRANITIC COMPLEXES

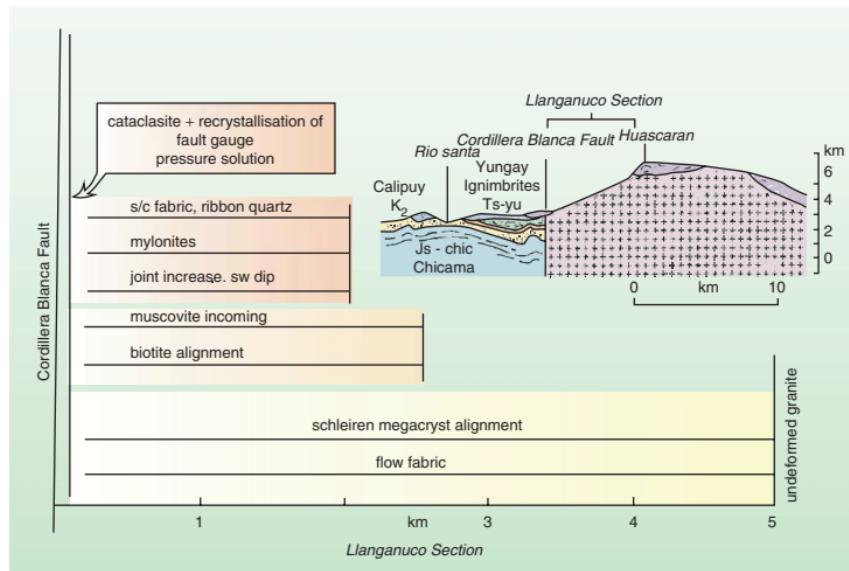


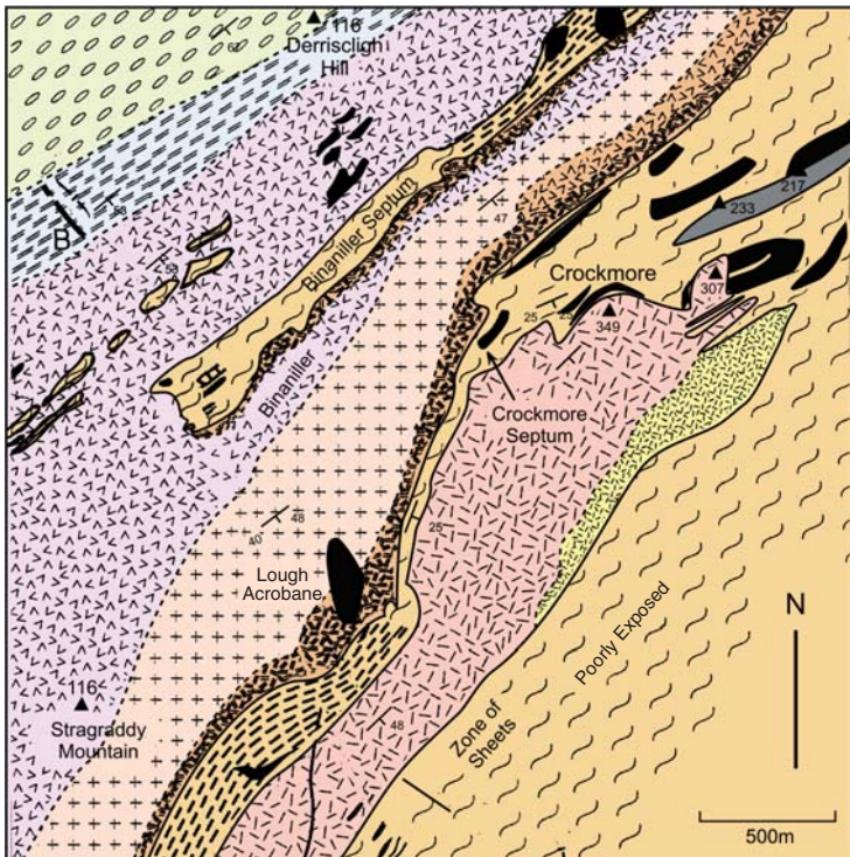
Figure 7.13 Summary of fabric and structural development in the Miocene Cordillera Blanca Batholith, Peru. The core of the batholith is relatively undeformed, but traversing westwards towards a major fault zone reveals a sequence of magmatic foliations (layering and schlieren) overprinted progressively by shear zones, mylonites and cataclasite.

7.5.1 Layering

Igneous layering similar to that associated with mafic–ultramafic magmas (e.g. Chapter 8), but involving different minerals, is also observed in calc-alkaline plutons (Figure 7.16). Small-scale lamination (2–3 cm band width) may be observed, in which the mineralogical distinction between more-mafic and more-felsic bands is clear cut. Disruption of early-formed mafic layers while the felsic magma remains mobile, perhaps as a crystal mush, produces rounded mafic *autoliths* or elongate *schlieren*. Schlieren are anomalous concentrations of mafic minerals (biotite, for example) and are commonly interpreted as forming by one of four mechanisms: (i) crystal sorting during magma flow, (ii) shearing of magmatic enclaves (see Chapter 9), (iii) crystal settling (iv) late-stage hydrothermal alteration.

The development of crystal mushes during incipient crystallisation of magmas may be associated with the extraction of residual liquid from the consolidating mush either by gravitational effects related to temperature or composition (buoyancy or compositional convection) or by deformation and shear-enhanced ditatancy of the granular mush framework. The efficiency of these processes is

GRANITIC COMPLEXES



	Leucocratic biotite - muscovite monzogranite		Heterogeneous granite with pelitic xenoliths
	Muscovite - garnet aplogranite		pelite and seimpelite
	Leucocratic biotite - granite		marble
	Coarse to medium - grained		matadolerite
	Fine to medium grained		basalt - dyke
	Heterogeneous granite with quartz xenoliths		granite sheet
	Heterogeneous granite with Thorh xenoliths		

Figure 7.14 Sheeted pluton: detailed section through Barnes Gap Traverse, main Donegal Pluton, Ireland showing the internal geometry, fine structure and lithology of granite sheets that comprise the intrusion.

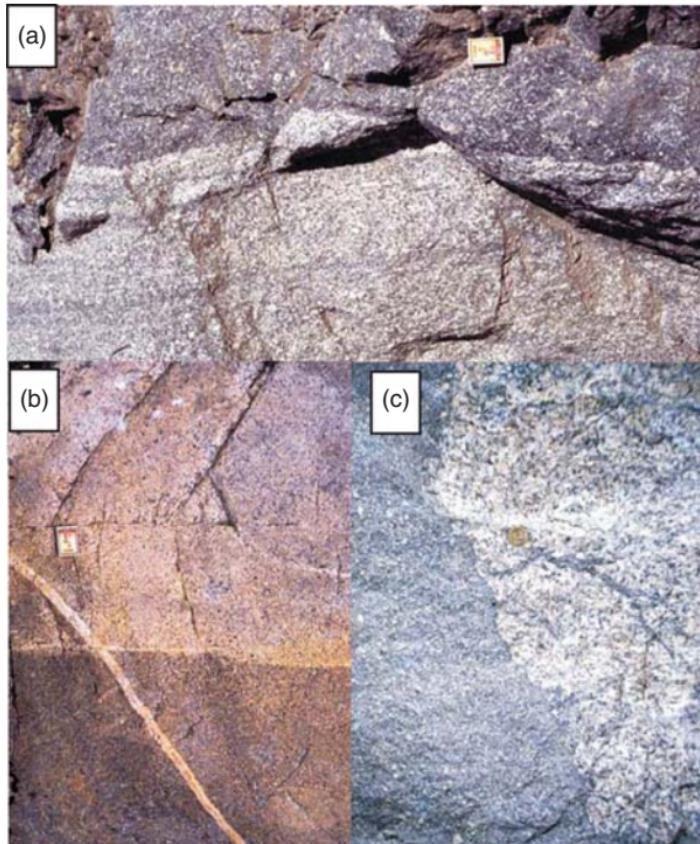


Figure 7.15 Details of internal contacts in granitoid plutons (see Figure 7.3 for description of contact types) (a) subhorizontal contact between tonalite and diotite, contact sharp and sinuous (b) Horizontal contact between granite and diorite. Contact marked by narrow aplite band and cut subsequently by diagonal aplite vein. Contact sharp, straight. (c) vertical contact between granodiorite and diorite. Note apophysis of dark diorite cutting granodiorite gives timing relationship, contact sharp and sinuous. (d) (on next page) Granite-microgranite contact, sharp but locally diffuse. Note local E-W alignment of feldspar phenocrysts (magmatic fabric).



Figure 7.15 (continued)

in nature there is a race between the physical/mechanical processes at work and the rate at which the system is cooling down – it is the relationship between the timescales of these competing processes that allows (or not) for the richness and diversity of structures preserved in igneous rocks. Unfortunately, there is no common norm, meaning that each pluton must be assessed on its own merits.

7.5.2 Jointing

All plutonic rocks contain fairly regular sets of joints (Figure 7.17). *Primary joints* are related to cooling of the intrusion. One kind of primary joint is referred to as cross jointing which are mostly vertical and confined generally to pluton margins where the rate of cooling is highest. Evidence that cross joints form late in the crystallisation of plutonic magmas comes from the observation that aplites and other vein structures are seen to follow them, implying that both are coincidental and related to internal stress patterns in the cooling (and contracting) magma. Cross joints are often affected by late stage hydrothermal fluid interaction associated with emplacement, and may be therefore be mineralised (see Chapter 10).

Secondary joints form during subsequent phases of crustal-scale tectonic deformation or uplift and unroofing. Sheet jointing is a good example: a pervasive, generally contact parallel/horizontal joint mass, which classically preferentially erodes giving rise to *tor* topography (Figure 7.18). At low pressure-temperature conditions, tectonic processes might be responsible for a wide or close-spaced fracture cleavage in plutonic rock. Jointing imparts a distinctive fabric in igneous rocks that varies with intensity towards and away from larger scale faults or shear zones. Together, these structures are referred to as discontinuities and have an important role in governing the porosity

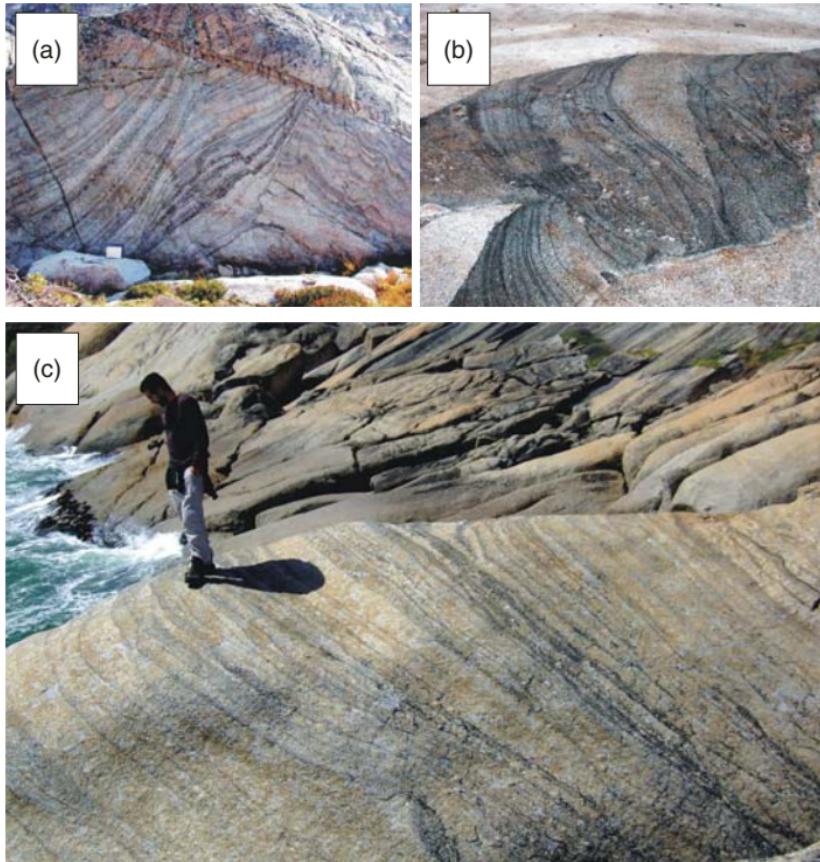


Figure 7.16 Examples of magmatic layering in granites. (a) and (b) show cross-cutting Schlieren layers with local megacryst accumulations, Sierra Nevada, USA. (c) Regular and truncated biotite layers, Cape Granites, Cape Town, South Africa.

and permeability of fractures in igneous rocks. They also act to weaken the overall rock structure or rock mass. Ways to classify their effects in the field are discussed in Chapter 10.

7.5.3 Aplites and pegmatites

Aplites and pegmatites are veins comprised mainly of quartz and alkali feldspar with subordinate muscovite and occasional garnet that form during the final stages of cooling and crystallisation (Figure 7.19). They are distinguished

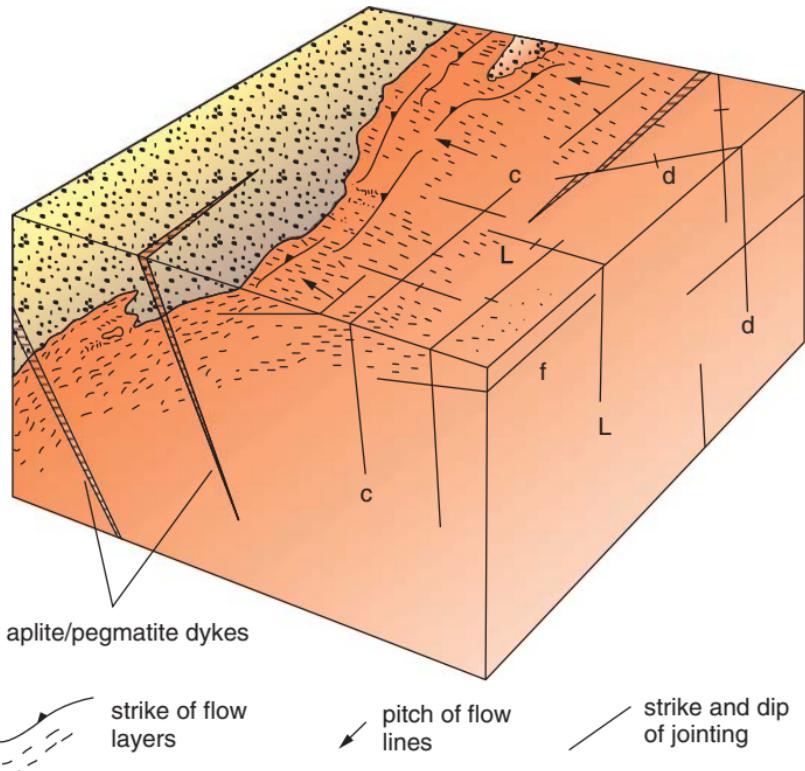


Figure 7.17 Block diagram showing the relationship between flow structures and joint systems in an intrusive mass. *c*, Cross joints; *d*, diagonal joints; *f*, flat-lying joints; *l*, longitudinal joints.

apart by their grain size – aplites are fine grained (sometimes with a sugary or saccahroidal granular texture) while pegmatites are much coarser grained rocks with individual crystals growing to several centimetres or even metres in size in extreme cases (see Section 2.5.2). Pegmatites can also be mineralised due to late stage enrichment in trace metals (Chapter 10). They often form contemporaneously with primary cooling joints through a process of auto intrusion (hydrofracturing) of cooling magma. Both can occur in isolation but are more commonly found as cognate sets with sharp margins. They provide excellent time markers and their dips and strikes useful structural information on the orientation of late stage thermal stresses set up during cooling. They can often be seen cutting surrounding country rock where they can be traced back to their host pluton.

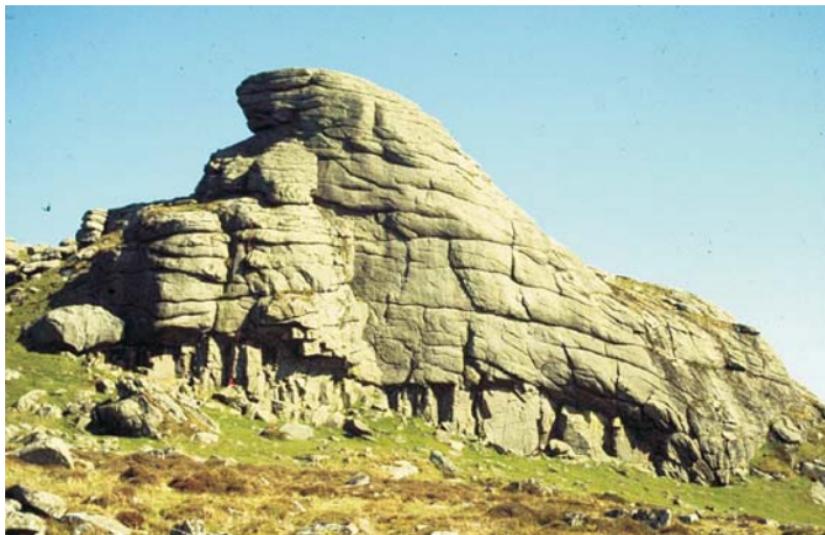


Figure 7.18 Distinctive granite Tor, Devon, UK. Two joint sets are visible, flat-lying sheet joints generally parallel to the surface and vertical cross joints.

7.6 Emplacement Timing

Most of the characteristics of calc-alkaline intrusions may be recognised in granitoid suites of all geological ages but, the older the intrusion, the greater is the probability that more complicated fabrics will have developed due to the additional effects of tectonic overprinting. In such cases, any tectonism (or associated metamorphic effects), such as regional jointing or post magmatic foliations, should be traceable laterally into the country rocks. Granitoids which contain internal structures that have survived a later stage of post-emplacement deformation during which new country rock fabrics, developed around the intrusion, are described as *pre-tectonic*. *Syn-tectonic* intrusions, including pre-tectonic intrusions from which all traces of an earlier fabric are lost and intrusions emplaced during metamorphism. These tend to have internal flow structures parallel to the fabrics in adjacent country rock schists or gneisses. Post-tectonic intrusions are generally unfoliated cutting pre-existing fabrics (Figure 7.20).

7.6.1 Case studies

Intrusions with different relationships to regional metamorphism may sometimes occur at the same level of exposure, for example Figure 7.21 illustrates a typical deeply-eroded area of southwest Finland where a syn-tectonic calc-alkaline suite (dated at 1900 Ma), with east-west flow structures parallel to the foliation in local amphibolitic gneisses, is cut by a later post-tectonic (1600 Ma) alkali granite. The

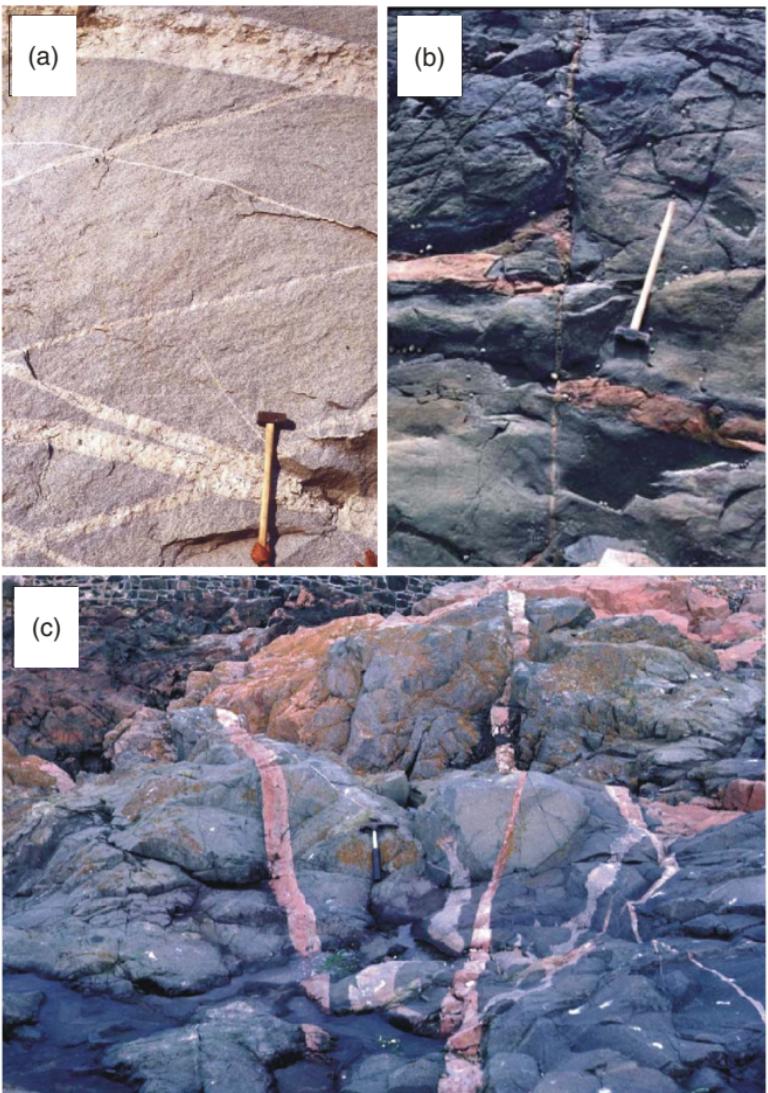


Figure 7.19 Aplite and pegmatite veins: (a) Cross cutting narrow aplites and coarser grained pegmatite veins, Peru. (b) Aplite dyke offset by dextral wrench fault with plane infilled with aplite, Channel Islands, UK. (c) Tentacle-like fine grained aplites, Channel Islands, UK.

GRANITIC COMPLEXES

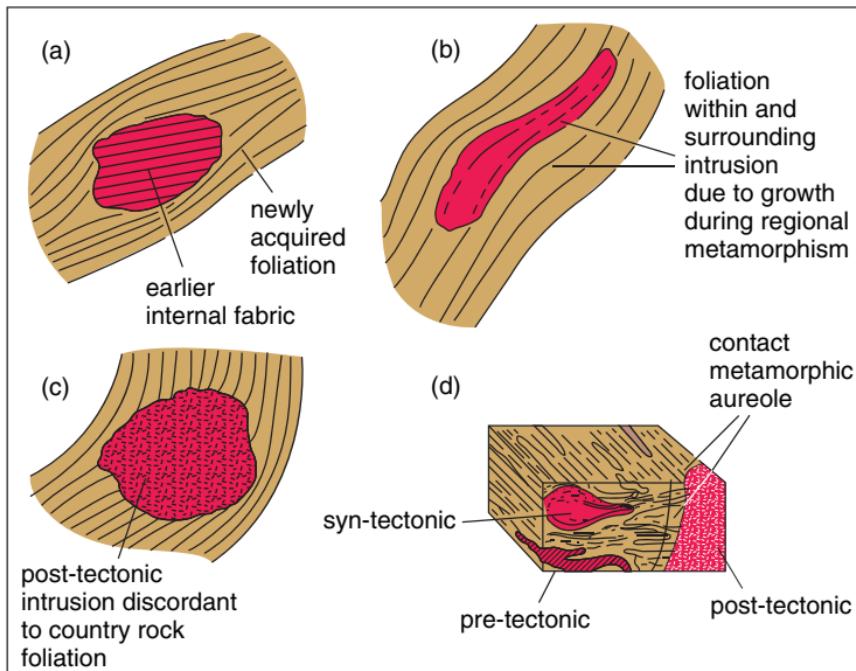


Figure 7.20 The relationship between intrusive rocks and phases of deformation: (a) A pre-tectonic intrusion with an earlier internal fabric, around which metamorphic fabrics have been developed without affecting the intrusion. (b) Syn-tectonic intrusion emplaced during regional metamorphism, and controlled in terms of both shape and internal fabric by the tectonic stress regime. (c) Post-tectonic intrusion emplaced through the discordant to pre-existing fabrics. (d) Block diagram showing syn-tectonic and pre-tectonic intrusions (light shading) developed due to strong vertical and moderate left-to-right compression in three dimensions. Fine lines indicate mineral lineations in the syn-tectonic intrusion and host rocks. Solid ornament (right) denotes a post-tectonic intrusion.

calc-alkaline rocks are dominated by *trondhjemite* with lesser amounts of tonalite and gabbro. This raises a general point about long-term changes in the petrology of calc-alkaline intrusions; whereas modern batholiths are composed of a granite-granodiorite-diorite or tonalite-gabbro suite, in older examples from the early Proterozoic and Archaean (that is, >1500 Ma before present) a trondhjemite-tonalite-gabbro suite is more common. So, in the field, ancient and modern calc-alkaline intrusions may contrast in both their degree of metamorphism and in their petrology. However, the contrast is also a function of tectonic setting and erosion level, because the plutons of young island arcs and much of the

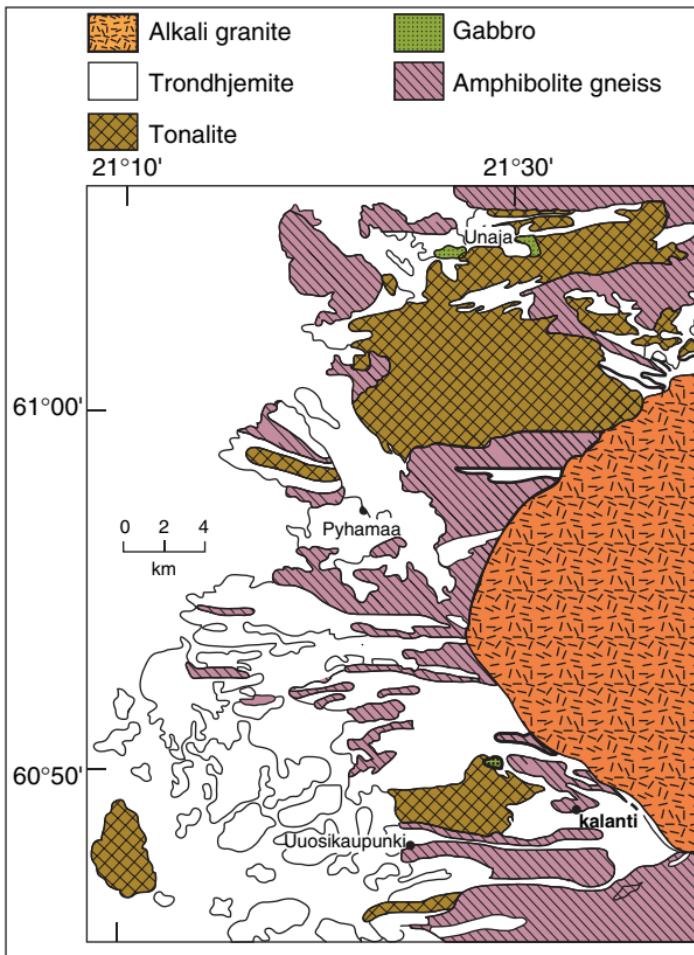


Figure 7.21 Simplified geological map of southwest Finland showing syn-tectonic calc-alkaline tonalite and amphibolitic gneisses in the west intruded by a post-tectonic alkali granite in the east.

Andean batholith both contain far more tonalite and gneissic granodioritic rocks than granite (s.l.).

When combined with other field measurements, notably Anisotropy of Magmatic Susceptibility (AMS) surveys, a comprehensive picture of pluton fabric development can be assembled that allows geologists to unravel the often complex emplacement histories of granite and other magma bodies. A well studied example is the Ardara pluton, Donegal, Ireland. Ardara is a zoned, syn-tectonic

pluton comprising an outer quartz monzodiorite surrounding a central core of granodiorite. In plan view its geometry is broadly circular but the outer facies shows evidence for pure flattening (plane strain) that increases in intensity towards the margin. The interpretation is that the pluton was emplaced as several magma pulses *in situ*, causing the intrusion to swell outwards on all directions, pushing and deforming magma against the surrounding country rocks. This type of emplacement mode is referred to as *ballooning*. S-C fabrics and flattened magmatic enclaves (Chapter 9) record the kinematic and strain history of pluton emplacement (Figure 7.22). The country rock (Dalradian metasediments of the Appin Supergroup) are contact metamorphosed, with an inner zone of sillimanite, garnet and cordierite, and an outer (lower temperature) zone containing staurolite, kyanite and andalusite.

7.7 Distinctive Granitoid Textures

7.7.1 Porphyries

One special kind of granitic rock whose location within plutonic terrains often points to mineralisation are termed porphyries. These rocks are recognised by their highly porphyritic nature, with large, mainly euhedral phenocrysts embedded in a fine (quenched) groundmass. Their overall composition ranges from tonalitic to leucogranitic and analysis of the groundmass can be extremely useful in determining the primary magma composition, something extremely difficult to establish in the medium to coarse grained rocks that comprise most granitic igneous terrains. These rocks are important due to their associated mineralisation. For example, at El Tienetie, Chile, one of the world's largest copper mines, the metal ore is hosted in granite porphyry with an 'adakite-like' chemical signature.

Here are brief descriptions of several distinctive, though rather uncommon, textural varieties that may be encountered in the field, among granitoids of various types.

Graphic granite refers to an inter-growth texture which develops in some quartz-alkali feldspar pegmatites. Aligned, rod-like, or triangular light grey blebs of quartz are enclosed within large feldspar crystals and take on the appearance of runic characters. If the quartz-alkali feldspar intergrowths are confined to smaller crystals, perhaps in the groundmass of a porphyritic granitoid, then the texture is *granophyric* (and is often only visible in thin section). Rocks with granophyric texture are usually fine-grained porphyritic granodiorites and granites ('high-level' granites) and they occur most frequently as minor intrusions.

Orbicular granite contains large ovoid bodies (orbicules), usually up to several centimetres in diameter, in which the components are arranged in concentric layers of leucocratic quartz – feldspar-rich material alternating with mafic layers consisting primarily of dark hornblende and/or biotite. Each orbicule contains a core, commonly an igneous xenolith fragment but sometimes composed of the

GRANITIC COMPLEXES

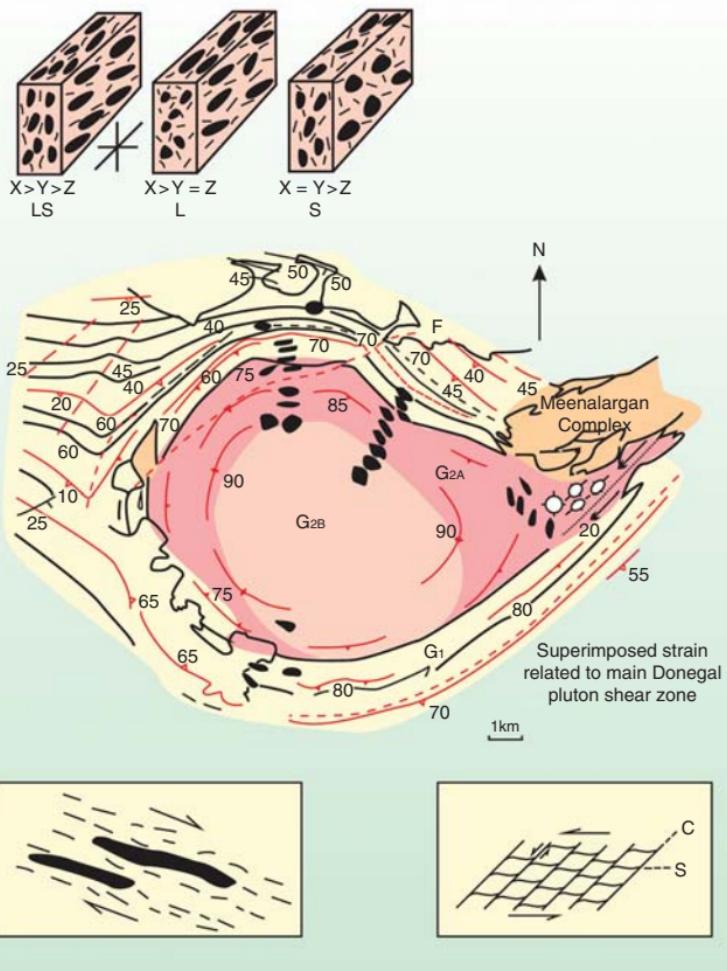


Figure 7.22 Synthesis of field data from in and around the Ardara pluton, Donegal. Fabrics and structures within the pluton and country rock including magmatic foliation planes, (dip degrees numbered), lineations, kinematic markers (insert shows S-C fabrics and asymmetric, sheared enclaves) and progressive flattening of magmatic enclaves (plane strain) towards the outer margin. Together with petrographic variations (Ardara pluton is compositionally zoned), these structures suggest a pulsed emplacement history dominated by in-situ ballooning of magma.

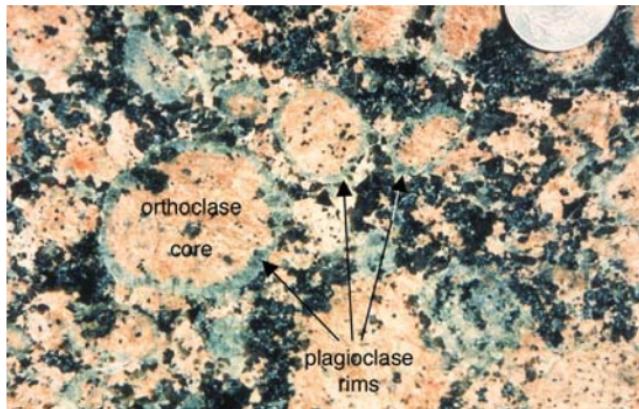


Figure 7.23 Rapakivi granite texture with corroded megacrysts of K-feldspar (pink) mantled by greyish plagioclase. (Courtesy of Robin Gill).

same granitic material in which the orbs are embedded. The thicknesses and mineral contents of the layers forming different shells of the orbicules are often similar, suggesting that they grew together. This may have happened during rhythmic crystallisation, analogous to that responsible for igneous lamination (Sections 2.4.2 and 8.2), with the difference that, instead of resulting from crystal settling, the orbicules accreted available crystals from the magma before final consolidation of leucocratic material in the interstices took place.

Rapakivi granite (Figure 7.23) contains large phenocrysts (several centimetres long) of alkali feldspar, often salmon pink or flesh-coloured, mantled by a rim (1–2 mm thick) of white plagioclase feldspar. These complex phenocrysts are usually embedded in a groundmass of quartz, feldspars, biotite and/or amphibole. Often the alkali feldspars are rounded, unlike most phenocrysts in igneous rocks which are euhedral, possibly indicating corrosion or remelting of their edges into the magma before crystallisation of the plagioclase rim. Rapakivi texture is most common in alkali granites (Chapter 9), notably the type localities in south Finland, but is also found in some calc-alkaline types and as decoration on many building walls.

7.8 Metamorphic Aureoles

This brief section on metamorphic aureoles is included here because aureoles are best developed around granitoid (granodiorite and granite) intrusions emplaced within a few kilometres of the Earth's surface, where there is liable to be a marked contrast between the temperature of the intrusion (circa 700–1000°C) and the country rocks (<200°C). Aureoles reflecting metamorphism in the temperature range 200–700°C are characteristically up to 2–3 km wide around

GRANITIC COMPLEXES

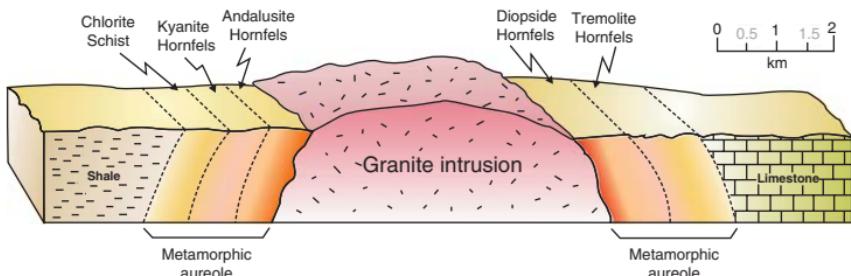


Figure 7.24 Schematic example of a metamorphic aureole surrounding a granite body with a shale protolith on one side and a limestone protolith the other.

granitoid intrusions 10–15 km in diameter, where the margins dip at 60–70° below the surface. Generally, the aureoles of basic intrusions such as dykes and sills are much smaller; for example many large sills with thicknesses of up to several hundred metres may show metamorphic effects up to only a few metres from the contact. The textures and mineralogical changes developed by contact metamorphism of common rock-types are summarised in Table 7.2 and highlighted in Figure 7.24; these characteristic minerals provide a basis for mapping and studying aureole zones. For further details of high temperature-low pressure metamorphism of sedimentary sequences, refer to Fry (this book series).

7.9 Summary of the Field Characteristics of Granitic Complexes

Granitic rocks are the final expression of extreme differentiation of the solid earth. These low density, silica rich materials comprise extensive tracts of earth's middle to upper crust and so far appear unique in the solar system. Problems surrounding the origin and emplacement of granitic rocks include the rate and mode of transport from the source region to site of emplacement, and how space is made for the invading magmas. Recent field and theoretical studies suggest that plutons are not the giant, amorphous masses once thought but are instead relatively thin sheets comprised of multiple pulses of magma. In this way, plutons and batholiths are built up incrementally, perhaps in relatively short bursts of emplacement activity. If so, then evidence for this should be preserved in the field in the form of sheeted or pulsed intrusions and evidence for internal contacts should be pursued judiciously.

Calc-alkaline plutons comprise different proportions of coarse grained rocks: granite, granodiorite and tonalite, with subordinate diorite and gabbro. Plutons may show evidence of emplacement by stoping; that is the fracturing of the roof and incorporation of country rock xenoliths. Together, a large number



Figure 7.25 Elongate crystals of andalusite (chiastolite) in contact metamorphosed (hornfelsed) slate.

of individual plutons (each 1–50 km in cross-section at outcrop, and roughly tabular), may comprise a batholith. When examining plutons, it is necessary to:

1. Describe and determine the relative proportions of rock types present; use Table 7.1 if a preliminary tectonic interpretation is required.
2. Examine contact relations to determine age relations.
3. Study the nature and distribution of enclaves (see Chapter 9).

Examination of individual plutons may reveal small-scale layering or lamination (0.01–10 m), or zoning on a pluton scale. These may be due to crystal accumulation under gravity or shearing during flow. Plutons may also contain igneous rocks derived by multiple injections of different magmas between which contacts may be recognised, with associated magmatic enclaves.

Important activities in the field are to:

1. Describe the nature and orientation of any planar igneous foliation and identify timing of formation as either magmatic (pre-full crystallisation) or post magmatic (crystal plastic strain fabric).

2. Measure any S-C fabrics and other kinematic indicators and record shear sense.
3. Look out for layering and ‘exotic’ magmatic structures such as pipes, ladder dykes and accumulations of megacrysts (log jams) and record their mineralogy, orientation and spatial association.
4. Look carefully for internal contacts. Record dip and strike of contacts, note changes in lithology. Is the pluton sheeted internally?
5. Study the nature and distribution of enclaves.
6. Decide whether a pluton is zoned on a regional scale.

It is important to remember that as with all rocks, the fabric you record is the last increment of finite strain and that a crystal plastic strain fabric may be superimposed on an originally magmatic one. This distinction may be critical when mapping in detail internal contacts inside plutons.

In addition:

1. Record any distinctive granitoid textures – graphic granite, orbicular granite and rapakivi granite, and jointing structures and mineralisation features (see Chapter 10).
2. The type of contact metamorphism in the country rocks surrounding high-level granites may be distinguished with the aid of Table 7.2.

GRANITIC COMPLEXES

Table 7.2 Mineralogical and textural effects of contact metamorphism of common rock-types (for details of minerals and textures not defined here, refer to Fry, this book series).

Pure limestone	No mineralogical changes; increase in degree of recrystallisation as contact is approached, leading to formation of completely recrystallised <i>marble</i> . The thickness of the zone of contact metamorphism and recrystallisation varies from a few centimetres adjacent to minor intrusions, to hundreds of metres adjacent to larger plutonic intrusions
Impure limestone (inc. dolomite)	Complex sequence of mineralogical changes; the following mineralogical assemblages may develop as the contact is approached: (i) calcite ± dolomite ± clay minerals; (ii) calcite ± dolomite ± olivine (Mg-rich olivine, forsterite) ± amphibole (tremolite); (iii) calcite ± serpentine ± chlorite ± tremolite; (iv) wollastonite (Cappyroxene) ± garnet (Ca-Mg-Fe rich garnet, grossularite and andradite) ± pyroxene (Ca-Mg rich pyroxene diopside). These changes are accompanied by recrystallisation to marble. The thickness of the zones corresponds to those described for pure limestone above. Typically, for a metamorphic aureole several hundred meters wide, the zones with diopside and wollastonite would be only a few metres thick
Shale	Complex sequence of mineralogical changes; the following mineralogical changes may develop as the contact is approached: (i) development of ‘spots’; (ii) large crystals (or porphyroblasts) of andalusite (chiastolite) (Figure 7.25); and (iii) cordierite-andalusite-biotite. These changes are accompanied by progressive recrystallisation to hornfels-textured rocks (these are usually banded and may show small-scale deformation as a result of partial melting). These zones may all be developed within 1–2 km of the igneous contact (cf. Figure 7.24)
Sandstones	No mineralogical changes; recrystallisation limited to within a few metres of the contact, and does not usually exceed a few tens of metres even around large intrusions
Basic volcanic rocks	Complex sequence of mineralogical changes; the following mineralogical assemblages may develop as the contact is approached: (i) epidote-chlorite-amphibole; (ii) plagioclase-hornblende; and (iii) plagioclase-pyroxene. These changes are accompanied by progressive recrystallisation to form basic hornfels-textured rocks (as above) within 1–2 km of the igneous contact

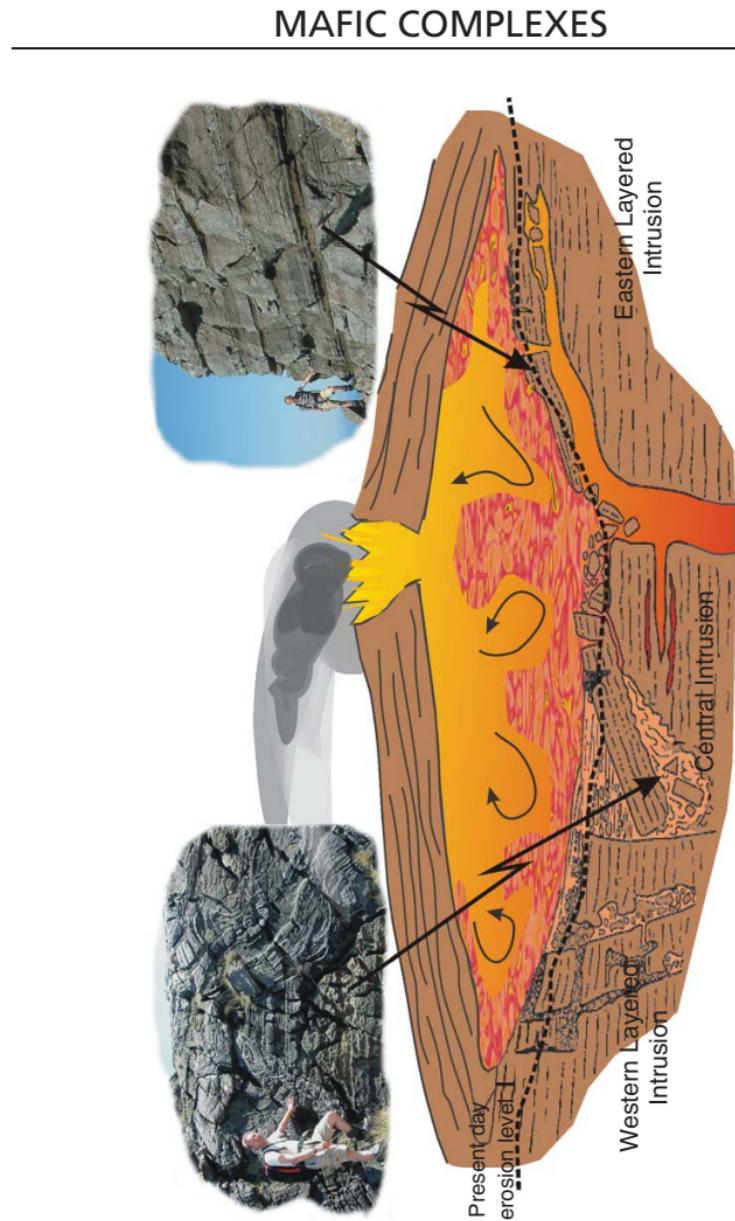


Figure 8.1 Schematic cross section through the Rum volcano, showing current erosion level and pictures of simple and complex igneous layering structures.

8

MAFIC COMPLEXES

Major occurrences of rocks that are solely or predominantly mafic/ultramafic compositions form a major sub-group of igneous rocks that are important in many geological settings, and can provide key information about igneous processes, crustal development, the Earth's temperature and are often of economic value. Here we take a look at these mafic/ultramafic complexes and how they can be tackled in the field.

8.1 General Features and Occurrence

Although the mafic rock-types, gabbro (35–65% plagioclase feldspar, 35–65% pyroxene and olivine) and melagabbro (65–85% mafic minerals) are well known as minor constituents of calc-alkaline and alkaline granitoid suites (Chapter 7), they are the major constituents of mafic intrusions and are important constituents of ultramafic intrusions (where most of the rocks contain over 90% mafic minerals, cf. Figure 3.14). These rocks occur in two main settings. First, they occur as isolated intrusions ranging from a few to hundreds of kilometres across with circular/sub-circular plan and elongate/elliptical intrusion cross-sections. Such intrusions have been emplaced in within-plate ocean island and continental margin settings (for example, the islands of Rum and Skye in north-west Scotland, for example, Figure 8.1). Second, mafic-ultramafic rocks occur as linear groups of intrusions, representing former oceanic lithosphere, which were emplaced tectonically by *obduction* (over-thrusting) at continent-continent or continent-island arc suture zones. The latter are known as *ophiolite complexes*; they include characteristic associations of lavas, dykes and mafic-ultramafic intrusions. The occurrence of mafic-ultramafic complexes associated with greenstone belts will be introduced in Section 8.4.

A wide variety of different rock-types may be present in a mafic-ultramafic intrusion. The mafic rocks vary from clinopyroxene-bearing gabbros through to rocks in which orthopyroxene is more abundant, and for which the name *norite* is often used instead of gabbro. Thus, norite is a gabbro-like rock with 35–60% of plagioclase feldspar and 35–65% of orthopyroxene, which is often bronze-coloured. However, it is the *ultramafic* rocks that show the greatest mineralogical diversity. When the mafic mineral constituents exceed 80% of



Figure 8.2 Fibres of serpentine/asbestos in serpentinite (altered peridotite) from Cyprus.

the rock, then one or more of olivine, clinopyroxene and orthopyroxene may become major constituents, and this leads to the discrimination of peridotites (>40% olivine), pyroxenites (>60% pyroxene) and the other less common varieties shown in Figure 3.14.

In general, *peridotite* is usually a mid-green to black, medium or coarse-grained rock-type when fresh (buff brown when weathered), though the olivine in many examples has become altered to the dull, light-green colour of serpentinite (a soft, fibrous, hydrated magnesium silicate, Figure 8.2). Serpentinitisation is also common in *dunite*, a medium- to coarse-grained rock almost entirely made (>90%) of olivine which, when fresh, may have a ‘sugary’ texture. *Pyroxenites* have less tendency to alteration and are usually black or greenish-black in colour, though abundant orthopyroxene may have a bronze colouration (*Bronzite*). Distinctive accessory minerals, particularly in continental ultramafic bodies and kimberlite xenoliths (Chapter 6), include occasional shiny grains of *phlogopite mica* and less commonly Fe-Ti oxide mineral grains such as ilmenite, magnetite and chromite (Table 3.4) which form small black granular clusters. A rare, but economically important and easily identified rock type, *chromitite* (see Section 10.2), occurs as layers in ultramafic intrusions. This is a heavy medium grained melanocratic granular rock in which black, euhedral chromite grains are enclosed in a matrix of clinopyroxene and/or olivine.

8.2 Continental Mafic-Ultramafic Intrusions

These bodies are large, discordant, funnel- or saucer-shaped intrusions, generally with inward dipping contacts, causing them to be thicker in the centre than at the perimeter, that is they are *lopoliths* (cf. Chapter 6). Most of them contain medium- to coarse-grained ultramafic, gabbroic, leucogabbro and even anorthositic lithologies as shown in Figure 8.3; together these rock-types form a series of repetitive, or rhythmic layers (with igneous lamination) which crystallised within a magma chamber fed by mafic/ultramafic magma. The terminology for layered rocks is based on classic studies which formulated the concept that the layers represent concentrations of minerals accumulated by mineral settling; the layered rocks have been termed *cumulates*. However, it is now realised that some layered rocks cannot be a result of such crystal settling, and that layering might form by a number of processes. Hence the term ‘cumulate’ may be applied to such rocks in a non-genetic sense, in which crystal settling is a possible but non-essential process in the origin of cumulate rocks. Examples of a mafic-ultramafic layered intrusion include the Bushveld intrusion (see Section 10.2); Skaergaard intrusion, Greenland, The Dufek and Dais intrusions in Antarctica; Stillwater, USA; and the Rum Complex, Scotland, among others.



Figure 8.3 Alternating layers of olivine rich (peridotite) and plagioclase rich (gabbro/leucogabbro) in the central complex of Rum, Scotland.



Figure 8.4 Layering on a variety of scales picked out by modal changes in abundance of key phases, for example olivine, plagioclase and pyroxene. Western layered series of Rum, Scotland.

Layers and laminae (see Section 2.4.2 for definitions) vary greatly in composition, texture and lateral extent. They may be *graded* in terms of mineral content and grain-size, and the commonest case is for upward-grading from dense to less-dense (dark- to light-coloured) minerals, or from coarse- to fine-grain size (Figures 8.3 and 8.4).

The contacts between layers vary between sharp and gradational, and between concordant and discordant with the adjacent layer. Both layers and laminae may be qualified by the terms ‘*grain-size*’, ‘*modal*’ (that is, mineral proportion) and ‘*textural*’ for a field description. In many intrusions, the layering may show complex medium-scale structures, such as channelling, cross-stratification (of discordant layers), slumping and complex convolution within layering (for example, Figures 8.5 and 8.6). In some cases, xenoliths of early-crystallised cumulate appear to have slumped into younger layered cumulates. These features are similar to common sedimentary structures and their occurrence provides evidence of processes that deform and slump soft sediments and do *not* necessarily require crystal settling.

Due to the abundance of layering and other ‘sedimentary’ type features (for example, slumping) that are common in mafic/ultramafic intrusions, they can be very well expressed by using graphic logs (see Chapter 2). An example of such



Figure 8.5 Complex slumping folded structure, formed by soft sediment deformation of the crystal mush. Western layered series of Rum, Scotland.



Figure 8.6 Flame structure caused by loading of more dense darker mush on plagioclase-rich lighter mush in the central layered series of Rum, Scotland.

MAFIC COMPLEXES

**MINERALOGY/
ROCK TYPE**

 opx (cpx, sp)
Orthopyroxenite

 opx, cpx (sp)
Websterite

 opx, cpx, gt (sp)
Garnet Websterite

 cpx, gt (opx)
Garnet Clinopyroxenite

 cpx, gt
Garnet Clinopyroxenite

opx

 cpx, gt, graph, (opx)
Graphitic Garnet
Clinopyroxenite

cpx, gt

 ol, opx, cpx, gt
Garnet Clinopyroxenite
enclosing Garnet Lherzolite

 cpx (gt)
Thinned (Garnet)
Clinopyroxenite

cpx, opx

Websterite

 cpx, cor
cpx, gt, cor

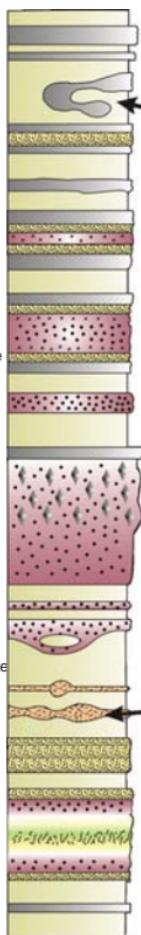
 Corundum Clinopyroxenite
Corundum Garnet
Clinopyroxenite


Figure 8.7 Graphic log with linked photos highlighting modal layering and textural variation from the Beni Bousera Peridotite Massif, North Morocco (courtesy of Graham Pearson).

a log is given in Figure 8.7, with example field photos indicated on the log. A combination of such graphic logs and detailed field sketches are often required, as the range of layering and other features can be on a variety of scales (for example, Figures 8.4 and 8.7).

In describing layered rocks, large-scale repetition of particular types of layering may be observed. Such repetition is usually conspicuous, involving the systematic recurrence of distinctive layers or sequences of the same kind of

layers, and is termed *rhythmic layering*. These were originally described from the Skaergaard intrusion, which is characterised by the occurrence of modally graded units, 5–50 cm in thickness. Finally, many cumulate rocks are also associated with coarse-grained segregations, such as *pegmatite* (for example, Section 2.5.2, Figure 2.15) which are potentially important as they can contain minerals that can be used to date the intrusions. Later minor intrusions, such as dykes and sills (Chapter 6), often cross cut layered intrusions and are useful in determining the relative ages of intrusion where clear cross cutting relationships can be established (for example, Chapter 6, Figure 6.7).

On a broad scale, in most of the large-scale intrusions, the proportion of felsic material increases upwards, in some cases giving way to granitic rocks in the uppermost layers. Individual layers vary upwards from a few centimetres (most typical) to metres and perhaps tens of metres (exceptionally in the upper regions of an intrusion where the contrast between layers tends to be much less than near the base). *Accurate observations of band orientations, widths and their variation normal to the layering therefore provide important criteria for later interpretation of fieldwork.*

Other mafic-ultramafic dykes, sills and plutons of all sizes occur in the continental crust. They may have been emplaced tectonically during metamorphism and thrusting, in which case they will be associated with sheared, deformed and high-grade metamorphic rocks. More simply, they may have been intruded as part of a larger igneous province (for example, the Cuillin gabbros of Skye). Various combinations of the mafic-ultramafic rock-types described earlier are known: some intrusions are homogeneous gabbros, peridotites or even dunites, others show igneous lamination, and many are complex, multiple intrusions with mapable internal boundaries. Small intrusions of dunite are sometimes found in high grade gneiss zones where they may be only a few tens of metres to a few kilometres in size, elongated parallel to the surrounding foliation and may, themselves, show syn-tectonic folding and foliation. They may be extensively serpentinised (Figure 8.2), particularly around their margins and, as well as serpentine, may also contain a range of secondary minerals such as chlorite, talc and magnesite.

Finally, it should be noted that the coarse-grained micaceous peridotites, known as kimberlites, that occur in deeply-eroded volcanic diatremes are classified as ultramafic intrusive rocks. These are considered in detail in Chapter 6.

8.3 Ophiolite Complexes

Many tectonically-emplaced mafic-ultramafic bodies have an origin as slices of oceanic lithosphere. Such bodies are termed *ophiolites* and are preserved as a result of thrusting and/or uplift associated with plate tectonic processes (cf. Chapter 1). Since these ophiolite complexes have assumed great importance in interpreting the tectonic history of a region, it is useful to explain

MAFIC COMPLEXES

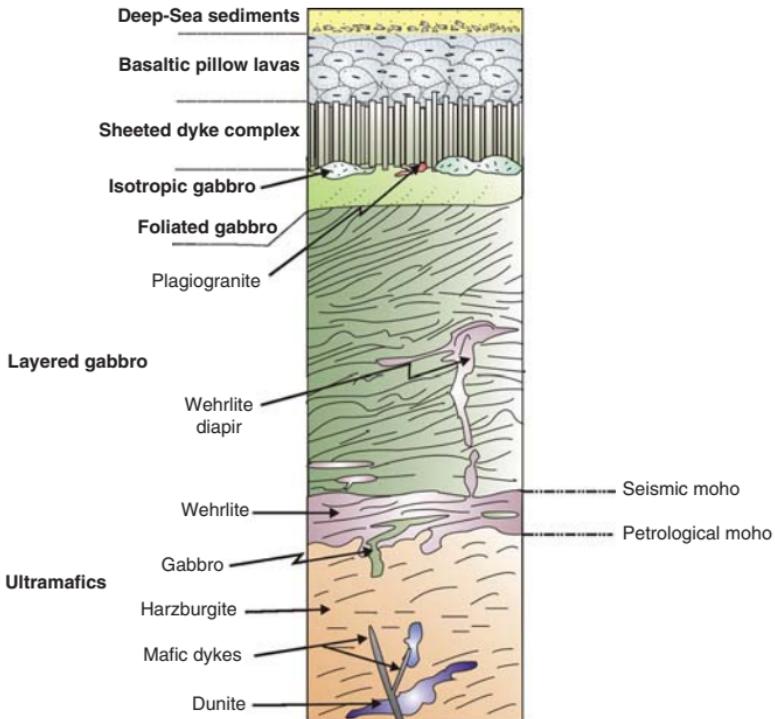


Figure 8.8 Schematic section through a typical ophiolite sequence (further details in text).

the field and petrological characteristics that enable them to be recognised. An *ophiolite* is defined as an assemblage of mafic and ultramafic rocks, ideally with a stratigraphy as in Figure 8.8. Thus a *complete* ophiolite sequence would provide representative samples of all the layers inferred to comprise oceanic lithosphere, but most mafic and ultramafic bodies with characteristics of ophiolites do not form complete sequences, possibly as a result of disruption during tectonic emplacement (for example, Figure 8.9). However, once several of the following characteristics have been found in the field, some confidence may be placed in the identification of an ophiolite.

First, ophiolite bodies are rarely found alone but tend to occur as intermittent elongate bodies aligned along a zone of major thrusting/faulting, such as the Alps and Himalayas. Within each ophiolite body, the rock-types often become progressively more ultramafic towards the base (remember, though, that the sequence may be dipping or even inverted due to tectonic emplacement). Ideally, the base may be seen to rest on a thrust zone which may be identified in the

MAFIC COMPLEXES

	Ocean Crustal Layers	Typical Ophiolite		Normal Ocean Crust	
		Thickness (km)	P-wave velocity (km/s)	Thickness (km)	P-wave velocity (km/s)
Sediments	1	~0.3	—	0.5	1.7 - 2.0
Pillow lavas	2A & 2B	0.5	3.3	0.5	2.0 - 5.6
Sheeted dykes	2C	1.0 - 1.5	4.1	1.5	6.7
Gabbro	3A	2 - 5	5.1	4.7	7.1
Layered gabbros	3B				
Layered peridotites					
Upper Mantle	4	<7	2.5-4.2	—	8.1

Figure 8.9 Comparison of ophiolites and normal ocean crust in terms of real and preserved thicknesses and P-wave velocities (used in seismic).

field by the presence of deformed and fragmented ultramafic rocks within the ophiolite, overlying, usually with sharp petrological and structural contrast, the local crustal rocks.

The *basal sequence*, itself the most variable in thickness, ranging up to 10 km normal to the dip of the base, may comprise generally unlayered mixtures of serpentinised peridotites, usually with strongly-deformed, ‘tectonised’ fabrics, but sometimes including zones of recognisable dunite and harzburgite. These rocks are interpreted as a ‘mantle sequence’ from which the partial melts that formed the overlying (ocean crustal) layers have been extracted (Figure 8.10).

Overlying the mantle sequence, there are usually several kilometres thickness of *layered mafic-ultramafic rocks*, resembling those of layered lopoliths (for example, Figure 8.11). Near the base, cyclic successions of dunites, wehrlites and other peridotites or pyroxenites may occur and there is often evidence that repeated influxes of magma at different levels within the layered sequence each generated a cyclic unit. Thus, whereas the majority of the layered sequence comprises interlayered gabbros, norites and olivine gabbros, with increasing amounts of leucogabbro and anorthositic material towards the top of the layered sequence, parts of the sequence may be repeated due to multiple intrusion. The white ‘anorthositic’ rocks of ophiolites are usually rich in sodium-rich plagioclase feldspar and some of them contain substantial amounts of quartz (up to 30%), so that the term *plagiogranite* is more appropriate. The layered sequence is usually interpreted as representing an ocean crustal magma chamber which, in the same way as it received successive magma batches, gave rise to the overlying dykes and lavas by venting and emplacement of magma from its upper surface.



Figure 8.10 Transition from olivine rich mantle rocks (dunites, harzburgites) to plagioclase bearing peridotites and gabbros in the basal sequence, Cyprus.



Figure 8.11 Layered gabbro section, Cyprus.



Figure 8.12 Sheeted dyke complex from the Cyprus ophiolite.

The top of the layered sequence is marked by a decreasing proportion of coarse-grained rocks and the appearance of medium- to fine-grained dyke rocks trending approximately at right angles to the layering below. Higher in the ophiolite sequence, these dykes rapidly become a *sheeted complex*, in which almost all the outcrop may be occupied by interpenetrating circa 1 m thick sheets of fine-grained gabbro (dolerite) or basalt (Figure 8.12). The sheeted complex of dykes may be circa 1 km thick before it gives way upwards to a series of pillow lavas, typically circa 0.5 km thick, though feeder dykes may continue through much of the lava sequence. Basalts predominate in the pillow lava sequence (for example, Figure 8.13) though more mafic picrites or more acid andesites and dacites may occur. Thick non-pillowed flows and occasional sills may also be found in the sequence, as well as fragmented pillows, hyaloclastites and more isolated dykes (for example, Figure 8.13). The uppermost lavas may be intercalated with various types of deep-sea sediments, ranging through cherts, shales and mudstones to occasional red-brown ferromanganese hydroxide sediments, known as *umbers*. A thin sequence comprising purely sedimentary rocks is sometimes preserved at the top of an ophiolite body.

Finally, the top 3–4 km of an ophiolite sequence is likely to show the effects of hydrothermal metamorphism due to the penetration of magmatically-heated

MAFIC COMPLEXES

10. Mineralisation/Geotechnical

7. Granitic Complexes

8. Mafic Complexes

9. Magma Mixing & Mingling

6. Shallow-level Intrusions

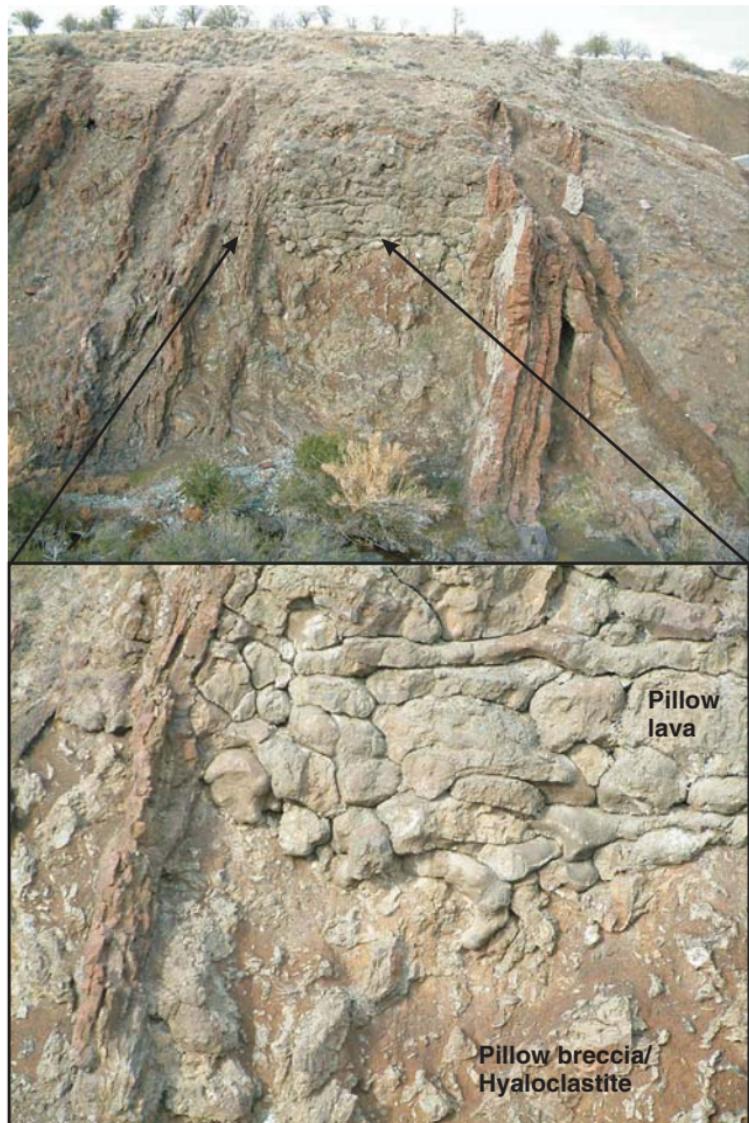


Figure 8.13 Dykes cutting through pillow lava sequence, from the Cyprus ophiolite. Blown up image shows beautifully preserved pillows above pillow-breccia/hyaloclastites (cf. Figure 4.18).

sea water through cracks and fissures when the ocean crust was newly formed. The pillow lavas and dykes of ophiolites frequently show minerals of the zeolite and greenschist facies of metamorphism, whereas other dykes, lower in the sequence, and the layered gabbroic rocks, may show amphibolite facies minerals. This means that, on a mineralogical scale, dark-green amphibole, originally at deeper levels, and pale-green chlorite, originally at shallower levels, should be recognised – both of which grew at the expense of pyroxene. White zeolite minerals may also be found in cavities or, more rarely, replacing feldspars in the dyke-pillow lava sequence. During hydrothermal metamorphism, in general, it is common for base metals to be remobilised and concentrated into ore deposits (see Chapter 10), and the upper regions of ophiolite sequences may contain *stockworks* along which mineralising fluids are able to penetrate and which are often lined with chalcopyrite and other sulfide minerals. Where a major fracture focused the flow of mineralising fluids into a sea-bed discharge zone, reaction between the hot, metal-enriched fluids and sea-water may have caused massive ‘exhalative’ sulfide ore bodies to be precipitated at the upper surface of the ocean crust; such sulfide ore bodies are also known from ophiolites. More commonly, however, the metals were precipitated as Fe-Mn-rich hydroxide umbers, carrying lesser amounts of copper, lead, zinc and other transition metals. So ophiolites can be economically important, not only for their deep chromite layers, but also for their high-level stockworks, massive sulfide mineralisation and the related sedimentary umbers.

8.4 Komatiites

Komatiites are a rare but important ultramafic rock type found predominantly in the Archaean. They contain very high MgO contents above 18%, as high as 32%. Komatiites are commonly associated with Ni-Cu-PGE-bearing sulfide deposits (for example, Yilgarn craton, Western Australia; Abitibi greenstone belt, Canada), and have also been used extensively in helping to determine mantle potential temperatures as they are thought to represent very hot primary melts. The rocks occur as lava flows and shallow intrusions, and are characterised by some key textural relationships. Figure 8.14 shows a schematic section through a komatiite lava with specific zones recognised by different olivine textures. The lower B2 zones contain mainly euhedral olivine crystals which were present as phenocrysts during the eruption of the komatiite and have settled out on emplacement. The most characteristic texture of the komatiite is the bladed interlocking olivine texture known as spinifex (after the long grass of the same name), where long olivines rapidly grow from the top of the flow downwards. Such textures are thought to form due to the high temperatures and super saturation of Mg in the melt. It should be noted that not all of the zones defined in Figure 8.14 will be present in a Komatiite, and careful mapping may be required as komatiites are commonly very weathered due to their old age and MgO content.

MAFIC COMPLEXES

6. Shallow-level Intrusions

7. Granitic Complexes

8. Mafic Complexes

9. Magma Mixing & Mingling

10. Mineralisation/Geotechnical

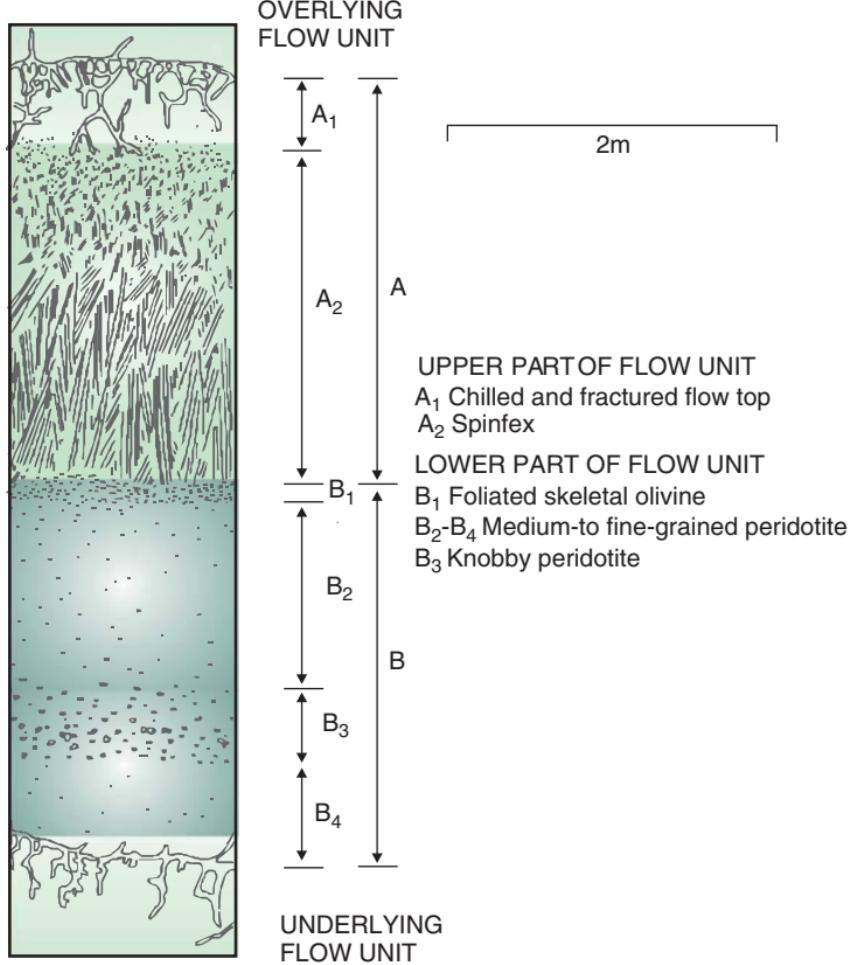


Figure 8.14 Classic schematic section through a komatiite flow highlighting the spinifex horizon (zone A) and cumulate horizons (zone B).

8.5 Summary of the Field Characteristics of Mafic-Ultramafic Intrusions

1. Medium- to coarse-grained gabbro ($\text{cpx} > \text{opx}$), norite (with $\text{opx} < \text{cpx}$) and ultramafic rocks occur as isolated within-plate intrusions, and also as part of ophiolite sequences that are tectonically emplaced into continental suture zones.

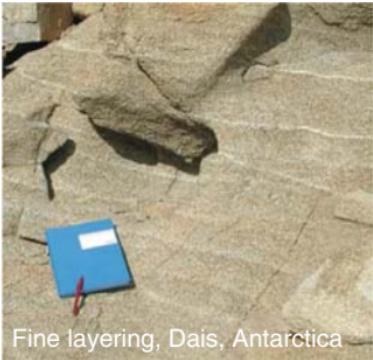
2. Large mafic-ultramafic intrusions often display excellent igneous layering and lamination. Individual layers may range petrologically from the extremes of anorthosite and economically-important chromitite, through a wide variety of peridotites, pyroxenites and gabbros; *the layers should be described in terms of grain size, modal mineralogy and texture*. Penetrative, disruptive and ‘sedimentary’ slump structures are frequently displayed by the layering, especially where strong density contrasts (for example, between peridotite/gabbro and anorthosite) exist between the alternating layers. In general, large mafic-ultramafic intrusions become more felsic upwards, and residual liquids from their fractional crystallisation may evolve towards granitic compositions.
3. Most ophiolite complexes are igneous masses of elongate outcrop caught within known continental sutures. They are generally most ultramafic towards the base, and from the bottom upwards should comprise combinations of:
 - a. a basal ‘mantle’ sequence of deformed dunite, harzburgite and pyroxenite,
 - b. a layered mafic-ultramafic sequence that becomes more leucocratic upwards,
 - c. a sheeted complex of dolerite dykes and
 - d. basaltic pillow lavas overlain by Fe-Mn umbers.

The effects of hydrothermal metamorphism should be recognised in the uppermost 3–4 km; high-level mineralised stockworks and, less frequently, massive exhalative sulfide ore bodies may also be found.

4. In ancient greenstone belts highly mafic volcanics can be found in the form of komatiites. These have characteristic spinifex textures and can be associated with mineral deposits.

MAFIC COMPLEXES

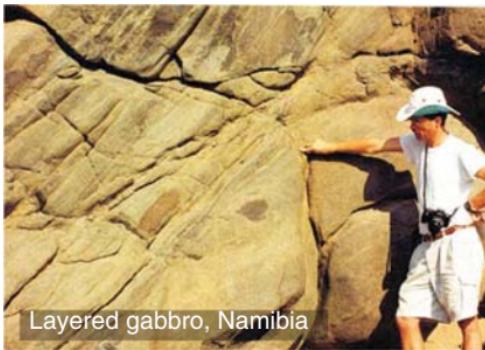
6. Shallow-level Intrusions



7. Granitic Complexes



8. Mafic Complexes



9. Magma Mixing & Mingling

10. Mineralisation/Geotechnical

Mafic/ultra-mafic - Examples



Mingling and mixing magmas can lead to some surprising results!!

9

MAGMA MIXING AND MINGLING

Igneous petrologists have long recognised that plutonic rocks in the field, and granitic rocks in particular, contain within them a variety of lithologies that are either not of igneous origin, or are igneous but appear compositionally distinctive from their host. The generic name given to this extraneous material is enclave and five broad categories are now recognised: (i) xenoliths, fragments of older country rock caught up during magma ascent and emplacement, (ii) magmatic enclaves, originally liquid or semi-liquid blobs derived from comagmatic intrusions, (iii) restite, solid material derived from the magma source region (iv), autoliths, cognate material derived from the host but differing slightly in composition, (v) xenocrysts or anitcrysts, individual crystals or accretions of crystals or pheoncrysts transferred into the host magma during mingling or mixing, or from a different part of the intrusion (for example, crystal mush on the chamber floor or walls). The structural and textural relationship between enclave type and host, if decoded and read properly, can reveal a wealth of information about the intrusive history of the magmatic system under investigation. For example, depending upon the rheology and relative volumes of host and invading magma, the nature of the preserved interactions range from near complete mixing and blending of two compositionally distinct fluids with similar viscosities to late stage dyking of fluid mafic magma into a rigid, near-crystallised plutonic host. This open system behaviour, where mafic and silicic magmas have the opportunity to interact more-or less continuously at any level from the base of the continental crust right through to eruption at the surface, is now regarded as commonplace. However, confusion can arise when, for example magmatic enclaves are misidentified as metasedimentary xenoliths, and vice versa, and where the terms magma mixing and magma mingling are used interchangeably to describe the same thing (they are in fact quite different). Thus when faced with unravelling the complexities of co-genetic magmatic systems in the field, caution and careful field observation are required.

9.1 Magma Rheology

Magmas are complex, multiphase fluids that contain solids (crystals, bits of pre-existing rocks), dissolved gas and bubbles, carried in a continuous liquid (melt) phase. The physical properties of magma strongly control their ability to

Table 9.1 Relationship between composition, physical properties and intrusion temperatures of crystal-free melts.

	SiO ₂ range (wt. %)	Average melt density range (kg/m ³)	Average melt viscosity range (Pa s)	Typical intrusion temperature range (°C)
Basalt/Gabbro	45–52	3.00–2.80	1–10 ³	1000–1200
Andesite/Granodiorite	52–65	2.80–2.60	500–10 ⁴	800–1000
Rhyolite/Granite	65–75	2.60–2.40	10 ⁴ –10 ⁹	700–900

flow and mix during ascent and emplacement. Two key properties relating to the liquid (melt) part are viscosity and density. Melt viscosity increases with SiO₂, making rhyolite more viscous than basalt (Table 9.1). Both also depend on temperature and other factors including strain rate and amount of dissolved volatiles (for example, H₂O, CO₂, Cl, F, SO₂). In general, the viscosity of the melt phase decreases with falling temperature and increasing volatile content, regardless of composition. Density is also related to composition, with mafic melts more dense than acidic ones while increasing water content lowers density overall regardless of composition. However, this is just part of the story as, by definition, magmas are mixtures of melt plus solids (crystals), where the volume fraction of suspended crystals can affect profoundly the overall effective viscosity of the magmas and hence their ability to flow and mix together. Generally, magmas with crystal contents <30% behave rheologically similar to crystal-free liquids. However, as the solids fraction exceeds about 40%, magma viscosity can increase rapidly by several orders of magnitude at least. This fact is self evident as by definition magma with 100% crystals is solid rock. The sensitivity of congested magma to small changes in crystallinity means that mingling and mixing phenomena are expected to be characterised by threshold behaviour, meaning that rheological transitions (for example, fluid-like to solid-like behaviour) in response to factors such as strain rate will not only be sudden but reversible, although overall as cooling proceeds, magma-magma interactions become generally more brittle-like. A further added complication is that a basaltic magma with a high crystal load (about >40%) can actually be *more viscous* than a crystal-free or dilute granitic magma (Figure 9.1). Thus, despite the general rule that acidic magmas are more viscous than mafic ones, like all rules this one can on occasion be broken where temperature, crystal content and compositional evolution are competing at a local level.

9.2 Magma Mixing

Magma mixing implies that two coexisting magma types have come together and blended to produce a new, hybrid magma composition, that is intermediate

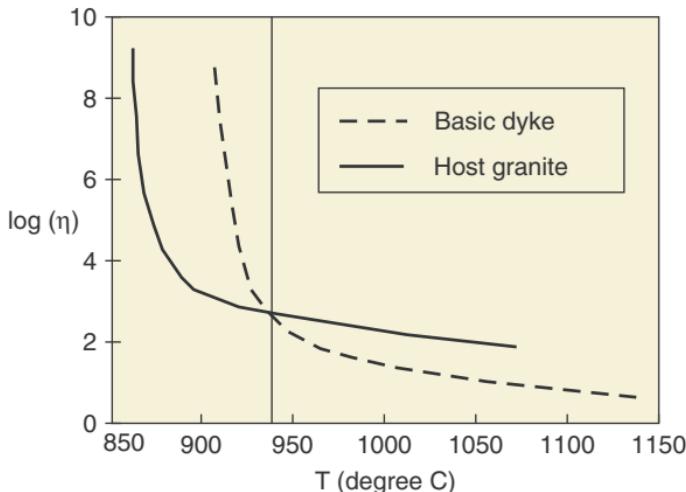


Figure 9.1 Variations of $\log \eta$ (in Pascal seconds) as a function of temperature for two magma types corresponding to basalt and granite (s.l.). The temperature where both magmas reach local thermal equilibrium (about 940°C) corresponds to a viscosity inversion, below which the originally more viscous granitic magma tracks the viscosity increase of the basaltic one. In this model calculation, the local equilibrium temperature corresponds to a crystal content of about 20% for the granite and 40% for the basic magma. The closer both magmas are to the inversion temperature, the easier it becomes for them to mix and blend as opposed to mingle.

between the two parents. An example would be basalt and rhyolite mixing together to make andesite. Just in the same way that oil and water don't mix, no matter how hard you stir, so it is with magmas where contrasting density and viscosity ratios prevent it. In magmas where these contrasts are low it is possible for mixing and blending to occur. Where the viscosity contrast remains high, magmas are effectively immiscible and co-exist as discrete entities with little chemical interaction, although material interaction in the form of crystal exchange is still possible. The timescale of cooling is an important factor controlling the degree of interaction and mixing, as is the volume of mingling material and the deformation history. Very fast cooling rates will inhibit mixing but shearing, if accompanied by stretching and folding of interacting fluids will enhance their ability to blend and homogenise by increasing the contact surface area available for chemical and thermal diffusion. This type of interaction is likely to be most effective when magma is flowing at speed in narrow dikes or conduits. Where magma mixing is suspected in the field, this can be tested in the laboratory using geochemistry, as whole rock major oxides and

trace elements form linear trends on Harker diagrams between two parent end member compositions. Isotopic data will show similar trends. When collecting samples for geochemical analysis it is vital therefore to select those rock types that reflect most closely pristine end member compositions.

9.3 Magma Mingling

Magma mingling describes the process whereby co-eval magmas of different composition come into contact but do not mix and blend. Instead they retain their compositional integrity and freeze as discrete, identifiable enclave types (Figure 9.2). This does not however imply that no exchange of material has occurred, and a defining characteristic of magma mingling is the ability of crystals to move from one magma type to another. Key observations include fluid-fluid contacts showing clear evidence of the higher temperature from (generally the invading mafic component) having chilled against its felsic host. The idea that mafic magma can intrude into incompletely solidified granitic magma and become frozen in a fashion analogous to pillow basalts in water might sound odd but where temperature contrasts fall in the range of several hundred degrees or more (Table 9.1) then local quenching is inevitable. The geometry of magmatic enclaves formed during magma mingling varies considerably, as do the relationships between enclave margin and host rock. Margins in particular require careful observation and can vary from sharp, brittle-like, through a range of cuspatc and cauliform types indicative of fluid-fluid mingling, into diffuse, irregular contacts indicative of blending and mixing.

Enclaves are often associated with schlieren and layering in grantitic plutons (see Chapter 7) and may cross-cut one another, implying successive episodes of magma mingling. In such cases it is possible to find enclaves with only a faint outline remaining (ghost enclaves), possibly the result of partial assimilation into the host magma due to prolonged exposure. Given the complex rheological nature of congested magmas (Figure 9.1) it is not uncommon to find evidence for simultaneous synmagmatic fracturing and viscous, fluid-fluid contacts between magmatic enclaves and host rocks in one small area of outcrop. What appear to be contradictory field relationships with respect to timing of intrusion are often displayed in the same magmatic enclave population (Figure 9.3).

9.3.1 Magma mingling: field example

An example of a pluton containing magmatic enclaves and metasedimentary xenoliths in close proximity is the Ross of Mull granite, Scotland. Here a zoned pluton was intruded into Precambrian metasediments of the Moine supergroup. Petrographical and geochemical analysis have revealed the fine grained magmatic enclaves are dioritic in composition. The dioritic enclaves are located

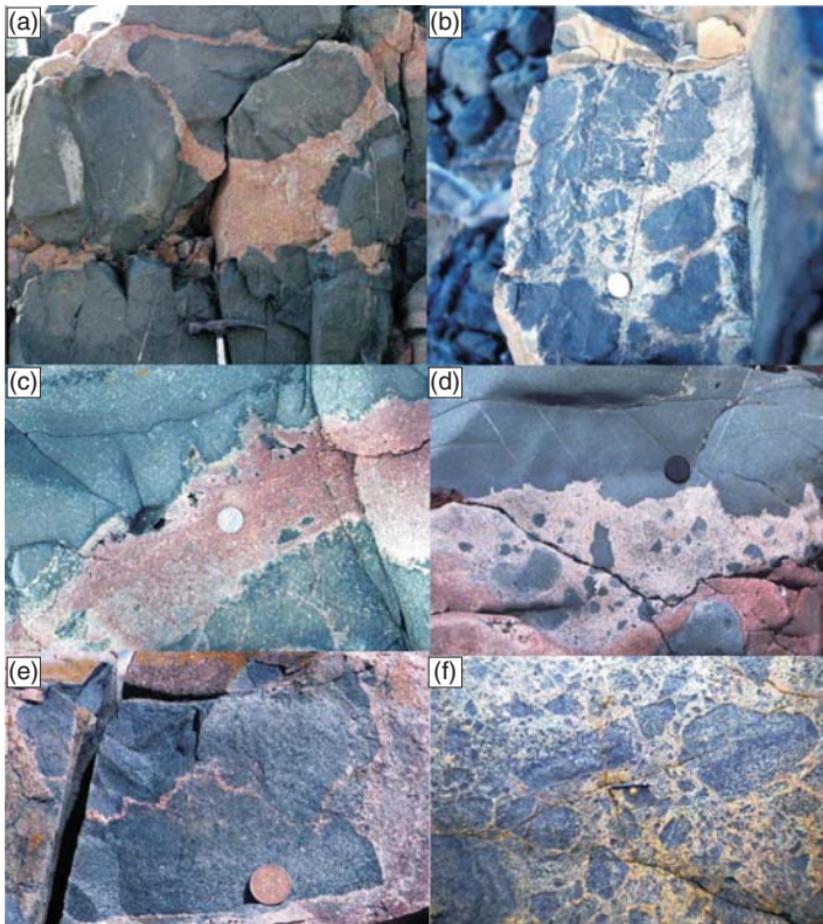


Figure 9.2 (a) Classic magma mingling showing ‘pillows’ of microdiorite with chilled margins intruded into pink granophyre, Jersey, Channel Islands. (b) Fine-scale dendritic structures in mafic enclaves resulting from ‘viscous fingering’ (disruption of mafic fluid within a non-Newtonian granitic host), Maine, USA. (c) Detail of cuspatate (fluid-fluid contacts) typical of magma mingling in the field. Cusps generally point towards the magma with the lowest viscosity, Jersey, Channel Islands. (d) Magma mixing and hybridisation zone between diorite and co-mingling pink granophyre, Jersey, Channel Islands. (e) Fine-scale, wispy granitic vein cutting angular diorite enclave, Jersey, Channel Islands. (f) Agmatite – magma ‘breccia’ of shattered, jigsaw-like mafic enclaves in a granitic host indicative of magma mingling at high strain-rates, Jersey, Channel Islands.



Figure 9.3 Type example of contemporaneous brittle-ductile magma mingling. Blocks of angular mafic enclaves appear to be wedged apart by invading, fluid acid magma (brittle-like) while occupying the same space as cauliform magma pillows indicative of more ductile, fluid-fluid interactions, Maine, USA.

in the southern and south-western sector of the pluton and range in size from less than 5 cm to several meters in diameter. They are either ovoid or ellipsoid in shape, with aspect ratios between 2 : 1 and 50 : 1. From careful field observation, the magmatic enclaves can be divided into two distinct populations: (i) megacrystic and (ii) megacryst free varieties. The megacryst free enclaves are composed largely of amphibole (~30%), biotite (~20%), albite twinned interstitial plagioclase (~45%) and show recrystallisation textures. Enclaves can be further subdivided into two groups depending on the megacryst assemblage: (i) plagioclase + quartz and (ii) plagioclase ± hornblende ± quartz ± alkali feldspar. Mantling of the megacrysts by coronas of plagioclase or amphibole ± biotite is common. From this it has been possible to map out on a large scale using an evenly-spaced 1 m grid, the spatial distribution of enclave type in the central region of the pluton (Figure 9.4).

9.3.2 Magmatic enclaves as strain markers

Magmatic enclaves in deformed plutons may become elongated parallel to the plane of foliation and can, with some caution, be used as strain markers in the same way rigid inclusions are used in tectonically deformed rocks (for strain analysis, see McClay this book series). The Ardara pluton, Donegal is a good example of where enclaves have been used to estimate strain during magma emplacement

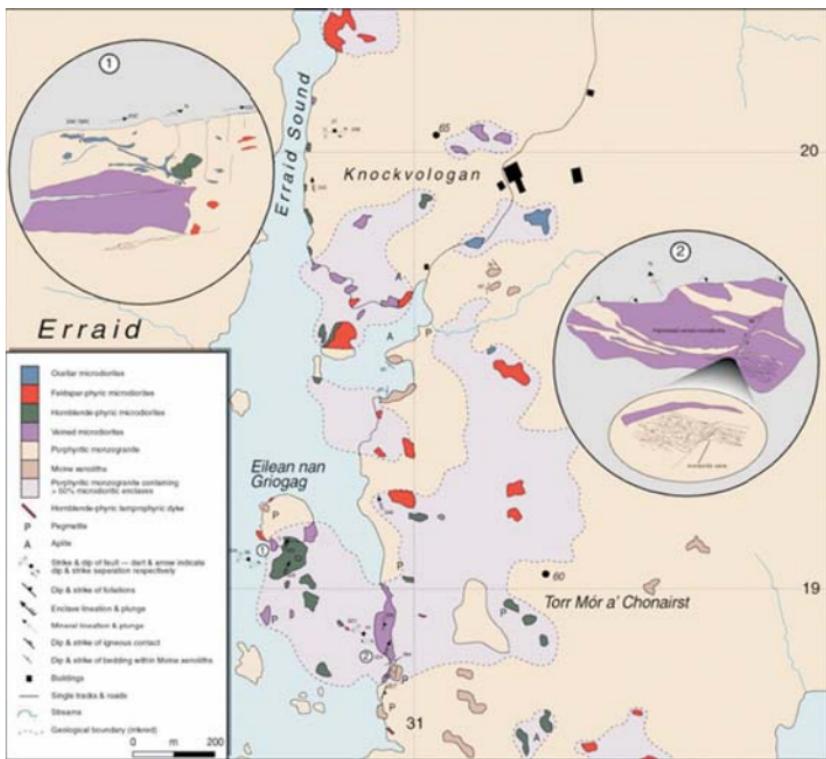


Figure 9.4 Large scale (1:200) map of the central part of the Ross of Mull Granite, Scotland, showing the distribution on enclave types base on mineralogy, and structural information on deformation fabrics in the host granite.

(Chapter 7, Figure 7.22). To do this requires measurement of the aspect ratio and orientation (dip and strike) of the long axis of individual enclaves. The degree of flattening can then be related to the amount of deformation needed to cause this (one weakness in the method is that traditional strain analysis assumes an originally spherical inclusion, most magmatic enclaves in undeformed plutons may be oblate to start with). When combined with other structural data such as S-C fabrics, deformed enclaves showing asymmetry can also provide information about the sense of shear. Careful interpretation of the relationship between enclave and host rock fabric is required, but with careful detailed mapping it is sometimes possible to use both to determine the shear sense during emplacement where the magmatic fabric is unanimously pre-full crystallisation (Chapter 7, Figure 7.11). It is often the case that deformation, hence enclave elongation, is most intense at or near contact zones, whether at the margin of a pluton with its country rock (Figure 9.5) or an internal contact where the pluton is sheeted (Figure 9.6).



Figure 9.5 Strained enclave in marginal tonalite facies flattened with long axis in plane of internal foliation, Ardara pluton, Donegal (see also Figure 7.22, Chapter 7).

It is also worth noting that syn-magmatic fabrics can be spectacularly developed where host magma and enclave are in close proximity. Shearing textures appear to be an almost inevitable consequence of magma mingling and at times the fabric development can be quite intense, resembling deformation fabrics (for example, mylonites) more commonly associated with tectonic deformation of completely crystalline rock. However, the highly localised nature of the sheared fabrics rules out an origin through large scale, pervasive deformation (Figure 9.7).

9.4 Synplutonic Dykes and Sills

Magmatic enclaves sometimes occur in parallel trains or bands, marking contacts between two different granite types (sheeting) or as remnants of syn-plutonic dykes. These are mafic to intermediate magmas that intrude into partially crystallised granite (*sensu lato*) plutons and are a common feature in the major Batholith terraines of Peru and Chile, and the Sierra Nevada, USA (Figure 9.8). Synplutonic dykes are important for several reasons. Firstly they confirm that crustal extension is an important aspect of batholith construction. In some regions of the Coastal Batholith Peru, individual plutons have undergone over 100% extension as measured by the intrusion volume of synplutonic dykes.



Figure 9.6 Deformed enclaves at contact between two granite facies (upper coarse grained granodiorite) and lower, finer-grained enclave-bearing facies in a sheeted pluton, Peru. Note enclaves are stretched out parallel to the contact.

Secondly, the composition of synplutonic dykes record important changes in magma type and source region that relate overall to the thermal and chemical evolution of plutonic complexes. Finally, it appears likely that in many instances, synplutonic dykes are themselves the source of magmatic enclaves (Figures 9.8 and 9.9). Evidence for this comes in the form of disrupted dykes that intrude into low crystallinity plutons. Here the hotter dyke magma chills and forms blobs and pillows that become dispersed within the pluton magma through a combination of momentum related to dyke intrusion and subsequent thermal convection. Similar effects may occur where mafic magma is intruded at the base of silicic magma chambers. Experimental evidence shows, by analogy, how the injection of basaltic liquid into a silicic magma chamber can lead to instability. Basalt intrusion causes a thermal plume that drags the basaltic liquid to the top of the chamber through a process called viscous coupling. This upwards movement causes mingling and mixing of the liquids in localised areas without influencing the bulk of the chamber. The timing of magma mingling and mixing interactions during different stages of crystallisation of host and invading magma are summarised in Figure 9.10.



Figure 9.7 Localised shearing at contact between mafic enclave and granite host. Fabric intensity drops off away from the enclave margin and is absent in the body of the host granite, implying a synmagmatic origin.



Figure 9.8 Parallel-striking synplutonic dykes of basaltic andesite cutting granodiorite that was only partially crystallised, Coastal Batholith, Peru.



Figure 9.9 Enclave swarm with clearly defined upper and lower boundary, marking the remnant of a synplutonic dyke (Coastal Batholith, Peru).

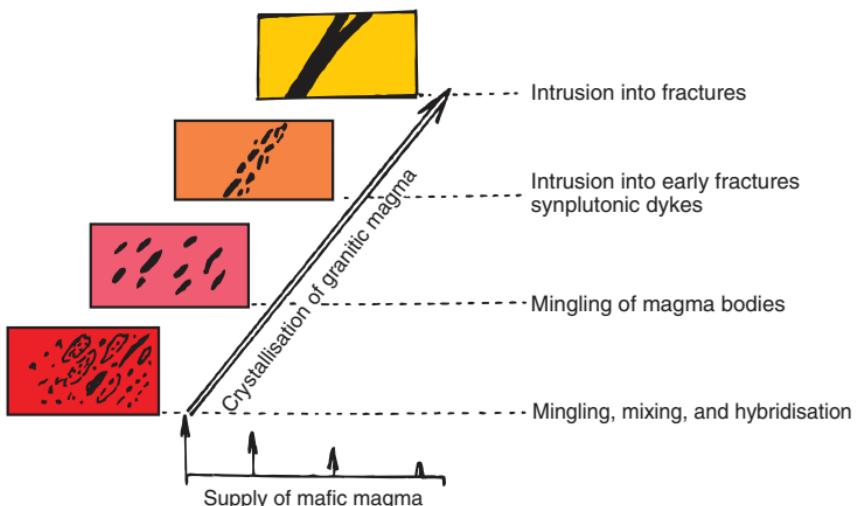


Figure 9.10 Summary of the different magma mingling structures preserved during an encounter between invading mafic magma and its granitic host. The style and degree of interaction is controlled by the crystallinity of the host magma, which will increase with time, and the supply (intrusion) rate of mafic magma.

9.5 Magma Mingling in Subvolcanic and Volcanic Environments

Subvolcanic complexes commonly preserve an enduring record of magma mingling, indeed the concept as a general principle gained widespread acceptance based on seminal field observations of quenched blobs of basalt in silicic ring dykes from central complexes in Slieve Gullion, Ireland and St Kilda, Scotland. The conduits responsible for channelling contrasting magma types simultaneously through the crust are known as composite dykes (refer to Dyke Chapter 6). Composite dykes occur in two forms, normal, where the dyke margins are mafic and the core acidic, or more rarely, reversed, where mafic magma forms the central part with acidic margins. As with normal dykes in the field, dip, strike and overall width should be measured, but in addition the widths of the mafic (or acidic) margins must be recorded, along with any visible evidence for magma mingling or chilling with interior contacts.

There is good evidence that magma mingling and mixing occurs during volcanic eruptions. Indeed magma mingling and mixing arising from injection of hot basalt into partially crystallised intermediate to acidic magma has been cited as a major trigger for the explosive eruptions of Krakatoa in 1883, Pinatubo in 1991 and Montserrat in 1995. There similar evidence for magma mingling at other centres of explosive volcanism such as Long Valley in California and



Figure 9.11 Streaky pumice from Mt Lassen, USA.

Yellowstone in Wyoming. Rapid withdrawal from high-level chambers has the potential to juxtapose magmas of differing composition in a highly dynamic environment where mingling and blending can take place. Compelling evidence for magma mingling and mixing during explosive volcanism is seen in the form of streaky pumice (Figure 9.11).

9.6 Xenoliths

Xenoliths are commonly sedimentary or metasedimentary country rock, often discernable in the field by the preservation of bedding or other sedimentary features. Xenoliths come in a range of shapes and sizes, from amorphous centimetre-scale inclusions to blocks the size of cars or bigger that are still recognisable as sedimentary in origin. Contact relationships between xenoliths and host rock are often brittle-like, characterised by sharp, angular shapes, implying a strong temperature contrast (thermal shattering) or high strain-rates at work during incorporation within the invading magma. Angular xenoliths are often cross cut by brittle veins of invading host magma, that in some cases, can be seen to run parallel with pre-existing bedding. Other tell-tale signs of metasedimentary origin include localised brown or rust-like staining (Figure 9.12). Not to be confused with magmatic enclaves, xenoliths are sometimes used as evidence for magma *ascent* via stoping, but are more likely formed during the



Figure 9.12 Flattened and angular metasedimentary xenoliths of Jurassic shale in host granite, Cordillera Blanca, Peru.

final stages of magma *emplACEMENT*. Their presence and abundance can indicate proximity to pluton margins. For example, xenoliths may occur as ‘trains’ of blocky material that can be traced over a short distance into surrounding the country rock.

In some cases xenoliths may show evidence of high temperature metamorphism and even internal melting. When encountered in the field, efforts should be made to identify their lithology with particular reference to surrounding country rock, the degree of metamorphism and preserved orientation, remembering that depending upon time of incorporation during magma emplacement the xenolith may have been transported some distance from its point of origin, or remain effectively in-situ.

9.6.1 Xenocrysts

The presence of corroded felsic and mafic minerals (xenocrysts) are often found in magmatic enclaves and generally interpreted as the result of mechanical transfer during the mafic-felsic magma interaction. A common type of xenocryst found in mafic to intermediate enclaves that have intruded into granite magma are *oscelli* (Italian for eye), quartz xenocrysts rimmed by biotite or amphibole. Other textural features indicative of magma mingling and mixing are rapakivi feldspars where sodic or potassium feldspar is rimmed by plagioclase (e.g. Chapter 7, Figure 7.23), and minerals of any kind which show evidence of reabsorption triggered by thermal or chemical reactions as a given crystal at home in its host melt suddenly finds itself displaced into magma of a different composition and or temperature under hybrid (enclave) magma conditions (e.g. Figure 9.13). If present together these features point to strong non-equilibrium conditions suggestive of an origin through magma mingling or mixing.

9.7 A Word of Warning

Telling apart xenoliths from magmatic enclaves in the field can at times be very difficult. Both have genetic implications for the origin and source region of granite magmas that warrant caution if you get them wrong. If in doubt the best thing is to record what you can (shape, size, orientation, if possible mineralogy) and take a sample for follow-up petrographical identification. Record the material in your notebook as generic enclave. Also, for magmatic enclaves, do not go by colour alone as a guide to composition. This is because magmatic enclaves range in composition from mafic to felsic. Felsic composition enclaves can and do occur in granitic rocks but look dark. In this case the dark colour is due to the fine grain size not composition and it is dangerous therefore to assume that dark coloured enclaves are always remnants of mafic or intermediate magma. Because of this, many geologists use the non-genetic term microgranular enclave to describe magmatic enclaves in the field.



Figure 9.13 Scale section cut through contact between microdioritic enclave and host granite, Ross of Mull, Scotland. The enclave has been ‘contaminated’ during emplacement into the unconsolidated granitic magma by mechanical incorporation of alkali feldspar and other granitic material.

9.8 Summary

Enclaves are a common feature of plutonic rocks and occur to some degree in most igneous intrusions irrespective of composition. Large screens of country rock are well preserved in mafic intrusions (for example, Skeargård), and magma mixing and mingling are described also from alkali igneous complexes and in carbonatite magmas. However, they are often most spectacularly displayed in granitic plutons and for this reason we have focused attention primarily on these.

Microgranular enclaves are common features of calc-alkaline granitic plutons that form during incomplete mixing of mafic to intermediate melts intruded into cooler granitic magma. The rheological behaviour of magmas of different chemical composition during interaction is governed by their physical properties (density and viscosity), temperature difference and strain rate. Factors such as degree of host and enclave crystallinity exert a significant affect on magma rheology and in particular their ability to mix together to form a hybrid composition (magma mixing) or retain their integrity as discrete blobs or enclaves (magma mingling).

MAGMA MIXING AND MINGLING

These interactions are likely to be geologically instantaneous, occurring over a period of hours to weeks. If interpreted correctly, the nature of the resulting interaction between coeval but contrasting magma types can reveal much about the timing and mode of pluton construction during cooling and crystallisation. Mixing and mingling is also preserved in erupted volcanic products as streaky pumice.

During magma emplacement, a process that requires displacement of pre-existing country rock, incorporation of surrounding material called *xenoliths* (Greek for ‘stranger rock’) is almost inevitable. This class of material picked up from rock hosting an intrusion can itself be igneous where different magma types are coalescing into a batholith, but is more commonly (meta) sedimentary in origin. It is vital not to confuse country rock (metasedimentary) xenoliths with magmatic enclaves in the field. *Xenocrysts* can be crystals of entrapped igneous country rock or crystals delved into the host by invading magma.

As a final point it is worth considering the intrusion geometry in which magma mingling takes place. In the *plutonic or subvolcanic zone*, this is most likely to be predominantly horizontal where mafic magma is injected near to the chamber floor, although where mafic magma is intruded as synplutonic dykes and sills the geometry of mingling may vary widely and apparently chaotically. In contrast, mingling and mixing during volcanic eruptions will take place in sub vertical conduits.

MAGMA MIXING AND MINGLING



Magma mingling - Examples

6. Shallow-level Intrusions

7. Granitic Complexes

8. Mafic Complexes

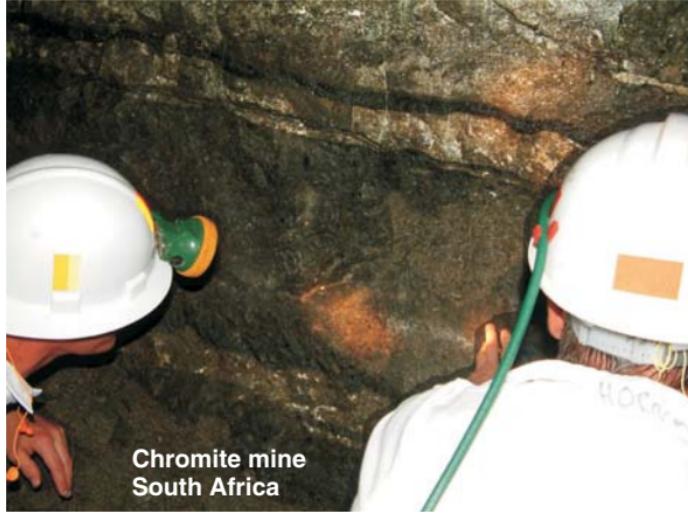
9. Magma Mixing & Mingling

10. Mineralisation/Geotechnical

MINERALISATION AND GEOTECHNICAL PROPERTIES



Chuquicamata copper porphyry mine, Chile



Chromite mine
South Africa

Figure 10.1 Examples of the economic importance of igneous rocks. Mineralisation around intrusions (for example, Chuquicamata mine, Chile). Mineralisation within intrusions (Chromite mine, Bushveld, South Africa).

10

MINERALISATION AND GEOTECHNICAL PROPERTIES

10.1 Mineralisation and Key Minerals

Mineralisation associated with igneous rocks, and calc-alkaline suites in particular, is related to the emplacement of a hot magma body at a high level in the crust which may initiate hydrothermal convection within both the cooling intrusion via veins and fractures and the surrounding permeable country rocks (see discussion of hydrothermal veins in Section 2.5.2). In some cases radiothermal heat associated with K,U,Th-rich granites may stimulate hydrothermal convection of meteoric groundwaters (that is, those derived by downwards-seeping rain-water) long after the intrusion has solidified. Trace elements, particularly base metals (iron, copper, lead, zinc), tin and tungsten, derived from the igneous body and/or from leaching of the country rocks by hydrothermal fluids, may be concentrated and precipitated when the fluids cool as disseminated oxide or sulfide crystals in the igneous rock or as veins, exploiting a joint network that may also invade the country rock. Disseminated mineralisation, usually most prominent near the periphery of a porphyritic diorite or porphyritic granodiorite intrusion, is termed a *porphyry mineral deposit* (notably the massive porphyry copper deposits which are dominated by chalcopyrite and pyrite), and networks of small veins, again usually concentrated near intrusion-country rock contact zones, are known as *stockworks*. Veins usually contain one or more minerals which are abundant but do not have commercial value, and are termed *gangue* minerals (for example, quartz, calcite, barytes) together with a lesser quantity (perhaps a few percent) of metalliferous minerals (for example, Figure 10.2); they may be simple (one generation of minerals) or complex comprising two or more generations, the timing identified by cross-cutting relationships. When mineralised veins are encountered in the field, the nature of the vein or fracture infillings, both metalliferous and gangue crystals and their proportions should be recorded. The width (aperture), orientation (dip and strike) and frequency should be measured in the field. It is also important to note the characteristics of the wall rocks, as explained below.



Figure 10.2 Arsenopyrite, chalcopyrite, stannite, quartz and apatite vein, Cornwall, UK.

Associated with mineralisation, the host rocks become altered by hydrothermal fluids, the effects of which are usually more pervasive and extensive than the mineralisation itself. Therefore, field occurrences of hydrothermal alteration may provide a means of determining whether the region around or within an igneous body might be mineralised. Alteration zones can form on a small scale in the vicinity of veins and fractures infiltrated by hot circulating fluids. Indeed, such alteration halos are regarded by economic geologists as the ‘smoking gun’ – clear proof of hydrothermal activity pointing to the likelihood of mineralised deposits nearby. The nature of the alteration halo around and inside the fracture can be instructive. Narrow veins and fractures are often infilled with secondary clay minerals or green epidote. An extreme form of hydrothermal alteration in granitic rocks is kaolinisation, a process whereby the entire rock mass is altered to secondary clay minerals (kaolin). These kinds of deposits, common in parts of Cornwall are of regional economic significance where the clay is mined to make pottery. Halos surrounding hydrothermal veins comprise phyllitic and propylitic zones where fluid flow is accompanied by dissolution-precipitation chemical reactions. These reactions can be complex but commonly involve the dissolution of feldspar. Plagioclase is particularly

susceptible, especially their calcitic (An-rich) cores. Feldspar pseudomorphs, mixtures of clay minerals, albite and chlorite can sometimes be identified with a hand lens.

Common forms that may be encountered and recognised in field samples are:

1. *silicification*, in which quartz overgrowths and replacements of existing minerals occurs;
2. *kaolinisation*, or *argillic alteration* – the alteration of feldspar to clay minerals;
3. *greisenisation* – the alteration to alkali feldspar to secondary white mica, sericite;
4. *propylitic alteration*, in which plagioclase feldspar alters to epidote mafic minerals to chlorite;
5. *potassium silicate alteration*, in which plagioclase feldspar and amphibole alter to K-feldspar and biotite mica.

For example, an outcrop of greisenised granite with disseminated pyrite (FeS_2) may indicate that deposits of chalcopyrite (CuFeS_2) are not far away. Figure 10.3 illustrates, in cross-section, the way in which both wall-rock alteration (for example, from K– silicate to greisenisation to kaolinisation or propylitic alteration) and the nature of the metalliferous minerals precipitated (for example, from chalcopyrite to pyrite for Cu-Fe bearing veins) both change with falling fluid temperatures and with distance from the centre of a mineralised stockwork. Thus, concentric zones of alteration and mineralisation may be recognised in field studies on a wide range of scales. Table 10.1 lists the properties of some of the more common metalliferous minerals that might be found often in association with calc-alkaline igneous intrusions. Careful examination of the Table 10.1 shows that most of these metallic minerals can be distinguished from one another using colour and hardness criteria.

10.2 Mineralisation in Layered Mafic Intrusions

Many layered intrusions contain chromite-rich seams (chromitites), which host the majority of the world's Cr reserves and may contain significant platinum group element (PGE) mineralisation. Examples with economic deposits include the Bushveld, South Africa and Stillwater Complex, Montana, United States, though many others have occurrences of chromities (for example, Rum, Scotland).

The largest example of a mafic-ultramafic layered intrusion associated with such economic deposits is the Bushveld Complex, shown by the map in Figure 10.4: the intrusion is about 450 km in diameter and about 9 km in thickness. At the margins of the Bushveld, where it is emplaced into

MINERALISATION AND GEOTECHNICAL PROPERTIES

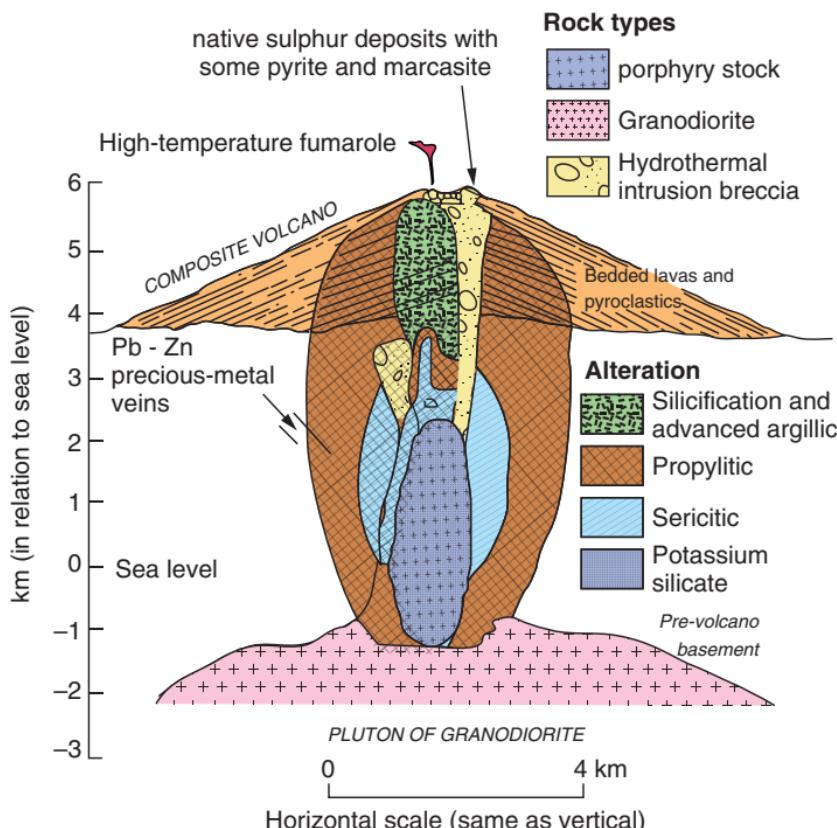


Figure 10.3 Summary of typical zones of wall-rock alteration that occur in and around porphyry copper mineral deposits. Similar alteration zones can occur on a much smaller scale around hydrothermal veins and fractures.

clastic sedimentary and volcanic lithologies, a ‘border group’ has been mapped, comprising thin sills of fine- to medium-grained gabbro, norite and pyroxenite. The vast bulk of the complex is divided into four zones on the basis of rock-type and internal variations (Figure 10.5). The *lower zone*, comprising thick, uniform orthopyroxenite layers, with interlayered olivine pyroxenites and harzburgites, is overlain by a layer termed the ‘*critical zone*’ (so-called because of its content of chromium and hence its critical economic importance), containing interlayered norites and anorthosites near the top with anorthosite-norite-pyroxenite layers further down. Within the critical zone there are several prominent chromitite layers, each up to 1.5 m thick, and many smaller layers (Figure 10.6), and also sulfide segregations, including a layer

Table 10.1 Characteristics of common metalliferous minerals found in association with igneous intrusions.

Mineral	Chemical formula	Usual colour and streak	Lustre	Usual form	Hardness	Density ($\times 10^{-3}$ kg m $^{-3}$)
Chalcopyrite	CuFeS ₂	Brass yellow to golden yellow, sometimes iridescent tarnish; black streak	Metallic	Usually occurs as vein encrustations, occasional small cubic crystals	3.5–4.0	4.1–4.3
Pyrite	FeS ₂	Pale brass yellow to bronze yellow; greenish-black streak	Metallic	Smooth, shiny or striated crystals, often forms good cubes	6.0–6.5	5.0–5.2
Arsenopyrite	FeAsS	Silver grey to steel grey; black streak	Metallic, steely	Massive, occasionally radiating clusters	5.5–6.0	5.9–6.2
Molybdenite	MoS ₂	Lead grey, occasionally greenish-grey, dark grey streak	Strongly metallic	Massive, scaly aggregates of flexible laminae. (NB Very soft, scratched by finger nail)	1.0–1.5	4.7–4.8
Galena	PbS	Lead grey, sometimes tarnishes; dark grey streak	Metallic	Massive cubic crystals, but occasional fibrous crystals	2.5	7.2–7.6

(continued overleaf)

Table 10.1 (continued)

Mineral	Chemical formula	Usual colour and streak	Lustre	Usual form	Hardness	Density ($\times 10^{-3}$ kg m $^{-3}$)
Sphalerite	ZnS	Yellow-brown, good crystals sometimes black; yellow-brown streak	Resinous	Massive, but occasional fibrous or cubic crystals	3.5–4.0	3.9–4.2
Wolfram	FeWO ₄	Black, sometimes dark brown often associated with scheelite; dark yellow-brown streak	Sub-metallic	Tabular, often forms striated, bladed crystals	5.0–5.5	7.1–7.5
Scheelite	CaWO ₄	White, pale green or pale yellow; white streak	Vitreous to resinous (cf. feldspar, but fluoresces in UV light)	Usually massive, occasional rectangular prisms, sometimes platy and striated	4.5–5.0	5.9–6.1
Cassiterite	SnO ₄	White to pale yellow; pale yellow streak	Brilliant on crystal faces, otherwise dull	Massive, fibrous, usually disseminated small grains	6.0–7.0	6.8–7.1

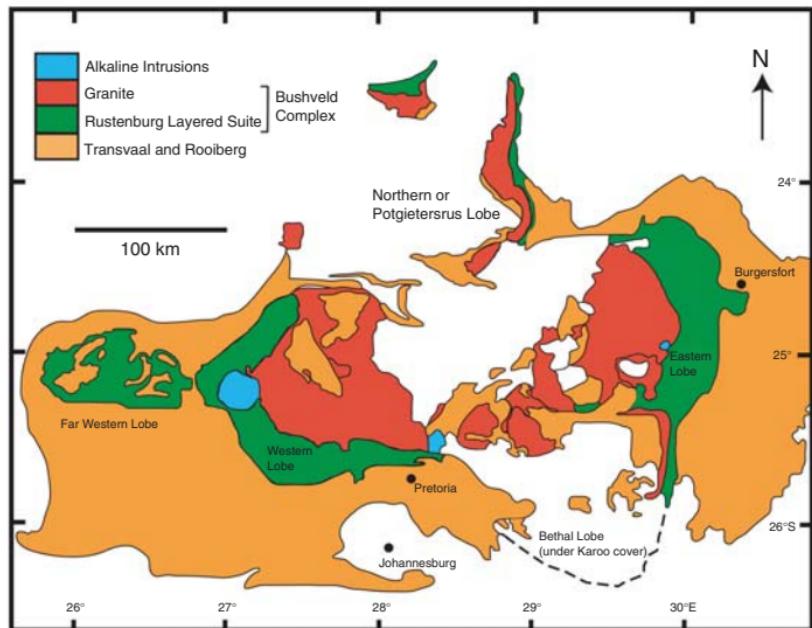


Figure 10.4 General geological map of the Bushveld intrusion and associated rocks, South Africa.

called the ‘Merensky reef’, which contains iron-nickel-copper sulfides and important platinum-group minerals. Above the critical zone is the *main zone* which comprises thick, rather uniform, layers of gabbro and norite, whereas the overlying *upper zone* contains olivine melagabbros, with magnetite-rich gabbros and anorthosite-magnetite layers near the top. A younger granite suite – thought to be unrelated, magmatically, to the Bushveld Complex – overlies and conceals the layered rocks in the centre of the outcrop (Figure 10.4).

10.3 Geotechnical Properties of Igneous Rocks

In this final section we look at the engineering properties of igneous rocks and review some of the on-site field methods used by practising civil engineers as part of routine geotechnical assessments. This applied aspect of igneous petrology has to date received scant attention but we feel it is important for geologists to understand something about the geotechnical properties of igneous rocks for the following reasons:

- Engineers may need to tunnel into a granite pluton or volcanic rock to build a new road or railway line and the characteristics used to define the

MINERALISATION AND GEOTECHNICAL PROPERTIES

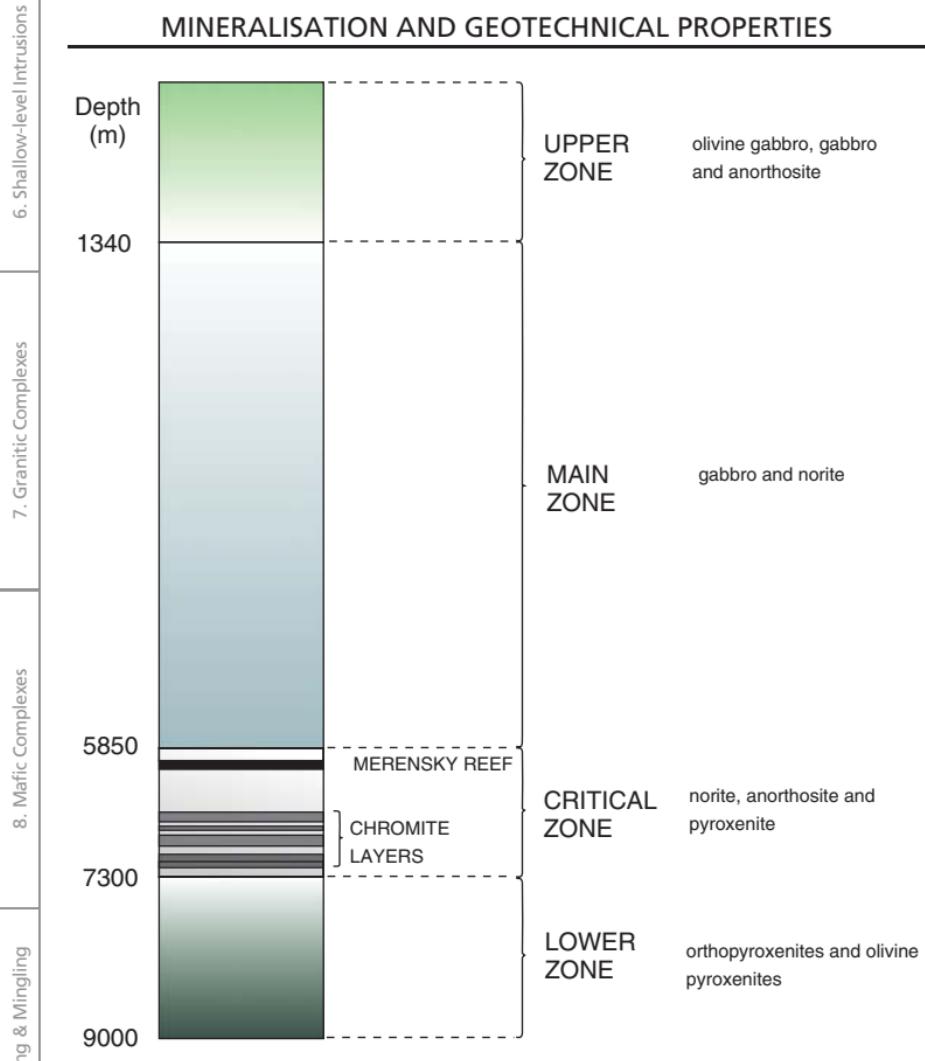


Figure 10.5 Generalised log through the Bushveld sequence highlighting the main zones and chromatite horizons.

geotechnical properties of igneous materials relate directly to their petrology, mineralogy, emplacement and cooling histories.

- ii) Geothermal energy. Fractured plutonic and volcanic rocks can act as geothermal reservoirs (for example, in Iceland). Understanding the detailed mechanical properties of these rocks is important as they determine how effectively and efficiently hot fluids can be transmitted in and out of them (see 10.1). The same reasoning applies when considering the use of igneous rocks as potential storage materials for radioactive waste (for



Figure 10.6 Chromatite layers in two Rivers Mine, Bushveld, South Africa.

example the Borrowdale Volcanic Group at Sellafield in the UK or Yucca Mountain in Nevada, USA).

- iii) Hydrocarbon potential. This may come as much of a surprise to igneous petrologists as petroleum geologists, but igneous rocks can and do act as hydrocarbon reservoirs. Several highly productive plays in South East Asia are extracting oil and gas directly from drilling fractured igneous reservoirs (referred to generically as ‘basement’ by petroleum engineers). As non-traditional sources dry up, these so-called exotic reservoirs may gain more prominence.
- iv) Volcano instability. Studies over the last decade have helped raise awareness of the life and property threatening hazards that result from volcano instability and collapse. Understanding the processes behind catastrophic edifice failure, and mitigating the associated risks, requires knowing about the physical properties, strength and cohesion of the rock mass comprising a volcano.

In the sections that follow, we introduce the key concepts of rock mass strength, intact rock strength, friction, cohesion and show how using simple field estimates it is possible to make some clear predictions on the overall strength properties of igneous rocks in the field as a first step towards a more detailed examination of their geotechnical properties using, for example numerical simulations. We focus here on the rock mass classification of volcanic rocks but note the same principles apply to plutonic rocks regardless of composition.

10.3.1 Fracture formation

Veins and fractures form by several mechanisms and are controlled by the local and regional stress fields acting on the rock mass. We have discussed the formation of cooling joints in plutons in Chapter 7. In brittle rock, fracture propagation is influenced by grain boundaries, smaller cracks and other pre-existing discontinuities. Increased stress at the fracture tip may activate new vein formation within a ‘fracture process zone’ surrounding the propagating crack. Fractures in rock can be classified into two main kinds and it is important to be able to identify each kind in the field as they contain information about the style of deformation responsible for their formation. *Mode I* fractures are formed by tensile stresses and are symmetrical in that little or no displacement occurs either side of the crack opening. *Mode II* fractures form under shear and are asymmetrical (Figure 10.7). In plutonic rocks both kinds can occur as primary fractures during cooling, or afterwards during regional tectonic deformation. Identification of primary Mode II shear fractures formed during cooling can yield important information about the local stress field in operation during pluton emplacement.

10.3.2 Permeability and porosity

The most highly altered igneous rocks occur in active geothermal areas where fracture development has led to elevated permeability and porosity, allowing large quantities of hot fluids to circulate through the rock mass. It is thus not depth beneath the surface but high fracture density and interconnectivity that allow hydrothermal fluids to permeate and alter the composition of igneous rocks.

Hydrothermal alteration of individual minerals can result in the formation of small voids that form a secondary porosity in some altered igneous rocks. While more commonly associated with sediments undergoing diagenesis, secondary porosity can nonetheless lead to a significant increase in the non-fractured permeability of hydrothermally altered igneous rocks. Figure 10.8 shows the range of total porosity in selected igneous rocks where the total is a function of both fracture and void (diffusive) porosity. Porosity and permeability have been measured in volcanic rocks and is a measure of their fracturing, primary gas content and vesicularity (for example, Chapters 4 and 5). Not surprisingly scoria is the most porous and permeable volcanic material (Figure 10.9).

10.4 Rock Mass Classification

At its most basic level, a volcanic edifice is made up from discrete layers of different rock types cut through on a range of scales by pervasive and penetrative discontinuities including faults, fractures and older contact surfaces (Figure 10.10). In addition, different scales of heterogeneity mean that assigning an overall strength is problematic. Characterisation of rock mass properties can be further hindered by long time gaps in volcanic activity that result in

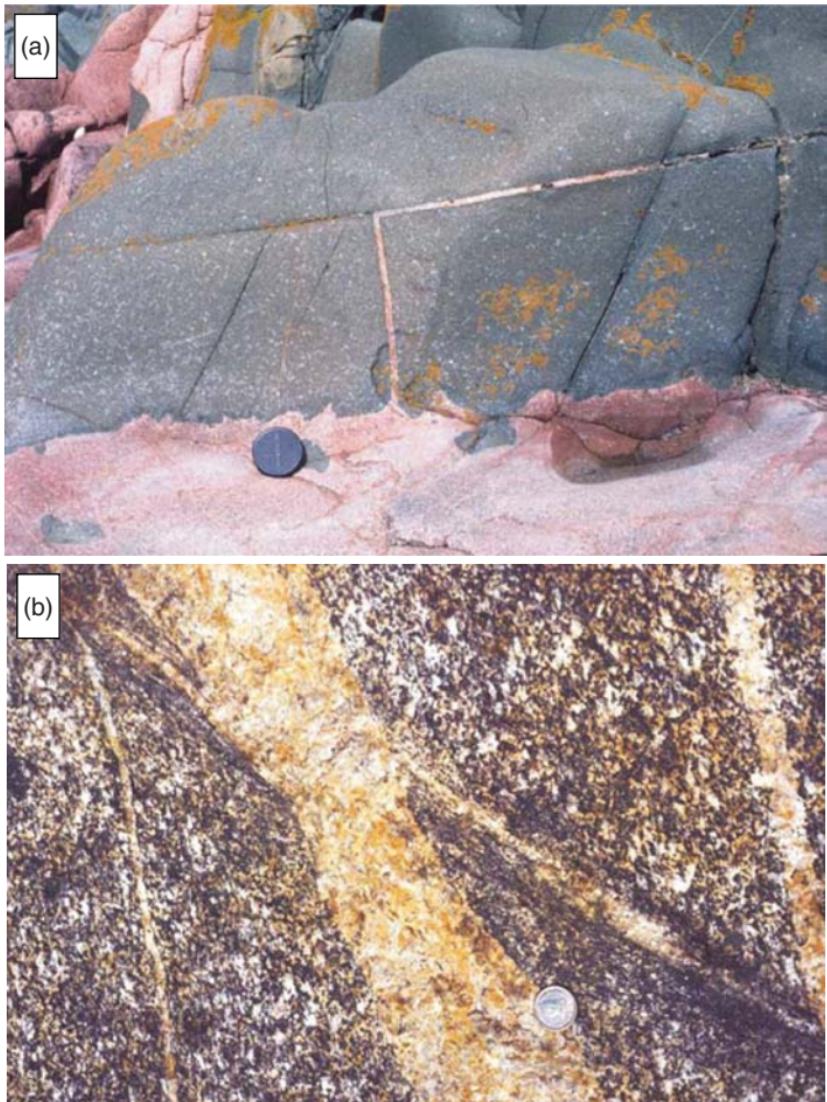


Figure 10.7 (a) Mode I (tensile) primary cooling joints in grey diorite marked out by pink granophyre intruded while both magmas were rheologically in a fluid-like state (note cuspatc contact). (b) Type II shear joint comprising quartz-pegmatite displacing an earlier sheared quartz vein in deformed granite, Peru.

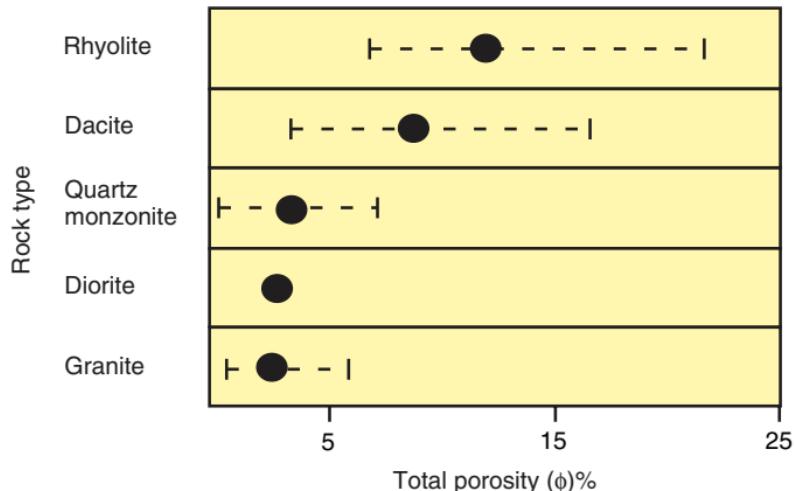


Figure 10.8 Plot showing the average total measured porosity (filled circles) and range in a selection of intermediate to acidic volcanic and plutonic igneous rocks. For the plutonic crystalline rocks, the total porosity range is low (generally <5%). The rhyolite has the highest measured porosity but this is still low in comparison to basaltic rocks (see Figure 10.9).

the development of residual soils and hydrothermal alteration, where the rock is chemically altered primarily to clay minerals. Both will cause a decrease in the strength properties of an edifice relative to its unaltered state. However, even without these degrading effects, a volcanic rock mass, and by implication the edifice itself, is an inherently weak structure when compared to the intact strength of its constituent rock (Figure 10.11).

10.4.1 The rock mass rating (RMR) system

In geotechnical engineering the rock mass rating (RMR) system forms the basis of an empirical approach that is used widely for estimating strength criteria and other rock mass properties. The geological strength index (GSI), another common rating system, is discussed later.

RMR classifies a rock mass using five main parameters (Table 10.2),

- i) the strength of the intact rock (known already from laboratory tests),
- ii) the rock quality designation (RQD),
- iii) spacing of discontinuities (collectively fractures, faults joints, etc.),
- iv) the condition of the discontinuities and
- v) the groundwater conditions.

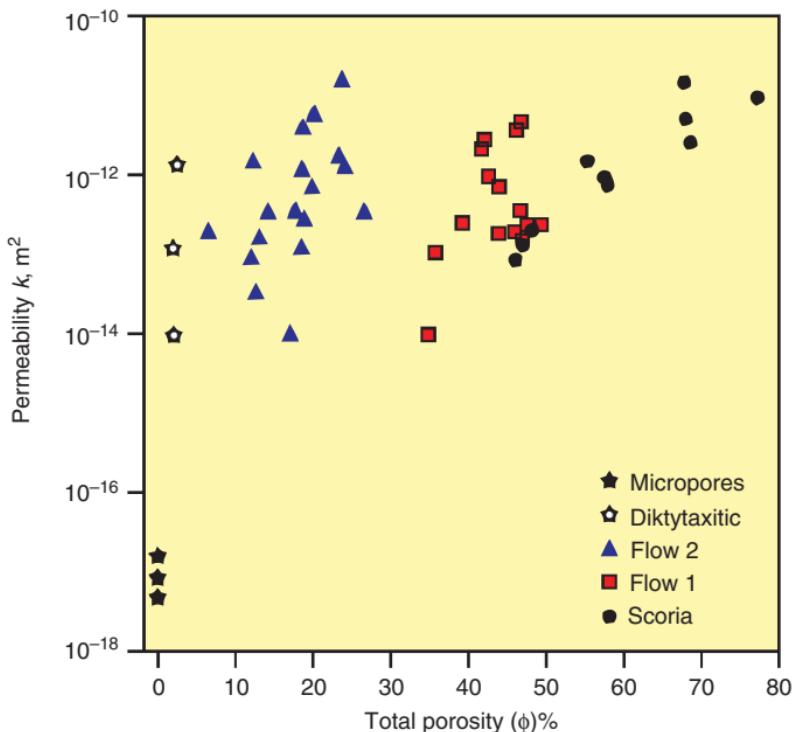


Figure 10.9 Total porosity versus permeability relationship in volcanic rocks (basaltic andesites and scoria) from the Cascades, USA. Porosity ranges from close to 0% to nearly 80% in highly vesicular scoria.

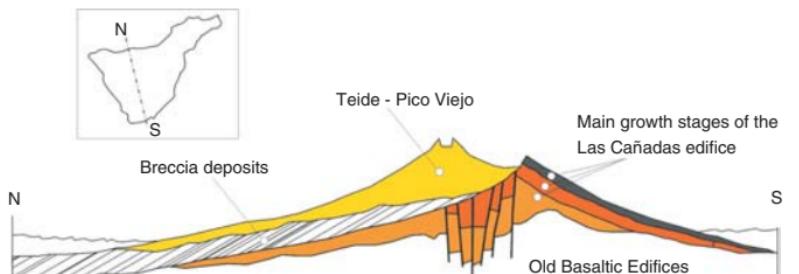


Figure 10.10 Schematic cross section through the island of Tenerife showing basement rock (old volcanic edifice) and breccia deposits to the north (interpreted as the products of successive northward directed lateral collapse), with the active Teide-Pico Viejo complex constructed on top. The structure and stability of the volcano are governed by the geotechnical properties of both the edifice and the basement rocks upon which it is built.

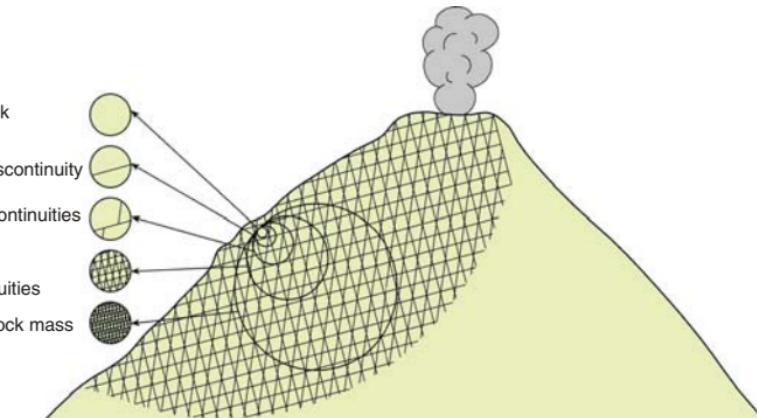


Figure 10.11 The effects of scaling of geotechnical properties on rock mass strength using a volcanic cone as an example (cf. Figure 10.10). In hand specimen the absolute strength of intact rock is dominant. But with increasing scale an ever greater volume of rock is available to move as discrete blocks within bounding discontinuities. In this way, the overall strength of the jointed rock mass is always less than that of the intact rock comprising it.

It is important to realise that descriptive, field-based rock mass classification schemes such as the RMR can be used to make predictions about the mechanical properties of the rock mass if used in conjunction with relevant empirical expressions. A particularly powerful combination involves combining the RMR with a method known as the *Hoek-Brown Criterion* devised to assess rock mass strength. To do this we need to follow a series of steps that progressively calculate several key variables through the cumulative processes outlined below. In this way it can be shown using RMR data that the rock mass strength of, for example, basalt is up to 80% weaker than its intact compressive strength as measured under laboratory conditions.

10.4.1.1 Step 1: data collection

The first step of the process involves data collection. Key data are the number and density of discontinuities, and their spacing, as measured along a linear traverse of the rock face. In addition we need to know something about the physical dimensions and properties of cracks and fractures, for example widths, whether they are closed or open, rough surfaces or weathered, and their moisture content as a measure of groundwater activity. These parameters are summarised in Table 10.2 along with a descriptive rating for each.

The first step is to work out the RQD. To do this you will need to measure the number of (i) intersections and (ii) spacings between all discontinuities along a

Table 10.2 The RMR classification parameters and their ratings (in yellow).

Strength of intact rock	Uniaxial compressive strength (MPa)	>250	100–250	50–100	25–50	5–25	1–5	<1
1	Rating	15	12	7	4	2	1	0
	RQD (%)	90–100	75–90	50–75	25–50	<25		
2	Rating	20	17	13	8	3		
	Spacing of discontinuities (m)	>2	0.6–2	0.2–0.6	0.06–0.2	<0.06		
3	Rating	20	15	10	8	5		
	Condition of discontinuities	Very rough, discontinuous, no separation, unweathered	Rough walls, separation <0.1 mm, slightly weathered	Slightly rough, separation <1 mm, highly weathered	Slippages or gouge <5 mm thick, or continuous separation	Soft gouge >5 mm thick or separation >5 mm continuous, decomposed wall rock		
4	Rating	30	25	20	10	0		
Ground water conditions		Completely dry	Damp	Wet	Dripping	Flowing		
5	Rating	15	10	7	4	0		

¹ After Bieniawski (1989).²RMR = 1 + 2 + 3 + 4 + 5 and ≤ 100 .

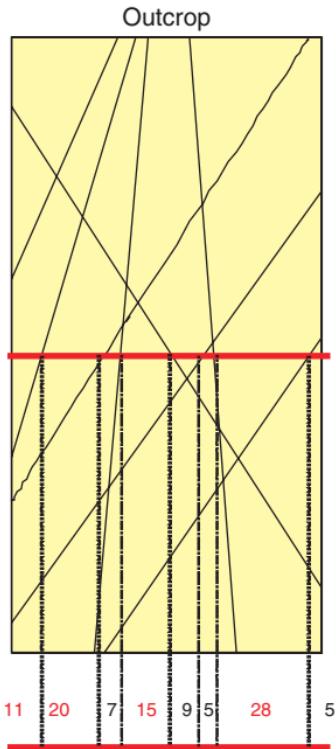


Figure 10.12 Measuring spacing of discontinuities. A tape is held in position and the number of discontinuities and the spacing between each one are recorded over a set traverse length (1.0 m). These data comprise the Rock Quality Designation (RQD), a key component of the RMR (Table 10.2).

known horizontal distance. For convenience we will assume a sample distance of 1.0 m (although in practice it can be any sensible length equal to or greater than this). Figure 10.12 shows the basic set-up using a horizontal tape measure. To begin, simply record the total number of fractures that intersect the 1.0 m traverse and the distance between them. From this, work out the RQD remembering that an RQD of 100% would mean that the rock mass is equivalent intact rock (for example, Figure 10.13). Continue working through Steps 1–5 in Table 10.2. At the end you should have five individual ratings that added together equal 100 or less. For most rock types encountered your RMR range should lie somewhere between 30 and 80.

10.4.1.2 Step 2: estimating rock mass strength using the Hoek–Brown criterion

In the 1980s, Hoek and Brown proposed a novel method for estimating rock mass strength that makes use of the RMR classification and the general empirical criterion that rock mass strength is defined by the relationship between the principal stresses at failure. Once you have a value for the RMR then the next



$$RQD = \frac{\sum \text{Length of Line Sections} > 0.1m}{\text{Total Length of Scan Line}} \times 100$$

$$RQD = \frac{0.11 + 0.2 + 0.15 + 0.28}{1.00} \times 100\%$$

$$RQD = 74\%$$

Figure 10.13 Estimating the Rock Quality Designation (RQD) from outcrop. Measure a horizontal distance across the rock face (for example, 1.0m) and sum the number of discontinuity intersections of spacing greater than 0.1m (highlighted red). Then divide by 1.0 and multiply by 100 to get %RQD.

step of the process involves calculating two Hoek Brown parameters m and s . To do this, look at Table 10.3. It is a list of equations that together tell you all you need to know about the key mechanical and geotechnical properties of a rock mass including cohesion and friction angle. The example below (Figure 10.14) shows how to calculate the numerical values of s and m in the field with the help of a calculator (or even better a laptop and spreadsheet). Values for the intact compressive strength (σ_c) and the Hoek Brown variable m_i are given in Table 10.4. The entire process is summarised in Figure 10.15.

To gain a comprehensive overview of the rock mass condition and characteristics, the RMR can be calculated both horizontally and vertically for each segment of the rock mass under investigation. The horizontal and vertical RMR

MINERALISATION AND GEOTECHNICAL PROPERTIES

Table 10.3 Key equations for calculating important rock mass geotechnical properties. RMR is a variable in both Hoek-Brown parameters m and s and can be calculated easily to obtain an estimate of the compressive and tensile strength of a rock mass.

Property	Symbol	Expression
Hoek-Brown criterion (biaxial strength)		$\sigma_1 = \sigma_3 + \sigma_c \left(m \frac{\sigma_3}{\sigma_c} + s \right)^a$
Hoek-Brown parameter m	m	$m = m_i \exp[(\text{RMR} - 100)/28]$
Hoek-Brown parameter s	s	$s = \exp[(\text{RMR} - 100)/9]$
Rock mass compressive strength	σ_{cm}	$\sigma_{cm} = (s\sigma_c^2)^{1/2}$
Rock mass tensile strength	σ_{tm}	$\sigma_{tm} = \frac{\sigma_c}{2} [m - (m^2 + 4s)^{1/2}]$
Instantaneous friction angle	ϕ_i	$\phi_i = \tan^{-1} \left[4h \cos^2 \left(30^\circ + \frac{1}{3} \sin^{-1} h^{-\frac{3}{2}} \right) - 1 \right]^{-\frac{1}{2}}$
	h	$h = 1 + \frac{16(m\sigma_n + s)}{3m^2}$
Mohr-Coulomb equation		$\tau = c + \mu\sigma_n, \mu = \tan \phi$
Instantaneous cohesion ($\sigma_n > 0$)	c_i	$c_i = \tau - \sigma_n \tan \phi$
Cohesive strength ($\sigma_n = 0$)	c'	$c' = \tau$

values can then be averaged to give one RMR value for that segment. Alternatively, just the horizontal RMR might be sufficient. You make the call. RMR values calculated using the steps outlined above and based on an averaged vertical and horizontal RMR for a range of igneous rock types are summarised in Table 10.4. Although the ages and tectonic settings of volcanic activity (Snowdonia, UK and Tenerife, Canary Islands) differ markedly, it is clear that the compressive strength (MPa) of each rock mass is considerably weaker than the intact rock itself. The implication is that a volcanic edifice will be a much weaker structure than the material comprising it, regardless of composition.

10.4.2 The geological strength index (GSI)

The GSI is a descriptive system similar to the RMR but regarded by many as providing better strength estimates where the quality of the rock mass is poor (for example, $\text{RMR} < 25$). The GSI differs from the RMR in that it relies much

1/4/2004

Fractured Rock Mass, Tenerife

GSI: $28^{\circ}09'28.72\text{ N}$
 $16^{\circ}38'13.86\text{ W}$

Rock type: phonolite lava, orthocryic, vesicular top.
 Slightly weathered, thickness 2-3 m.

$$\frac{RQD}{RMR} = 72\%$$

$$RMR = 50$$

Hoek Brown Parameters

$$\textcircled{1} \quad M = 25 \exp\left(\frac{50-100}{28}\right) = \underline{4.192}$$

$$\textcircled{2} \quad S = \exp\left(\frac{50-100}{9}\right) = \underline{3.886 \times 10^{-3}}$$

$$\textcircled{3} \quad \sigma_{cm} = \left[3.886 \times 10^{-3} (250)^2 \right]^{\frac{1}{2}} \\ = \underline{15.544 \text{ MPa}}$$

$$\textcircled{4} \quad \sigma_{tm} = \frac{250}{2} \left[4.192 - (4.192^2 + 4 \cdot 3.886 \times 10^{-3})^{\frac{1}{2}} \right] \\ = \underline{-0.231 \text{ MPa}}$$

NB $\sigma_c = 250 \text{ MPa}$, $M_i = 25$

Figure 10.14 Specimen calculation showing workings for Hoek Brown parameters, rock mass compressive and tensile strength.

more on descriptive observations. For better quality rock masses (for example, RMR > 25), the GSI can be estimated directly from the RMR. An inherent source of uncertainty with the RMR system is that up to 15% of the rating can be attributed to ground water conditions, which can change on the scale of hours or less from dry to wet due simply to localised rainfall. The GSI system does

Table 10.4 A summary of the rock mass properties obtained from volcanic rocks in the Snowdonia volcanic group^a and from the island of Tenerife^b. Values of intact laboratory compressive strength and the Hoek Brown variable m_i are also included for reference.

Rock type	Rhyolite ¹	Tuff ¹	Micro granite ¹	Basalt ²	Phonolite ²
RMR	52–74	47–74	57–79	48–60	38–60
Laboratory compressive strength (MPa) ³	175	150	250	150	175
Hoek Brown variable m_i	25 ± 5	13 ± 5	32 ± 3	25 ± 5	25 ± 5
Rock mass compressive strength (MPa)	12–41	7–35	23–78	8–16	6–19
Rock mass tensile strength (MPa)	(−0.29) to (−1.5)	(−0.18) to (−1.4)	(−0.3) to (−1.55)	(−0.2) to (−0.5)	(−0.1) to (−0.5)

³Values taken from Brown (1981).

not require this (or other) variables to be assessed and while relatively untested may ultimately prove more the reliable rating system in a volcanic context.

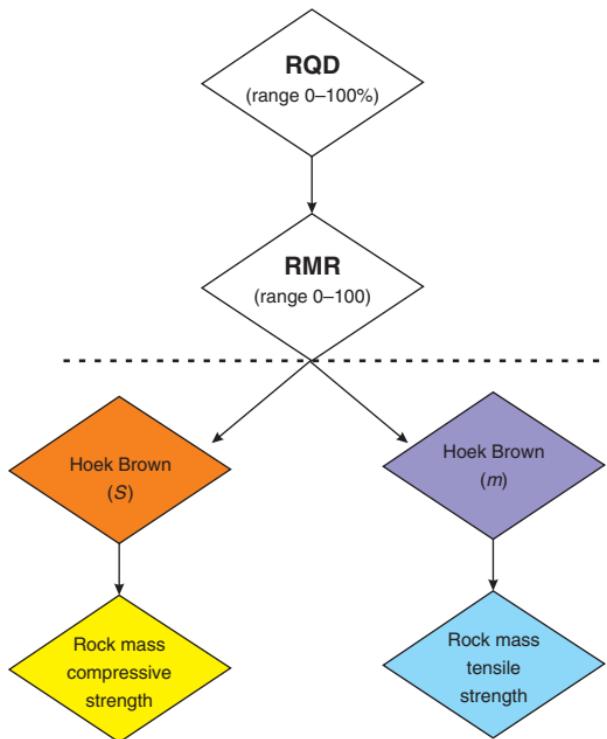
10.5 Summary

Igneous rocks are often hydrothermally altered and are on occasion host to economically significant mineral deposits including copper, tin, tungsten, platinum and clays. They can also act as reservoirs for hydrocarbons where fracturing has resulted in elevated permeability and porosity. In the field alteration zones around hydrothermal fractures and veins should be recorded along with structural data (dip and strike), spacing, relationship to larger structures (for example, faults) vein width and number density. Any associated metaliferous minerals should be identified (Table 10.1) and their occurrence noted. Where possible the mode of fracture (tensile or shear) should be discerned along with any cross-cutting relationships so that a detailed picture can be constructed to help shed light on the emplacement and post-emplacement deformation history.

Intact rock, measured in a laboratory, is a relatively homogeneous and continuous material. However, on larger scales in the field, the notion of continuity does not hold. Instead the structure of a rock mass on the mesoscale is highly discontinuous and heterogeneous and it is the presence of discontinuities that

Step 1: field data

Refer to table 10.2

**Step 2: Calculate**

Refer to table 10.3

Figure 10.15 Flow diagram summarising the stages involved in calculating the compressive and tensile strength of a rock mass from field data. Colour relates to equation in Table 10.3.

governs the geotechnical properties of a rock mass in response to deformation and its hydraulic properties. The RQD, RMR and ultimately the compressive and tensile strength of the rock mass can be assessed using the simple steps outlined here. If the stresses acting on igneous rock mass are altered in some way, either through man-made activities such as drilling or construction, or through magmatic activity (inflation of magma chambers inside a volcano or intrusion of sills and dykes), then the rock mass can lose integrity and fail. If the rock mass happens to be an active volcano, then the consequences may be severe indeed.

APPENDIX

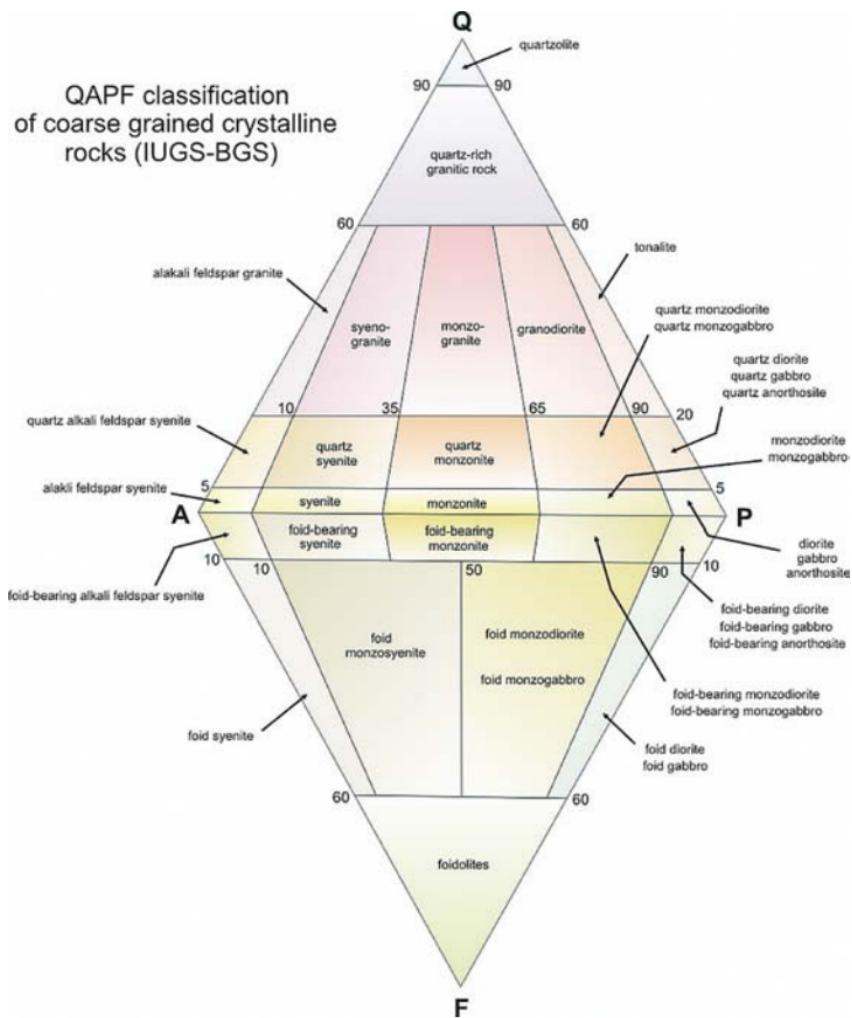


Figure A.1 QAPF classification of coarse grained crystalline rocks (IUGS-BGS).

FURTHER READING

- Bieniawski, Z.T. 1989. *Engineering Rock Mass Classifications*. Wiley, New York.
- Bouchez, J.L., Hutton, D.W.H. and Stephens, W.E. (eds) (1997) *Granite: From Segregation of Melt to Emplacement Fabrics*, Kluwer Academic Publishing, Dordrecht, The Netherlands.
- Branney, M.J. and Kokelaar, P.B. (2002) *Pyroclastic Density Currents and the Sedimentation of Ignimbrites*, Geological Society Memoir, Geological Society Publishing House, Date published: 2002-12-01. ISBN: 13: 9781862391246, ISBN: 1862391246.
- Brown, E.T. (ed.) 1981. *Rock Characterisation, Testing and Monitoring – ISRM Suggested Methods*. Pergamon, Oxford.
- Didier, J. and Barbarin, B. (eds) (1991) *Enclaves and Granite Petrology*. Elsevier Science Ltd, 626 pp. ISBN: 978-0444891457.
- Gill, R. (2010) *Igneous Rocks and Processes – A Practical Guide*, Wiley-Blackwell.
- Jerram, D. (2011) *Introducing Volcanology*, Dunedin Academic Press.
- Lockwood, J.P. & Hazlett, R.W. (2010) *Volcanoes: Global Perspectives*. Wiley-Blackwell ISBN: 978-1-4051-6250-0
- Petford, N., Sparks, R.S.J. and Hutton, D.H.W. (2006) *Plutons and Batholiths*. Earth and Environmental Science Transactions of the Royal Society of Edinburgh, 97, 4, v–vi.
- Pitcher, W.S. (1993) *The Nature and Origin of Granite*, Chapman & Hall, 408 pp. ISBN: 0412758601.
- Sigurdsson, H. et al. (1999) *Encyclopedia of Volcanoes*, Academic Press, October 1999. ISBN: 13: 978-0-12-643140-7, ISBN: 10: 0-12-643140-X.
- Tappe, Sebastian, Foley, Stephen F., Jenner, George A. and Kjarsgaard, Bruce A. Integrating ultramafic lamprophyres into the IUGS classification of igneous rocks: rationale and implications, *Journal of Petrology*, Sep 2005; 46: 1893–1900.
- Thomas, M.E., Petford, N. and Bromhead, E. (2004) Volcanic rock-mass properties from Snowdonia and Tenerife: implications for volcano edifice strength. *Journal of the Geological Society*, 161, 939–946.

INDEX

- A-type granite 138–139
aa lava 73–75
accessory mineral 49, 54–55
accidental lithic fragment 96
accretionary lapilli 97, 103
acid (chemical composition)
 38, 43
air-fall pyroclastic deposits
 6, 94
alkali feldspar (inc. Orthoclase)
 39–40, 45, 52, 72, 58–63,
 138–139, 147, 156, 164,
 209, 229
alkali granite 60–63, 139,
 158–161, 229
amphibole (inc. Hornblende) 40,
 50, 53, 59, 66, 72, 139
amphibolites facies 161, 183
amygdales/amygdaloidal texture
 76, 78–79
andalusite/chiaxtolite 166,
 168
andesite 3, 13, 24, 59, 66,
 72, 80
anhedral 46–47
anorthosite 64, 173, 210,
 213–214
apatite 30, 54, 58
aphanitic 44
aphyric 44, 71
aplite 30, 140, 154–159
arsenopyrite 211
ash 94–101, 105–109
ash-ring 109
back veining 129
basalt 3, 13, 24, 49, 59, 66, 69,
 71–73, 75–80, 89–90,
 120–121, 130, 137,
 177–182, 190–192, 197,
 200, 220, 226
basaltic andesite 71–72, 80,
 198, 219
basic (chemical compositions)
 42–43, 63–64
batholiths 3, 9, 136 138–144,
 150, 152
biotite 27, 39–40, 45, 49, 54,
 59, 63, 65–66, 72, 101,
 138–139, 144, 147–148,
 152–153, 156, 162, 164,
 168, 194, 197–199, 202,
 209
Block and ash flows 101–108
blocky lava surface 74–75,
 80, 90
bomb (volcanic) 98, 100, 103,
 108, 116
bridge structures (sills/dykes)
 124–125

- calc-alkaline rocks 64, 137–138, 152, 158, 160, 164–165, 207, 209
calcite 29–30, 51, 55, 58, 65, 79, 168, 207
carbonatite 43, 65
cassiterite 212
chadocryst 46–47
chalcopyrite 183, 207–211
chlorite 57–58, 168, 209
chromite/chromitite 178, 183, 186, 206, 209–210, 214
cleavage (minerals) 49–57, 65, 67
clinopyroxene 40, 63, 64, 65, 171–172
coarse grained 44
colonnade 76–77
colour (minerals) 49, 52–57, 211–212
colour index 38–42, 58–59
colour (rocks) 38–42
columnar joints 76–77, 90
composite intrusion 125–128
composite volcano 5–6, 210
compound lava flow 75
cone sheet 123
contacts/contact relationships 7–10, 18–33, 119–129, 141, 150–157, 192–201
contact metamorphism 164–165, 168
cordierite 139, 162, 168
crystallite 73
cumulate 173–176
cupola 10
- dacite 3, 13, 66, 72, 95, 101
devitrification 71–73
diabase 65, 121
diatreme 118, 131–132
diorite 3, 13, 59–64, 66, 139
dolerite 3, 59, 65, 121, 129, 130, 181
domes 82–85
druses/drusy cavities 34, 133–134
dunite 63–64, 172, 177–180
dyke (dike – US) 7–9, 119–129, 142, 159, 177–183, 196–199
- Enclave 193
entablature 76–77
epidote 56, 207–209
essential mineral 49
euhedral 46–47
- feldspathoid 39–40, 60, 63–65
felsic/-mineral, -rock 26, 30, 38–44, 49, 52, 59–61, 66, 71, 80, 82, 96, 123, 128, 129–130, 152, 177 192, 202.
femic/-mineral/-rock 40, 43
Fe-Ti oxides 49, 59, 72, 129, 130, 131, 172
fiamme 99, 104, 106, 110
fine-grained 38, 44
flame structure 27, 175
flow banding/-structures 24–25, 47, 71, 82, 157–158
Fluidisation/fumarole pipes 99, 104
foid diorite 60, 229
foid gabbro 60, 229
foid/-minerals 60, 130, 229
foid syenite 60, 64, 229
fractional crystallisation 26, 128, 129, 143, 185
- gabbro 3, 4, 13, 26, 31, 39, 43, 59–66, 124, 126, 137, 138,

INDEX

- 160, 171, 173, 177–183,
190, 213–214
- galena 51, 211
- garnet 50, 58, 138, 139, 153,
156, 162, 168, 176
- geological strength index (GSI)
224
- geothermal energy 214
- glass/glassy rocks 24–25, 33,
41, 43–44, 71–73, 84, 86,
89, 93, 96, 99, 101, 110,
112–113, 121, 135
- glassy (luster) 50, 52–57
- glass shard 93, 96, 101
- grain size (crystalline rocks)
43–44
- grain size (pyroclastic rocks)
97–98
- granite 3–4, 9, 13, 25, 31–32,
39, 41–42, 45, 59–66,
121, 133, 136–168, 190,
191, 195, 197, 198, 201,
207, 218
- granodiorite 3, 13, 59–61, 64,
66, 137–139, 142, 143,
144, 150, 154, 160, 162,
197, 198
- granophyres 103, 162
- granular texture 46
- graphic granite 162
- greenschist facies 183
- greisens 209
- habit/-crystal form (minerals)
43–44, 49–50, 52–57,
211–212
- hardness (minerals) 29, 49–58,
211–212
- harzburgite 63, 178–180, 210
- high level intrusions 7–8,
133–134
- hornfels 165–166, 168
- Hoek-Brown Criterion
222–224, 227
- hyaloclastite 75, 82, 87–89,
109–111
- hydrocarbon potential 215
- hypautomorphic 46
- hypidiomorphic 46
- I-type granite 138–139
- idiomorphic 46
- igneous layering/-lamination/
-mineral layering 23–27,
146, 152, 170, 173–177,
179–181, 209–210, 213,
214, 215
- ignimbrite 96, 97, 101–108, 116
- ilmanite 55, 139, 172
- Intact rock strength 218,
220–226
- intermediate (chemical
composition) 43, 59
- joint(s)/jointing 28–30, 33,
75–81, 83, 90, 122, 129,
155–158, 207, 216–218,
220
- juvenile fragment 93, 95, 104,
110, 113–114
- kaolinization 208–209
- kimberlite 65, 131–132
- komatiite 183–184
- laccolith 7–8, 118–119,
123–124, 133, 135
- lahar 94, 113–114
- lamprophyre 65, 121
- lapilli 97–98, 100, 103, 107,
109, 116
- lava delta 87–88
- lava flow 4–6, 68–91, 110,
114

INDEX

- lava tube 79–80
leucite 39, 55, 58, 60, 66
leucocratic (leuco-) 41–43, 49, 64, 66, 139, 153, 162, 164, 185
lherzolite 63, 176
lithic fragment 93–94, 96, 97, 99, 101, 106–107, 110, 114
Lithophysae 99
lopolith 7–8, 123, 173, 179
lustre (mineral) 49–50, 52–57, 211–212
- Mafic/-mineral, -rock 9–11, 38–43, 44, 45, 53–54, 58, 59, 61, 62, 63–66, 170–186, 209, 189–205
magnetite 55, 172, 213
Magma Mingling 192–201
Magma Mixing 190–192
Magma rheology 186–190
major mineral 39, 49
medium-grained 44
megacryst 45, 144, 145, 156, 164, 194
melanocratic (mela-) 41–43, 64, 65, 66
mesocratic (meso-) 41–43, 65
metalliferous mineralization 207, 209, 211–212
metamorphism 164–165
miarolitic (cavities) 34, 134
minette 65
minor intrusions 7–9, 29, 32, 118–135, 138, 162, 177
modal minerals/mineralogy 26, 146, 174, 176, 177
molybdenite 211
monazite 55, 218, 229
monchiquite 65
mudflow 113–114
- multiple intrusion 121, 123, 126, 128, 129, 165, 166, 177, 179
muscovite 39, 40, 53, 59, 138–139, 152, 153, 156
- nepheline 39, 40, 52, 60, 64, 66, nepheline syenite 64, 66
net-veining 129
norite 171, 179, 184, 210, 213, 214
normative (norm) minerals/mineralogy 39–40
- obsidian 25, 71, 82, 121
oikocryst 46–47
olivine 39–40, 49, 50, 53, 59, 63–66, 72, 131, 171–184, 214
ophiolite complex/-association 3, 177–183
orbicular granite 162
orthopyroxene 40, 63, 64, 65, 171–172
- pahoehoe lava surface 68, 73–75, 79–80, 86, 89, 90
palagonite 84, 89,
parasitic volcano/cone 6, 108
pegmatite 29–32, 45, 130, 156–157, 162, 177
peridotite 3, 39, 59, 62, 64, 131, 171–185
peridotite xenolith 27, 131
Permeability 216
perthite 52
phaneritic 44
phenocrysts 44–45, 65–66, 71–72, 93
phonolite 66
pillow lava 82, 84, 86–90, 178–183

- pitchstone 71, 73, 121
plagioclase feldspar (inc. albite, anorthite) 39, 43, 45, 52, 59–66, 72, 121, 139, 164, 171, 713–174, 179, 180, 194, 209, 229
plagiogranite 178–179
plateau (lava) 4–5
poikilitic 46–47
porphyritic (phyric) 44–45, 48, 65–66
porphyry mineral deposit 207–208, 210
post-tectonic intrusion/injection 158, 160
potassium silicate alteration 209–210
pre-full crystallisation fabric 148–150
pre-tectonic intrusion 158, 160
propylitic alteration 208–210
pumice 93–101, 103, 106–107, 112
pyrite 57, 207, 209, 210, 211
pyroclastic cone 6, 108
pyroclastic flow/deposit (surge) 5–7, 82, 94, 95, 97, 95, 101–108
pyroxene (inc, clinopyroxene, orthopyroxene) 39–40, 49, 50, 53, 59, 63–66, 72, 121, 129, 130, 131, 143, 168, 171–172
pyroxenite (ortho-, clino-) 63, 64, 172, 176, 179, 210, 214
Porosity 216
QAPF diagram 60, 229
quartz 29–32, 39–40, 45, 49–52, 58, 59–66, 71–73, 130, 137–138, 162, 194, 207–209, 229
radial dyke 122
ramp structure 81
rapakivi texture/-granite 139, 164
rhyolite 3, 13, 24–25, 59, 66, 71, 72, 80–82, 120, 121, 128, 137, 190–191, 218, 226
riebeckite 53, 64, 139
ring-dyke 122–123
ring intrusion/complex 9, 138, 140
rock mass rating system (RMR) 218–226, 227
rock quality designation (RQD) 220–223, 227
roof pendant 9–10
S-type granite 138–139
salic (minerals/rocks) 40, 43, 60–62
scheelite 212
schist/schistosity 158, 165
schlieren/-structure 146, 147, 152, 156, 192
scona cone 108, 111, 116
scoria 93, 96, 108
secondary (mineral) 55–58
serpentine/serpentization 172, 177
sheeted (-dyke) complex 177–183
shield volcano 6, 75
silicification 209, 210
sill 7–9, 71, 76, 118–122, 124, 125–127, 129, 130, 135, 142, 165, 196
sillar 99
slickensides 29
slump(ing) structures/deposits 27, 174–175, 186
soviet 65

INDEX

- sphalerite 212
spheine 54, 58, 139
spherulite 73, 90
spinel 55, 131
strain markers 194
stock 9, 210
stockwork 183, 185, 207, 209
streak (metalliferous minerals)
 211–212
subhedral 46–47
syenite 60, 62–64, 66, 139, 229
syn-tectonic intrusion/injection
 158, 160, 161
- tachylite 71, 84
tectonic setting (igneous rocks)
 2–4, 12–13
tonalite 60–61, 138–139, 229
tourmaline 30, 54
trachyte 66, 71
tremolite 53
trondhjemite 138–139,
 160–161
tuffcone/-ring 109, 131
tuffisite 131
tuff pipe 131
twinning (minerals) 50, 52
Tuyas 110
- Ultramafic (chemical
 composition) 43
- ultramafic rock 187–188,
 209
umber 181
- vein 29–33, 55, 56, 57, 58, 149,
 152, 155, 156, 159, 193,
 201, 207–209, 210, 211,
 216, 217
vesicles/vesicular texture
 76–80, 101, 122, 128,
 129
volcanic agglomerate/-breccia
 131–132, 140, 210
volcano instability 215, 219
- websterite/olivine websterite 63,
 176
wehrlite 63, 178–179
welding (ignimbrite) 104,
 106–107, 109
wolfram 212
wollastonite 168
- xenolith (-inclusions) 10, 27,
 38, 127, 201–202
xenocryst 127, 202
- zeolite/-facies 56, 58, 65,
 78–79, 130, 134, 183.
- Zoned pluton 142–145