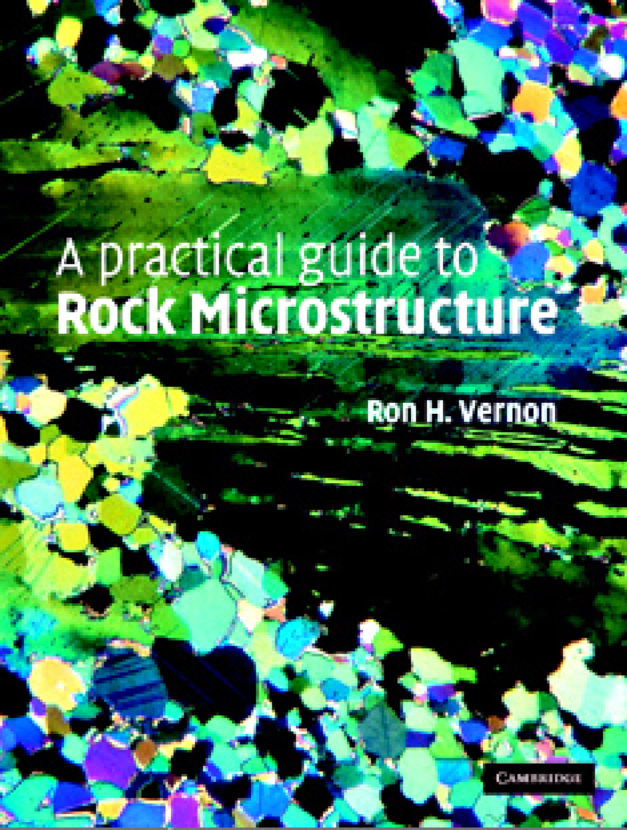

A practical guide to
Rock Microstructure

Ron H. Vernon

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A Practical Guide to Rock Microstructure

Rock microstructures provide clues for the interpretation of rock history. A good understanding of the physical or structural relationships of minerals and rocks is essential for making the most of more detailed chemical and isotopic analyses of minerals.

Ron Vernon discusses the basic processes responsible for the wide variety of microstructures in igneous, sedimentary, metamorphic and deformed rocks, using high-quality colour illustrations. He discusses potential complications of interpretation, emphasizing pitfalls, and focussing on the latest techniques and approaches. Opaque minerals (sulphides and oxides) are referred to where appropriate. The comprehensive list of relevant references will be useful for advanced students wishing to delve more deeply into problems of rock microstructure.

Senior undergraduate and graduate students of mineralogy, petrology and structural geology will find this book essential reading, and it will also be of interest to students of materials science.

RON VERNON is Emeritus Professor of Geology at Macquarie University, Conjoint Professor of Geology at the University of Newcastle and Research Professor at the University of Southern California. He has taught undergraduate geology courses in Australia and Italy, as well as graduate courses and workshops in the USA, Italy, Germany, Finland and Mexico.

He has written two books, *Metamorphic Processes* (1976) and *Beneath Our Feet* (2000). The latter provides a clear and enthusiastic introduction to rocks for the non-geologist.

A Practical Guide to Rock Microstructure

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For Katie

Contents

Preface

Learning about rocks can give much pleasure to anyone interested in Earth and its development. I hope that readers of this book will share my enthusiasm for examining rocks with the microscope. I planned the book to be an introductory review of the main processes responsible for the microstructures of Earth rocks. However, I soon realized that if I did that, the book would be a collection of half-truths, with little scientific value. Though many rock microstructures are understood fairly well, the interpretation of many others involves considerable controversy, and new ideas are being published all the time. So, I have felt compelled to mention problems of interpretation and to present alternative views, where appropriate. Thus, the book has evolved into (1) a basic explanation of the main processes, (2) an introduction to more complex issues of interpretation and especially to the relevant literature, and (3) an outline of modern approaches and techniques, in order to emphasize the ongoing, dynamic nature of the study of rock microstructure. Because complicated problems cannot be discussed in detail in a book of this kind, I have tried to provide a sufficient number of references to enable the reader to delve more deeply.

I assume that the reader has a basic knowledge of geology, rock types and microscopic mineral identification. Thus, the book is aimed mainly at senior geoscience undergraduates and above. Emphasis is placed on higher-temperature processes, i.e. those that occur under igneous and metamorphic conditions, although the book begins with a brief discussion of sedimentary microstructures, as background for some of the metamorphic microstructures. The mineral abbreviations used follow those suggested by Kretz (1983), as extended by Bucher & Frey (1994), and are listed at the start of the book. There is an extensive glossary of microstructural terms at the end of the book.

I also hope that materials scientists may also gain some benefit and interest from the microstructures discussed and illustrated, because rocks are the 'materials' of Planet Earth, in the sense of 'materials science': the branch of science that links all solid materials, such as metals, ceramics, glass, organic polymers and, of course, rocks.

I took all the photographs, except where otherwise acknowledged. I am also responsible for most of the line drawings, with the assistance of Dean Oliver

(Figs. 5.11, 5.37) and Daleth Foster (Fig. 5.93). I thank David Durney, Dick Flood, Scott Johnson and Scott Paterson for critically reading parts of the typescript, Judy Davis for assistance with computer techniques, Geoff Clarke for access to specimens at the University of Sydney, Ross Both, John Fitz Gerald and Neil Mancktelow for providing images, and John Lusk, John Ridley, David Durney and Pat Conaghan for providing specimens of opaque minerals, deformed rocks/veins and sedimentary rocks at Macquarie University. People who kindly provided other samples or thin sections are acknowledged in the figure captions.

Mineral symbols used in this book

After Kretz (1983), extended by Bucher & Frey (1994).

Ab	albite
Act	actinolite
Alm	almandine
Als	aluminosilicate
Am	amphibole
An	anorthite
And	andalusite
Apy	arsenopyrite
Bt	biotite
Cal	calcite
Ccp	chalcopyrite
Chl	chlorite
Chr	chromite
Cld	chloritoid
Cpx	clinopyroxene
Crd	cordierite
Crn	corundum
Cv	covellite
Czo	clinozoisite
Dol	dolomite
Ep	epidote
Gln	glaucophane
Gn	galena
Gr	graphite
Grs	grossular
Grt	garnet
Hbl	hornblende
Ilm	ilmenite
Kfs	K-feldspar
Ky	kyanite

Lws	lawsonite
Mag	magnetite
Ms	muscovite
Ne	nepheline
Ol	olivine
Omp	omphacite
Opx	orthopyroxene
Or	orthoclase
Pgt	pigeonite
Phl	phlogopite
Pl	plagioclase
Prh	prehnite
Py	pyrite
Qtz	quartz
Rt	rutile
Scp	scapolite
Ser	sericite
Sil	sillimanite
Sp	sphalerite
Spl	spinel
Spr	sapphirine
Sps	spessartine
Srp	serpentine
St	staurolite
Stp	stilpnomelane
Tlc	talc
Toz	topaz
Tr	tremolite
Ttn	titanite (sphene)
Tur	tourmaline
Wo	wollastonite
Zo	zoisite
Zrn	zircon

Chapter 1

Background

1.1 Introduction

I wrote this book to help you to interpret what you see when you look at thin and polished sections of rocks with the microscope. I say 'help', rather than 'teach', because I don't want to give the impression that every microstructure you see can be easily and unambiguously interpreted in terms of processes that produced the rock. Many can, but in many other instances, conventional interpretations are ambiguous or poorly understood. So I intend the book to be only a guide, and I present alternative ideas where appropriate. A healthy scepticism should be maintained when interpreting rock microstructures yourself and also when reading the interpretations of others.

1.2 History of the examination of rocks with the microscope

Rocks in natural outcrops, in samples knocked off these outcrops and in drill cores, are beautiful and instructive. We can see different minerals, and identify many of them with the aid of a hand lens. We can also see some of the more obvious structures in the rocks. However, cutting a slice (section) though a rock with a diamond-impregnated circular saw and polishing the sawn surface shows us the various minerals alongside each other, rather than piled confusingly all around each other. This reveals the structure even more clearly, as can be seen in the polished facing slabs on many buildings and bench tops.

But we always want to see more. So, when D. Brewster in 1817 and William Nicol about 1830 showed how to make a slice of crystalline material thin enough to transmit light (0.03 mm is the standard thickness), stuck to a glass microscope slide (Shand, 1950, p. 6; Loewinson-Lessing, 1954), it wasn't surprising that a curious person, such as Henry Sorby, should start looking at these *thin sections* of rocks (Sorby, 1851, 1853, 1856, 1858, 1870, 1877, 1879, 1880, 1908). Sorby learnt the technique of making thin sections from W. C. Williamson in 1848 (Judd, 1908; Folk, 1965) and made the first rock thin section in 1849 (Judd, 1908). He was the first to look seriously at rock sections with the microscope, beginning with a study of chert, a siliceous sedimentary rock that was a very appropriate choice for microscopic investigation, in view of its very fine grain-size. He described and suggested a mechanical origin for slaty cleavage (Sorby, 1853, 1856), noticed many of the basic features of igneous and metamorphic rocks, made many important observations on sedimentary rocks, including carbonate rocks (Sorby, 1851, 1879), investigated pressure-solution (using fossil crinoids), described meteorites, and published the first papers on the examination of polished sections of metals with the microscope (Sorby, 1864, 1887). So not only is he the founder of petrography (the description of rocks), but the founder of metallography as well (Smith, 1960). He also investigated fluid inclusions in minerals, heating crystals to watch the gas bubbles disappear, in order to get an estimate of the temperature of crystallization of the mineral (Sorby, 1858; Folk, 1965).

Sorby was followed soon after by many others, as discussed by Johannsen (1939) and Loewinson-Lessing (1954). Among them were Dawson (1859), Zirkel (1863, 1866, 1876), Vogelsang (1867), Fouqué& Michel-Lévy (1879), Allport (1874), Rosenbusch (1873, 1877) and Teall (1885, 1886). Since those days, the light microscope has become the main tool for identifying minerals and examining their microstructures, although it has been augmented by many modern techniques (Section 1.6).

1.3 How relevant is the microscope today?

Many petrologists concentrate on the mineralogical and chemical aspects of rocks, without spending much time looking at rocks with the microscope. In fact, in these days of marvellous techniques for the chemical and isotopic analysis of minerals, some people may feel that simply looking at and measuring the shapes and arrangements of crystals in rocks with the microscope is a little out of date. However, carrying out detailed chemical and isotopic analyses of minerals when you don't understand the relationships of these minerals to other minerals in the rock makes little sense. It's a waste of expensive resources, at very least.

On the other hand, many structural geologists look at the physical or structural aspects of minerals and rocks, especially from the viewpoint of deformation processes and preferred orientations of grains, without being concerned about the chemical aspects of these processes. Both approaches are valuable, of course, but their interrelationships can be particularly illuminating. Fortunately, many researchers are attempting to integrate the chemical and physical approaches, and the study of rocks with the microscope provides a link between them. In fact, the detailed study of processes in rocks at the microscopic scale is now a major area of research, especially among younger people, in many universities and other research institutions. Moreover, new observational techniques are being developed and used, as discussed in Section 1.6.

Research microscopes commonly have both transmitted and reflected light facilities. An excellent example of the simultaneous use of transmitted and reflected light microscopy is the study of Columbia River basalts by Long & Wood (1986), in which reflected and transmitted light photographs are arranged side by side, clearly revealing the dendritic shapes of the opaque Fe–Ti oxide minerals and their relationships to the transparent and translucent silicate minerals. Some leading books and review articles on minerals in reflected light, with emphasis on microstructures, are those of Bastin (1950), Edwards (1947, 1952), Cameron (1961), Ramdohr (1969), Stanton (1972), Craig & Vaughan (1994) and Craig (1990a,b).

1.4 Mineral identification

Learning to identify minerals takes time and practice, and is outside the scope of this book. Close teaching in a laboratory situation is the best way to learn about the optical properties of minerals, using textbooks specifically written for the purpose (e.g. Fleischer *et al*., 1984; Shelley, 1985a; Nesse, 1991; Gribble & Hall, 1992; Deer *et al*., 1992). Ideally, this should go hand-in-hand with learning about microstructures.

1.5 The concept of a section

Thin and polished sections are two-dimensional sections through three-dimensional objects, and this must always be kept in mind, as explained in some detail by Hibbard (1995). Mineral grains can have unexpectedly complex three-dimensional shapes (see, for example, Rigsby, 1968; Byron *et al*., 1994, 1995, 1996). Two or even three orthogonal sections may be necessary to reveal the structure of structurally anisotropic rocks, and some recent detailed microstructural studies have used: (1) *serial sectioning* (see, for example, Byron *et al.,* 1994, 1995, 1996; Johnson & Moore, 1996), coupled with *image analysis* by computer, to construct a three-dimensional image of the microstructure, and (2) *computed X-ray tomography* (Section 1.6), to reveal the three-dimensional distribution of large crystals (porphyroblasts) in metamorphic rocks (Carlson & Denison, 1992; Denison & Carlson, 1997), plagioclase chains in basalts (Philpotts *et al*., 1999), and former melted rock (leucosome) in migmatites (Brown *et al*., 2002).

1.6 Newer techniques

This book deals mainly with microstructures visible in the optical (light) microscope, in standard thin or polished sections, with the use of polarized light. However, some newer techniques are very useful for revealing features not apparent or less clearly shown in polarized light, as outlined below. Several examples of photographs taken with these techniques will be presented in the book. The new techniques underline the fact that the study of rock microstructure is a dynamic, progressive field of research.

(1) **Cathodoluminescence** (CL) is a technique that can reveal internal microstructures of grains of some minerals, for example, compositional zoning, overgrowths, microcracking and replacement veining in quartz, calcite, dolomite, magnesite, zircon, plagioclase, K-feldspar, diamond, fluorite, sphalerite, kyanite, pyrope garnet, corundum, cassiterite, anhydrite and apatite (Sippel & Glover, 1965; Smith & Stenstrom, 1965; Zinkerngel, 1978; Matter & Ramseyer, 1985; Owen & Carozzi, 1986; Sprunt, 1978, 1981; Sprunt & Nur, 1979; Field, 1979; Reeder & Prosky, 1986; Padovani *et al*., 1982; Hanchar & Miller, 1993; Marshall, 1988; Ramseyer *et al*., 1988; Morrison & Valley, 1988; Mora & Valley, 1991; Yardley & Lloyd, 1989; Hopson & Ramseyer, 1990; Barker & Kopp, 1991; Shimamoto *et al*., 1991; Williams *et al*., 1996; Mora & Ramseyer, 1992; D'Lemos *et al*., 1997; Watt *et al.,* 1997, 2002; Hayward, 1998; Müller *et al*., 2000; Pagel *et al*., 2000; Janousek *et al*., 2000; Ahn & Cho, 2000; Götze, 2000; Rubatto & Gebauer, 2000; Penniston-Dorland, 2001; Rubatto *et al*., 2001; Hermann *et al*., 2001; Barbarand & Pagel, 2001; Peppard *et al*., 2001; Rusk & Reed, 2002; Viljoen, 2002; Rougvie & Sorensen, 2002). CL is especially useful for revealing microstructural details in minerals that are colourless in the light microscope, for example calcite, quartz and feldspar. For example, quartz, untwinned K-feldspar and untwinned plagioclase may be distinguished by their CL colours in fine-grained aggregates, and CL can assist in provenance and diagenetic studies in sedimentary rocks. The technique is not suitable for iron-rich minerals. Some applications of CL are discussed in Sections 3.11.7, 3.11.9 and 5.9.3.

It can be used with the light microscope (producing true CL colours with a spatial resolution of 1–2 μm), or the scanning electron microscope (producing grey-scale variations with a spatial resolution of less than 1 μm), and may be combined with CL spectroscopy to enable spectral spot analysis of revealed features. CL is combined with X-ray tomography to reveal the internal structure of diamonds (see, for example, Field, 1979).

CL is caused by defect structures in the crystal lattice, such as impurity atoms (transition metals, rare earth elements, lead, titanium, actinides), vacancies and dislocations produced during formation and/or deformation of the mineral, which therefore reflect conditions of crystallization, deformation and alteration. The technique involves coating a polished thin section with carbon and bombarding it with electrons in a vacuum. This bombardment produces light from substitutional atoms in an excited state.

(2) **Laser-interference microscopy** is a relatively new optical technique that detects small differences in refractive index, and so can reveal, in great detail, subtle compositional differences (on which refractive index depends), for example in zoned plagioclase (Chao, 1976; Pearce, 1984a,b; Pearce *et al*., 1987a,b).

(3) **Scanning electron microscopy** (see, for example, Lloyd, 1987) is capable of revealing sharp microstructural details in shades of grey, although arbitrary colours may also be assigned to form a false-colour image. It involves

backscattered and forescattered imaging in the scanning electron microscope (SEM). This is particularly useful for (a) revealing the detailed microstructure of small grains and fine-grained aggregates and intergrowths (see, for example, Vernon & Pooley, 1981; Wirth & Voll, 1987; Cashman, 1988; Simpson & Wintsch, 1989; Swanson *et al.*, 1989; Johnson & Carlson, 1990; van der Voo *et al.*, 1993; Brodie, 1995; Lloyd & Prior, 1999; Drüppel *et al.*, 2001; Blundy & Cashman, 2001; Rickers *et al.*, 2001; de Haas *et al.*, 2002; Schieber, 2002), (b) identifying very fine-grained minerals (see, for example, Prior *et al.*, 1999), (c) revealing fine-scale compositional zoning in minerals (see, for example, Yardley *et al.*, 1991; Müller *et al.*, 2000; Piccoli *et al.*, 2000; Kuritani, 2001; Alexandrov, 2001; Rubatto *et al.*, 2001; Hermann *et al.*, 2001; Ginibre *et al.*, 2002a,b; Lentz, 2002), (d) measuring orientation differences between grains and subgrains as small as 1 μm across (Prior *et al.*, 1996, 1999; Lloyd *et al.*, 1997; Trimby *et al.*, 1998; Wheeler *et al.*, 2001; Bestmann & Prior, 2003), and (e) revealing domains of different orientation in optically isotropic minerals, such as garnet (Spiess *et al.*, 2001; Prior *et al.*, 2000, 2002) and pyrite (Boyle *et al.*, 1998).

(4) **Transmission electron microscopy** (TEM) assists in the interpretation of some microstructures that can be optically ambiguous, such as some recovery features in deformed quartz (Section 5.4) and fine exsolution lamellae (Section 4.11). The principles and some applications have been reviewed by Champness (1977), Putnis & McConnell (1980), McLaren (1991) and Putnis (1992). TEM resolves very small objects, such as very fine to submicroscopic intergrowths, exsolution features, inclusions and twins and can reveal the arrangement of defects (including dislocations, discussed in Section 5.3.2) in the atomic structure of individual grains of both optically transparent and opaque minerals (see, for example, McLaren *et al.*, 1967; McLaren, 1974, 1991; McLaren & Retchford, 1969; Phakey *et al.*, 1972; Green & Radcliffe, 1972; McLaren & Hobbs, 1972; Champness & Lorimer, 1976; Champness, 1977; McLaren & Etheridge, 1976; Zeuch & Green, 1984; Doukhan *et al.*, 1985; Allen *et al.*, 1987; Cox, 1987a; Couderc & Hennig-Michaeli, 1989; Hennig-Michaeli & Couderc, 1989; Green, 1992; Ando *et al.*, 1993; Doukhan *et al.*, 1994; Vogelé*et al.*, 1998).

(5) **X-ray tomography** (in full: high-resolution computed X-ray tomography) is a more recent development in the study of rock microstructure (see, for example, Mees *et al.*, 2003). This technique maps the variation of X-ray attenuation within solid objects, the attenuation varying with each mineral present. A source of X-rays and a set of detectors revolve around the rock sample, producing images in layers or cross-sections. The series of two-dimensional images can be computed into a three-dimensional representation of the grains and aggregates in the rock (see below), which gives a clearer picture of spatial relationships and crystal size distributions (see, for example, Carlson & Denison, 1992; Carlson *et al.*, 1995, 1999; Denison *et al.*, 1997; Brown *et al.*, 1999; Philpotts *et al.*, 1999).

(6) **Computer-aided construction of three-dimensional images** is a technique whereby serial two-dimensional optical or X-ray tomographic images can be scanned and imported into suitable computer graphics programs to provide three-dimensional constructions (Johnson & Moore, 1993, 1996; Carlson *et al.*, 1995, 1999; Pugliese & Petford, 2001; Castro *et al.*, 2003). This approach is used in confocal laser scanning microscopy (CLSM), which enables objects (such as inclusions in minerals) to be viewed in focus through a thickness of sample in a single image (Petford & Miller, 1992; Petford *et al.*, 1995; Sheppard & Shotton, 1997; Bozhilov *et al.*, 2003), as described in Section 4.4.5. Readily available computer software can also be used to animate images, producing a more complete visualization of features such as grain shapes, grain distributions and vein networks (see, for example, Johnson & Moore, 1996; Carlson *et al.*, 1999; Pugliese & Petford 2001).

(7) **X-ray compositional mapping** produces maps of compositional zoning in crystals (see Sections 3.11, 4.12). Such maps are produced by multiple stage-scan chemical analyses made with wavelength-dispersive spectrometers on an electron microprobe, different colours being assigned to different concentrations of the analysed element. Examples are shown in Section 4.12. The technique can also be used for more clearly revealing mineral or compositional domains in fine-grained aggregates (see, for example, Lang & Gilotti, 2001; Williams *et al.*, 2001; Clarke *et al.*, 2001; Daczko *et al.*, 2002a,b). Raw X-ray intensity maps may be converted to maps of oxide mass percent by appropriate matrix corrections (Clarke *et al.*, 2001).

1.7 Quantitative approaches

Although most work on rock microstructures is qualitative, involving description and interpretation, quantitative methods are also used. For example, grain measurement is important in the classification and interpretation of clastic sedimentary rocks in terms of transport and depositional environments (Section 2.2.3). Grainsize is also used in the classification of igneous rocks, though less precisely, and crystal size distributions (CSD) are being increasingly investigated in igneous and metamorphic rocks (Sections 3.4, 4.3.1). Various statistical techniques are used to determine spatial distribution patterns (SDP) of grains and crystals in rocks (see, for example, Flinn, 1969; Kretz, 1966b, 1969; Jerram *et al.*, 1996; Denison & Carlson, 1997; Daniel & Spear, 1999; Jerram & Cheadle, 2000). Numerical modelling has been used to convert two-dimensional measurements of grain shapes and sizes in thin section to three-dimensional grain shapes and true crystal size distributions (Higgins, 1994, 2000; Peterson, 1996). Moreover, computer software is readily available to do this and to make animated images, as mentioned in the previous section. Interfacial angles have been measured in many metamorphic rocks, sulphide rocks and igneous cumulates, as indicators of mutual solid-state growth of minerals (Section 4.2). In addition, the orientations of inclusion trails in porphyroblasts have been used as indicators of tectonic processes (Section 5.10). Numerical simulation of the development of metamorphic and deformation microstructures is also well under way (Jessell, 1988a,b; Jessell *et al.*, 2001).

1.8 Some terms

Although no hard and fast rule exists, it is probably best to use *crystal* for a volume of crystalline mineral with well-formed, planar faces (called crystal faces or facets), and *grain* for any other volume of crystalline mineral. For me, the shapes, arrangements and orientation of the minerals constitute a rock's *fabric*. At the microscope scale, the fabric (*microfabric*) consists of the grain shapes and arrangement (the *microstructure*) and the spatial orientation of the minerals (the *preferred orientation*). However, many people equate 'fabric' with 'preferred orientation', as recommended by the IUGS Subcommission on the Systematics of Metamorphic Rocks (Brodie *et al.*, 2002).

It would be good to make materials scientists more interested in rocks, as they are the great class of natural solid materials. Therefore, because 'texture' means 'preferred orientation' to most materials scientists and an increasing number of structural geologists, it would be best not to use it instead of 'microstructure' as many petrologists do. However, although 'microstructure' is gaining in popular usage, 'texture' is common, and no ambiguity is caused among petrologists by using it. Actually, 'microstructure' appears to have priority, because the first publications on the microscopic examination of rocks referred to 'microscopical structure' or 'microscopic structure' (see, for example, Sorby, 1851, 1858; Dawson, 1859; Allport, 1874). Moreover, the IUGS Subcommission on the Systematics of Metamorphic Rocks has recommended that the term 'texture' be replaced by 'microstructure', which is defined as 'structure on the thin section or smaller scale' (Brodie *et al.*, 2002). The term 'microtexture', which unfortunately is starting to enter the literature, is a tautology, because 'texture' refers to the microscopic scale.

Of course, every gradation in scale exists between the microscopic and macroscopic (outcrop) scales, and so I have not been able to confine the discussion to the microscopic, although this is by far the main scale discussed.

1.9 Traditional rock groupings

Many rock-forming processes apply to more than one of the traditional igneous, sedimentary and metamorphic rock groups. For instance, similar basic principles governing the nucleation and growth of crystals apply to all rocks, and grain growth in the solid state occurs not only in metamorphic rocks (in which it is a universal process), but also in the late stages of formation of some rocks conventionally regarded as igneous. In addition, growth of new minerals in the solid state (*neocrystallization*) occurs not only in metamorphic rocks, but also in the late-stage alteration (*deuteric alteration*) of igneous rocks, and in the low-temperature alteration (*diagenesis* or *burial metamorphism*) of rocks that many people would consider to be still sedimentary. Moreover, metamorphic rocks begin to melt at high temperatures, producing rocks with both igneous and metamorphic features. In addition, radiating crystal aggregates (spherulites) commonly grow in glass, which, though technically solid, is liquid-like with regard to its atomic structure. Furthermore, exsolution, which is a solid-state process, occurs in both igneous and metamorphic minerals. As if that isn't enough, fragmental material thrown out of explosive volcanoes produces rocks that are technically sedimentary, but consist entirely of igneous material, and may also show evidence of solid-state flow of glass. The result of this cross-linking of processes is that, though this book adheres roughly to the traditional sedimentary–igneous–metamorphic subdivision, processes discussed under one of these headings may also be relevant to another of these groups. These instances are cross-referenced.

1.10 Importance of evidence

Science relies on *evidence*. An assertion made without evidence isn't worth very much. Yet I often read statements such as: 'the microstructural (textural) evidence indicates . . .' This implies that the writers are asserting that their interpretations are so obviously right that they don't have to go to the bother of describing what they saw and evaluating the evidence.

Of course, recognizing evidence takes practice. As noted by Chalmers (1999):

> It is necessary to learn how to see expertly through a telescope or microscope, and the unstructured array of bright and dark patches that the beginner observes is different from the detailed specimen or scene that the skilled viewer can discern.

Whenever you make interpretations based on microscopic examination of rocks, you should: (1) *describe* clearly what you see and (2) *evaluate* the possible interpretations. If one or more interpretations are valid, you should not arbitrarily favour one of them, unless other evidence (e.g. field or chemical evidence) clearly points in that direction. This is the 'method of multiple working hypotheses' advocated by Chamberlain (1890). In many instances, the microstructural evidence may not be at all clear, in which case you shouldn't use it to support a hypothesis. Maybe you will have to suggest equally valid alternative interpretations and leave it at that.

The paramount importance of evidence in making scientific inferences is emphasized in the following quotation.

> On so important a question, the evidence must be airtight. The more we want it to be true, the more careful we have to be. No witness's say-so is good enough.

> People make mistakes. People play practical jokes. People stretch the truth for money or attention or fame. People occasionally misunderstand what they're seeing. People even sometimes see things that aren't there.

Carl Sagan (1996) was referring to UFOs, but at least some of these statements could refer to petrologists interpreting rock microstructures. People do make mistakes and even see things that aren't there, and though practical jokes may be uncommon in such a serious pursuit as petrology(!), people certainly do occasionally misunderstand what they're seeing. We all do, in fact. Most important, we often want something to be true so much that we may be tempted to gloss over the evidence, whereas we should be doubly careful, in order to save ourselves falling into the trap of a woefully wrong interpretation, no matter how attractive it may seem.

It doesn't matter how many times an assertion is repeated or how loudly it is trumpeted in conversation, in the scientific literature, in textbooks or even on the Internet, it is only as good as the evidence for it. Another point to remember is that an interpretation presented by a great authority on the subject, though worthy of respect perhaps, is also only as good as the evidence for it. Such 'arguments by authority' can subdue interpretations based on careful accumulation of evidence (Vernon, 1996b).

Too often we see examples of interpretations based on inadequate evidence used to support a preferred model. Even some well-accepted interpretations may be wrong. A good example is the common belief that an 'order of crystallization' in igneous rocks can be inferred by looking at the microstructure. Generally this is impossible, as explained in Section 3.6. If the microstructure cannot give you the evidence, please don't try to extract it anyway!

In fact, the more I examine and read about rock microstructures, the more cautious I become about interpreting them, and that will be a constant theme in this book. As mentioned below (Section 1.11), recent work on the direct microscopic observation of developing microstructures in organic compounds used as mineral analogues has revealed many unexpected processes, and has shown that similar microstructures may have very different histories. They remind us of the necessity for caution in the interpretation of natural rocks.

So I will try to give explanations that are sufficiently general to be regarded by most people as 'reasonable' on the available information, and that students can infer largely from the optical microstructure. Where alternatives need to be discussed, the relevant publications will be mentioned. Moreover, where pitfalls exist, they will be pointed out, and it must be re-emphasized that the book is only the most general of guides. It indicates what to look for and how to start (not stop) thinking about what is observed. Finally, it is as well to keep in mind the wise words of John Spong (*Sydney Morning Herald*, June 9, 2001): 'Explanations are always compromised by the levels of knowledge available to the explainer.' That applies to me just as much as to any other explainer!

1.11 Kinds of evidence used

What evidence is useful in interpreting rock structures? Imagine you had never seen a rock section, either a thin section or a slab cut through a hand sample. How could you begin to interpret the crystal shapes and arrangements you see? You must have some guides. These are *field relationships* and *experimental evidence* on rocks and minerals, assisted by some general inferences from experiments on other materials, such as metals, ceramics, organic polymers and synthetic ice. For example, when Sorby first looked at thin sections of slates with the microscope, he would have already known that slates are formed by strong deformation and that the deformation is in some way responsible for their characteristic strong foliation (slaty cleavage). Furthermore, once Sorby had observed and described the features shown by the microstructure of the slate, others were in a position to recognize similar cleavages in thin sections of rocks from other areas. In this way, general guides to the interpretation of rock microstructures have been established.

If we can observe rocks forming, as with sedimentary and volcanic rocks, we are on firm ground for making inferences about how the microstructures were formed. We are on much shakier ground when it comes to intrusive igneous and metamorphic rocks. However, we can learn much from careful interpretation of field relationships, although strong differences of interpretation are common. In addition, experiments on the cooling of melted rocks and the melting of solid rocks are valuable guides to the interpretation of rocks involving melts, and many recent experimental advances have been made in the interpretation of igneous microstructures (see, for example, Lofgren, 1971b, 1973, 1974, 1976, 1980; Lofgren *et al.*, 1974; Fenn, 1974, 1977, 1986; Donaldson, 1976, 1977, 1979; Swanson, 1977; Swanson & Fenn, 1986; London, 1992). However, we should keep in mind possible problems caused by the short duration of experiments.

Experimentally determined stability fields of mineral assemblages in different bulk chemical compositions reveal the conditions of pressure, temperature and fluid composition that occur during metamorphism. However, it is not as easy to conduct successful experiments on the development of microstructures in metamorphic rocks, because of the high temperatures and pressures involved in the experiments and the generally small size of the samples used.

Many important experiments on mineral and rock deformation have been carried out (Chapter 5), but again we can only observe the finished product, not the stages along the way. Fortunately, experiments on ice deformation have helped our understanding of progressive microstructure development during deformation (see, for example, Wilson, 1984, 1986; Wilson *et al.*, 1986). Moreover, a new experimental technique based on transparent and translucent organic compounds that behave somewhat similarly to minerals has been developed, and is being applied with great effect to the interpretation of microstructures, especially deformation features (see, for example, Means, 1977, 1981, 1983, 1989; Means & Jessell, 1986; Means & Park, 1994; Means & Ree, 1988; Park & Means, 1996;

Urai, 1983a,b, 1987; Urai & Humphreys, 1981; Urai *et al.*, 1980, 1986; Ree, 1991; Ree & Park, 1997). Because these compounds deform, melt and crystallize rapidly at room temperatures, the processes can be observed and photographed in progress in the microscope ('see-through' experiments). Of course, we must keep in mind that these materials are generally not minerals, but they have provided some startling insights into possible grain-scale processes that may occur in natural rocks.

In the absence of reliable experimental evidence, it is necessary to fall back on 'common-sense' interpretations, based on accumulated experience of the type outlined previously. This applies especially to metamorphic rocks. Unfortunately, common sense isn't so common, and what makes perfectly good sense to one person may make no sense at all to somebody else. The most important thing is to be as honest and logical as possible, and to evaluate (and if necessary retain as possibilities) every interpretation that can reasonably explain the observations. If the end result is the unsatisfying conclusion that you cannot make an unambiguous interpretation on the available evidence, leave it at that. No harm will be done. On the contrary, many a doubtful interpretation, presented as being reliable, has been accepted at face value and used in later work, thereby misleading subsequent researchers.

1.12 Complexity

A rock's microstructure is the product of a complicated sequence of events and processes. So is a rock's chemical analysis. Both may tell us something about the rock's history, but neither can fully reveal all the historical complexities. This is a problem that petrologists have to accept. We do our best with the evidence available, without taking it too far, and we must acknowledge that our interpretations are often incomplete.

Another point to add to the complexity is that superficially similar microstructures may be formed in different ways, as with exsolution and epitactic (epitaxial) replacement producing similar intergrowths (see, for example, Craig, 1990). For example, hematite lamellae in magnetite, usually inferred to be of replacement origin, may be due to exsolution in some rocks (Edwards, 1949). Another complication is the optical similarity between subgrains formed by recovery and similar features formed by fracture, as discussed in Section 5.4. Other complexities of rock microstructure will become apparent in the following chapters.

Chapter 2

Microstructures of sedimentary rocks

2.1 Introduction

Although this book is mainly concerned with igneous, metamorphic and deformation processes, in this chapter I briefly review the main sedimentary microstructures, partly because they need to be understood in order to interpret residual sedimentary microstructures in some metamorphic rocks.

The basic sedimentary microstructures are relatively straightforward, but variations (reflecting variable sedimentary environments) can be very complex, and many complicated classification schemes to deal with this complexity have been suggested. Classification schemes, microstructural details and discussions of sedimentary environments can be found in many excellent books (e.g. Pettijohn, 1949; Williams *et al.*, 1954; Carozzi, 1960; Milner, 1962; Folk, 1968; Selley, 1970; Blatt *et al.*, 1972; Bathurst, 1975; Friedman & Sanders, 1978; Adams *et al.*, 1984, Greensmith, 1989; Boggs, 1992; McPhie *et al.*, 1993; McLane, 1995).

Sediments are loose, unconsolidated fragments, and sedimentary rocks are the consolidated or lithified equivalents. *Residual (pedogenic)* sediments are essentially *in situ* deposits of the products of rock weathering, for example weathering crusts, soils and regoliths. *Epiclastic* (*terrigenous*) sediments are accumulations of solid fragments formed by erosion of existing rocks. *Pyroclastic* sediments are deposits of fragmented igneous material (e.g. volcanic glass and crystal fragments) ejected from volcanoes in explosive eruptions directly onto Earth's surface. *Bioclastic* sediments are accumulations of organic skeleton or shell fossils that have been at least slightly transported. *Chemical* sediments are precipitated directly from aqueous solution or by replacement of existing sediment. *Polygenetic* sediments consist of mixtures of the foregoing types, and are named according to the dominant kind of sediment present (e.g. fossiliferous limestone, tuffaceous sandstone). Moreover, epiclastic rocks commonly contain chemical components in the form of a *cement* (Section 2.2.5).

2.2 Epiclastic ('terrigenous') sedimentary rocks

2.2.1 Detrital (clastic, fragmental) minerals

Quartz and the clay minerals constitute up to about 70–80% of the epiclastic sedimentary rocks, with less abundant feldspar, mica and carbonate. The clay minerals belong to a group of very fine-grained, water-rich, complex aluminosilicates with various other elements, especially potassium, magnesium and iron. Generally they are too fine-grained to be identified with the optical microscope, and so other techniques (especially X-ray diffraction) are used for their accurate determination. Other chemically and physically resistant minerals, such as zircon, tourmaline, ilmenite, magnetite, monazite, rutile, topaz and garnet, are also commonly present in very small quantities: generally no more than 1%, except in local 'black sand' concentrations.

The detrital (clastic, fragmental) minerals in epiclastic sediments depend on (1) their presence in the source rocks, (2) their resistance to mechanical abrasion, and (3) their chemical stability in the surface environment. Common fragmental minerals are those that are most stable in Earth's atmosphere, especially quartz and the clay minerals. Quartz is the most common mineral that is stable in Earth's atmosphere, and is released from rocks undergoing weathering as individual grains or fragments. In contrast, clay minerals are formed by the chemical breakdown of minerals that are not as stable in the atmosphere, especially feldspars and the ferromagnesian minerals (pyroxene, olivine, biotite and amphibole). However, these less stable minerals can occur as fragments in sediment that has been transported relatively short distances and/or deposited rapidly.

2.2.2 Fragment size and sorting

Transport at normal conditions in water tends to sort the fragments (detritus) into different sizes, namely: pebbles ($> 2\,\mathrm{mm}$ in diameter), sand ($2 - \frac{1}{16}$ mm in diameter), silt ($\frac{1}{16} - \frac{1}{256}$ mm in diameter) and clay ($< \frac{1}{256}$ mm in diameter). Prolonged washing of sediment, for example in waves in relatively shallow water (involving traction currents), leads to a well-sorted sedimentary rock with a relatively even grainsize (Figs. 2.1, 2.2), whereas if the sediment is deposited rapidly (for example, in turbidity currents produced by submarine slumps on continental slopes), the fragments tend to have very different sizes (said to be poorly sorted or unsorted), as shown in Figs. 2.3–2.7.

Generally, well-sorted sedimentary rocks tend to have rounded fragments (Section 2.2.4) and *vice versa* (Figs. 2.1, 2.5), but not necessarily (Figs. 2.2–2.4). Sedimentary rocks composed mainly of pebbles, sand and clay are called conglomerates or rudites (Figs. 2.7, 2.8), sandstones or arenites (Figs. 2.1–2.6) and shales, claystones, mudstones or pelites (Fig. 2.9), respectively.

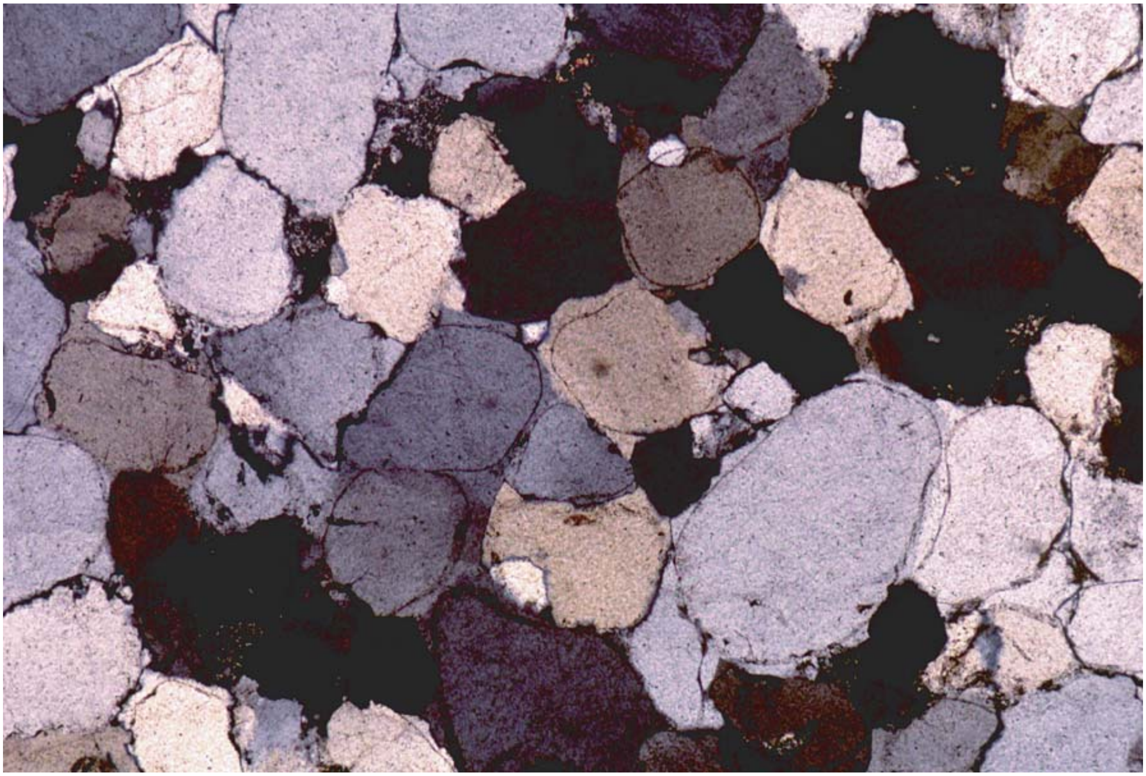


Fig. 2.1: Orthoquartzite, consisting of well-sorted, rounded clastic quartz grains (outlined by minute, dusty-looking inclusions) with secondary overgrowths (forming a quartz cement) in optical continuity with the clasts. The resulting new grains (clasts plus their overgrowths) form a roughly polygonal aggregate (compare with Fig. 2.13). Some of the grain contacts are irregular to sutured (stylolitic surfaces; Section 5.9.2), owing to 'pressure-solution' in response to local stress increases as the grains were pressed together, during either compaction or weak tectonic activity (Chapter 5). Crossed polars; base of photo 2.8 mm.

Poorly sorted sediments commonly contain a *matrix*, which is finer-grained material of detrital origin that partly or completely fills spaces (interstices) between the larger clastic grains that form the framework of the rock. For example, poorly sorted sandstones may have a clay and/or silt matrix, and poorly sorted conglomerates may have a sand matrix. Clay minerals in the matrix tend to bind the larger clasts, especially after being neocrystallized (Section 2.2.7) during burial, and so help produce a consolidated rock. In many volcanic sandstones (often called 'greywackes'), what appears to be a primary clay matrix may actually be an aggregate of clay minerals and/or chlorite that replace volcanic glass fragments and are squashed between the quartz, feldspar and rock fragments (Figs. 2.5, 2.6).

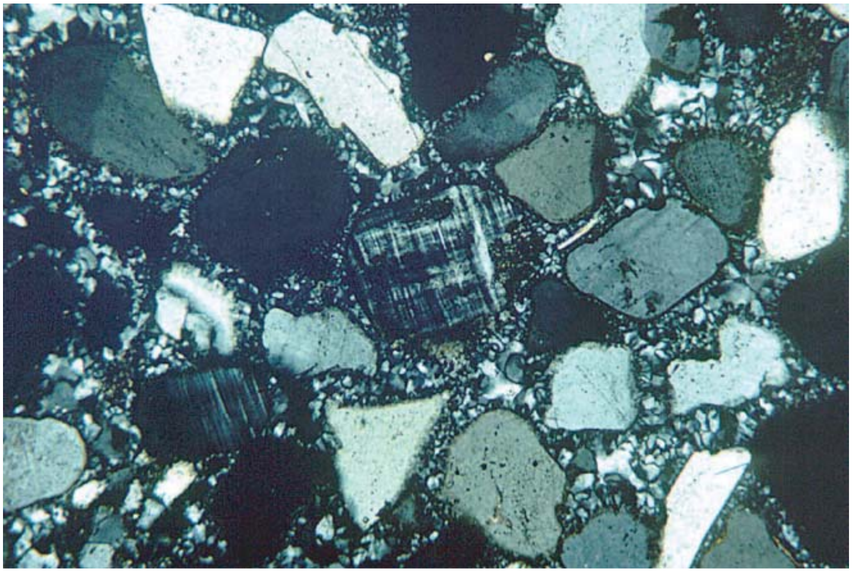


Fig. 2.2: Relatively well-sorted sandstone consisting of fragments of quartz and feldspar (mainly microcline, showing tartan twinning; Section 4.8) that are mainly well rounded, though some are sub-rounded to angular. The fragments have been cemented by fine-grained aggregates of quartz precipitated from hydrous solutions percolating between the clasts; the quartz cement grains nucleated on the clasts, forming 'micro-vughs'. From Vernon (2000b, fig. 103). Crossed polars; base of photo 1.3 mm.

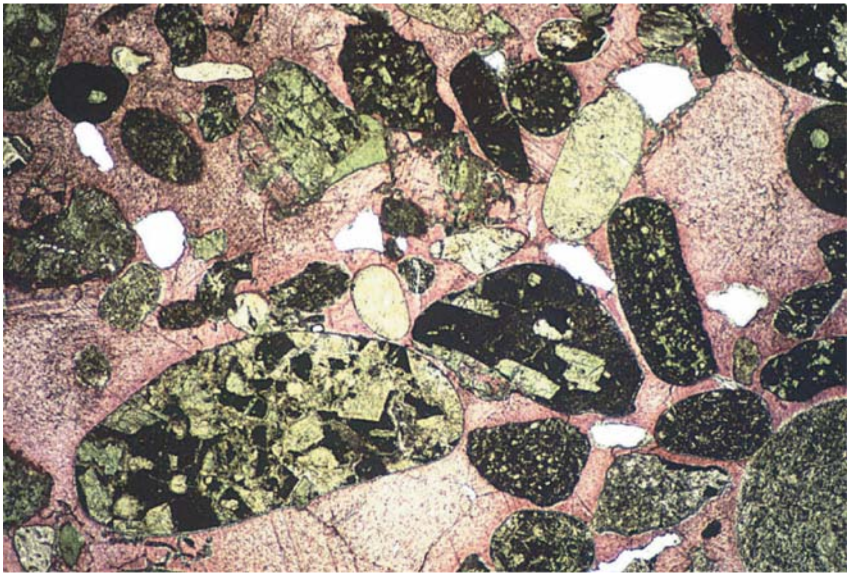


Fig. 2.3: Moderately poorly sorted volcanic sandstone composed of clasts of volcanic rock fragments with igneous microstructures, including phenocrysts (Chapter 3). The feldspar of the rock fragments has been replaced by fine-grained, green chlorite. The clasts have been cemented by calcite in grains that are much larger than the clasts ('*lustre mottling*' structure). The calcite (which is normally colourless in thin section) has been stained with an organic dye (alizarin red S), to distinguish it from other carbonate minerals. From Vernon (2000b, fig. 104). Plane-polarized light; base of photo 4.4 mm.

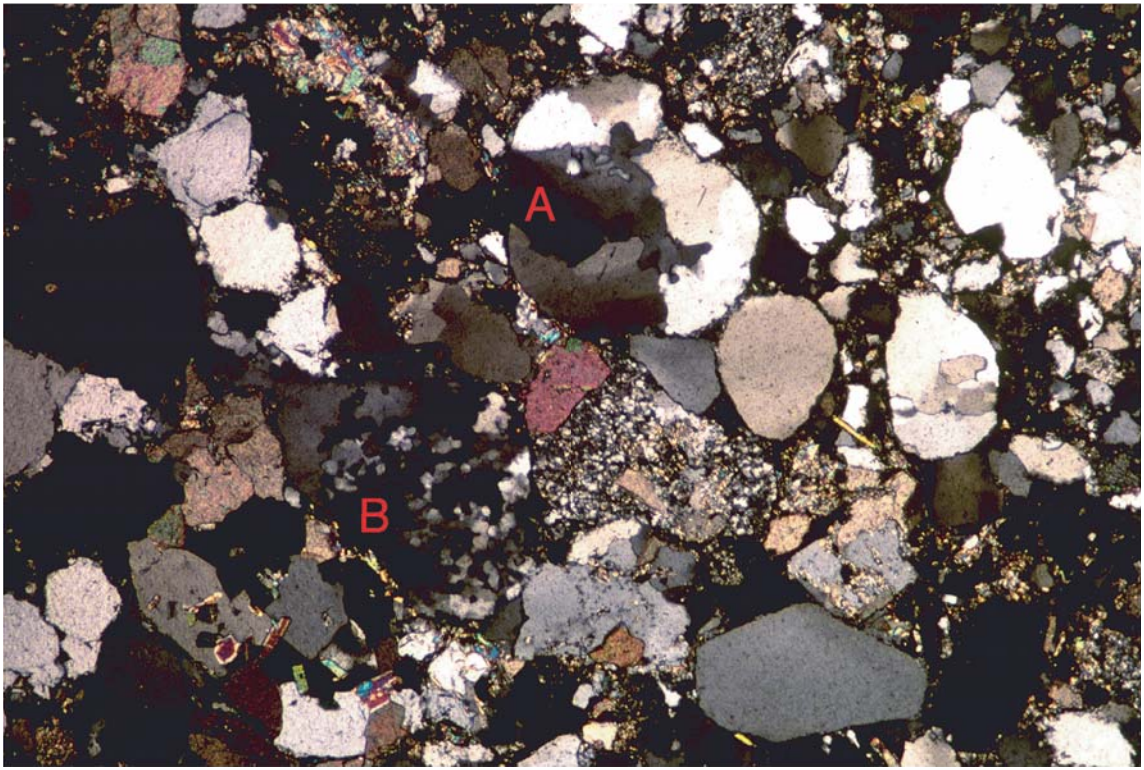


Fig. 2.4: Poorly sorted sandstone, consisting mainly of angular to rounded quartz fragments, with some altered feldspar, carbonate and fine-grained rock fragments, all cemented by opaque iron oxide. Some of the quartz clasts show evidence of recovery (A) and/or recrystallization (B), as discussed in Section 5.4, owing to deformation in the source area; such microstructures assist in determining the source (provenance) of the sediment. Crossed polars; base of photo 4 mm.

Turbidity currents (turbid flows) that transport and dump unsorted material commonly briefly interrupt slow deep-water (pelagic) deposition of clay-size fragments. The rapidly moving currents commonly disrupt the underlying mudstone beds and incorporate fragments of them (rip-up clasts), as shown in Fig. 2.10.

2.2.3 Fragment shapes

The general shapes of clastic grains depend largely on shapes inherited from the parent rock, modified by abrasion during transport. Abrasion of fragments wears away their sharp corners, eventually producing rounded sand grains and pebbles. Grains are said to be *angular, subangular, subrounded, rounded* or *well-rounded*, depending on their degree of rounding (Fig. 2.11). Softer mineral and rock fragments become rounded much more quickly than harder ones, quartz being so

A

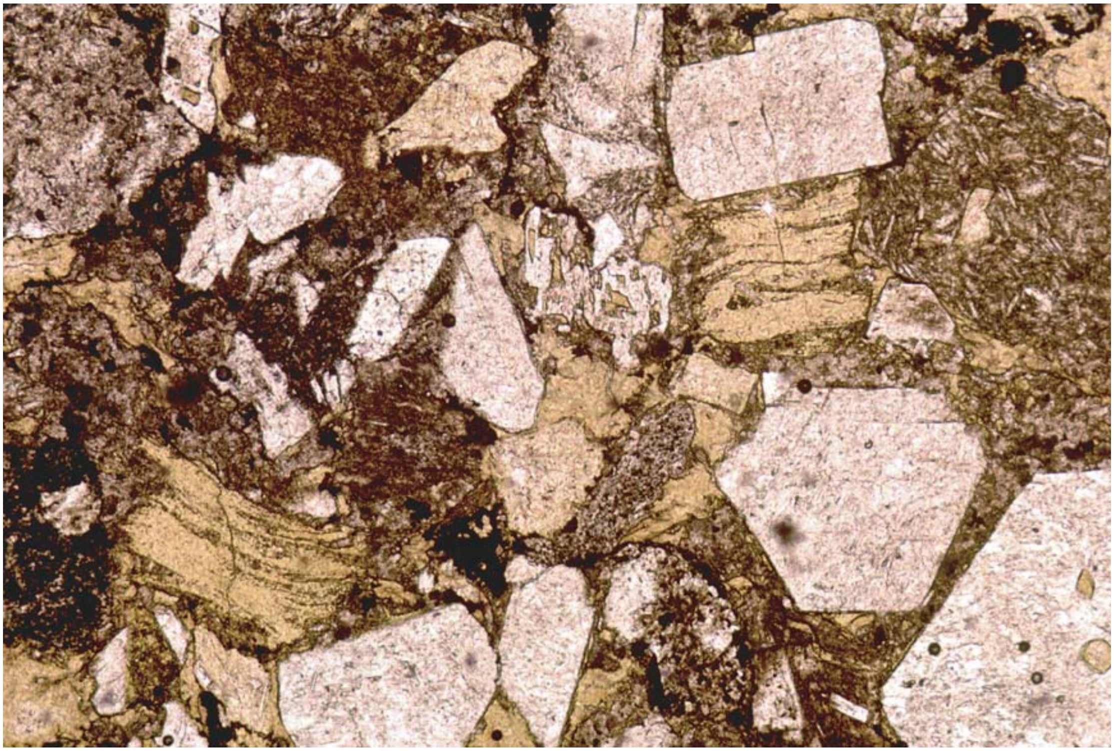


Fig. 2.5: (A) Poorly sorted volcanic sandstone ('greywacke') consisting of angular fragments of quartz and plagioclase (some with remnants of crystal faces) and volcanic rock fragments, some with feldspar laths aligned in a magmatic flow structure (right), as discussed in Section 3.9, with a fine-grained chlorite-rich matrix (pale yellowish green). The presence of former glassy volcanic rock fragments with residual flow lines, in which the glass has been replaced by similar chlorite, suggests that the chlorite in the matrix may also have replaced volcanic glass fragments, after which it was squeezed between the clasts by compaction. Plane-polarized light; base of photo 2.8 mm. (B) Same field of view. The almost isotropic nature of the matrix is due to the abundance of very fine-grained chlorite, with a low birefringence, and not unaltered volcanic glass. Crossed polars; base of photo 2.8 mm.

hard that it resists rounding in all sedimentary environments except high-agitation areas, such as beaches and wind dunes. Fragments that are not transported very far tend to have angular shapes (Fig. 2.10), whereas fragments transported over long distances and/or rubbed together in energetic water environments (such as waves) tend to have more rounded shapes (Figs 2.1, 2.7, 2.8).

Elongate grains (said to have a *low sphericity*) may be well rounded, and equant grains (*high sphericity*) may be angular. Roundness may also be inherited from grains in parent rocks, for example older sedimentary rocks with rounded quartz grains.

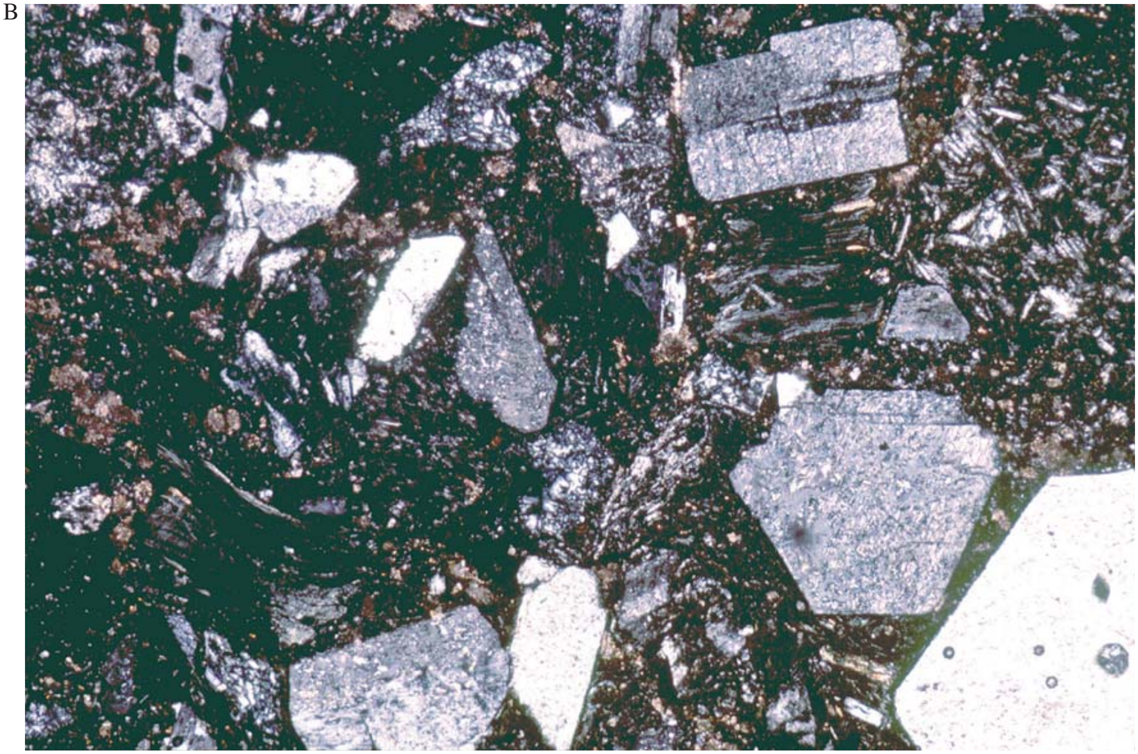


Fig. 2.5: (*cont.*)

2.2.4 Maturity

Epiclastic sediments may be classified on the basis of 'textural maturity'. *Immature* sediments have at least 5% matrix, with poorly sorted, angular grains; *submature* sediments have less than 5% matrix, with poorly to moderately sorted, angular grains; *mature* sediments have less than 5% matrix, with well-sorted, angular grains, and *supermature* sediments have no matrix, with well-sorted, rounded grains.

Catastrophic conditions, such as floods, coastal cyclones and mass-flows, are very unfavourable for sorting. For example, in mountainous areas, where the rain runoff and hence the erosion rate are very high, rock fragments of various sizes may accumulate with sand and soil in the lower regions as alluvial fans. Similarly, mass-flows in submarine canyons at the edge of a continental shelf dump poorly sorted material in turbidite fans. In addition, transport of sediment by ice in alpine glaciers does not involve much sorting or abrasion, so that fragments deposited in glacial moraines tend to be poorly sorted and angular.

Fig. 2.6: Clast of formerly glassy volcanic rock with embayed phenocrysts of quartz and smaller phenocrysts of altered feldspar, the glass having been replaced by very fine-grained chlorite (green). Embayments are common in volcanic phenocrysts, and so are relatively good indicators of volcanic source rocks. A few of the quartz clasts in this rock also show embayments, suggesting former volcanic phenocrysts (Section 3.13). Plane-polarized light; base of photo 5.2 mm.

2.2.5 Cement

Loose sediment is deposited in beds in subsiding areas of Earth's crust. As the sediment becomes buried, it is compacted, and groundwater begins to circulate through the spaces between the particles. Chemical compounds dissolved in this water may be precipitated as new minerals (*cement*) in spaces between the fragments, forming a solid sedimentary rock by cementing the clasts together (Figs. 2.1–2.4, 2.7, 2.8, 2.12).

The most common cement minerals are silica minerals (quartz, chalcedony, opal), carbonates (calcite, aragonite, dolomite, siderite), hematite, 'limonite' (hydrated iron oxides), chlorite and clay minerals (especially kaolinite). Cement should be contrasted with matrix (Section 2.2.2). Clay from outside the site of deposition and either deposited at the same time as the larger fragments (during

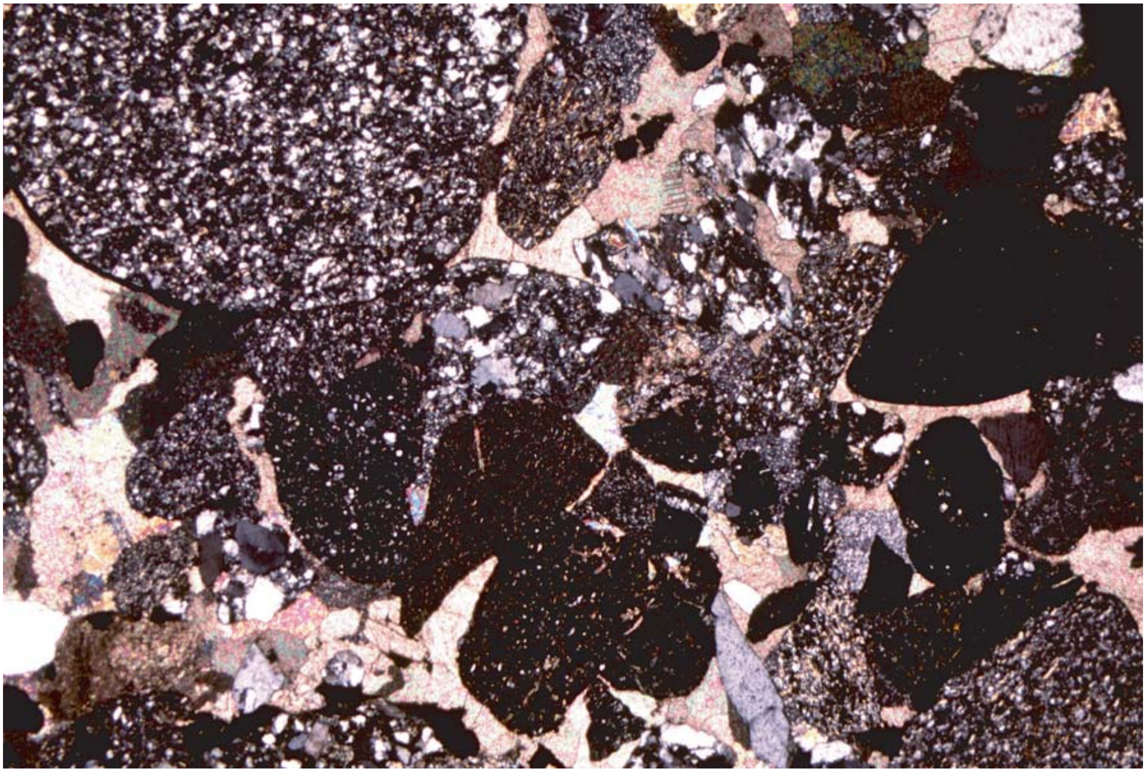


Fig. 2.7: Poorly sorted volcaniclastic conglomerate (sandy conglomerate) composed mainly of rounded clasts of volcanic rocks, cemented by very large grains of calcite (compare with Fig. 2.13). Crossed polars; base of photo 5.2 mm.

rapid deposition) or mixed with sand by bioturbation (churning by burrowing animals) forms a matrix between the larger clasts. This is called *allogenic* clay, in contrast to *authigenic* clay minerals, which are either precipitated chemically as a cement or formed by replacement of chemically unstable minerals, such as feldspar. Cement and matrix can coexist in coarser-grained, poorly sorted rocks consisting of larger clasts and finer-grained matrix, with cement-filled spaces between the larger fragments in the matrix.

In some rocks the cement grains are at random (Fig. 2.7), whereas in others they are arranged perpendicular to the clast surfaces, forming a kind of miniature vugh ('micro-vugh') structure (Section 5.9.3), as shown in Figs. 2.2 and 2.8. In other rocks, cement minerals (e.g., calcite) may occur locally as radiating aggregates (Fig. 2.12). Some cement minerals, especially quartz, nucleate in the same crystallographic orientation as clasts of the same mineral (epitactic nucleation; Section 4.3.1), forming *overgrowths*, the process being known as secondary enlargement (Figs. 2.1, 2.13). The overgrowth continues until the

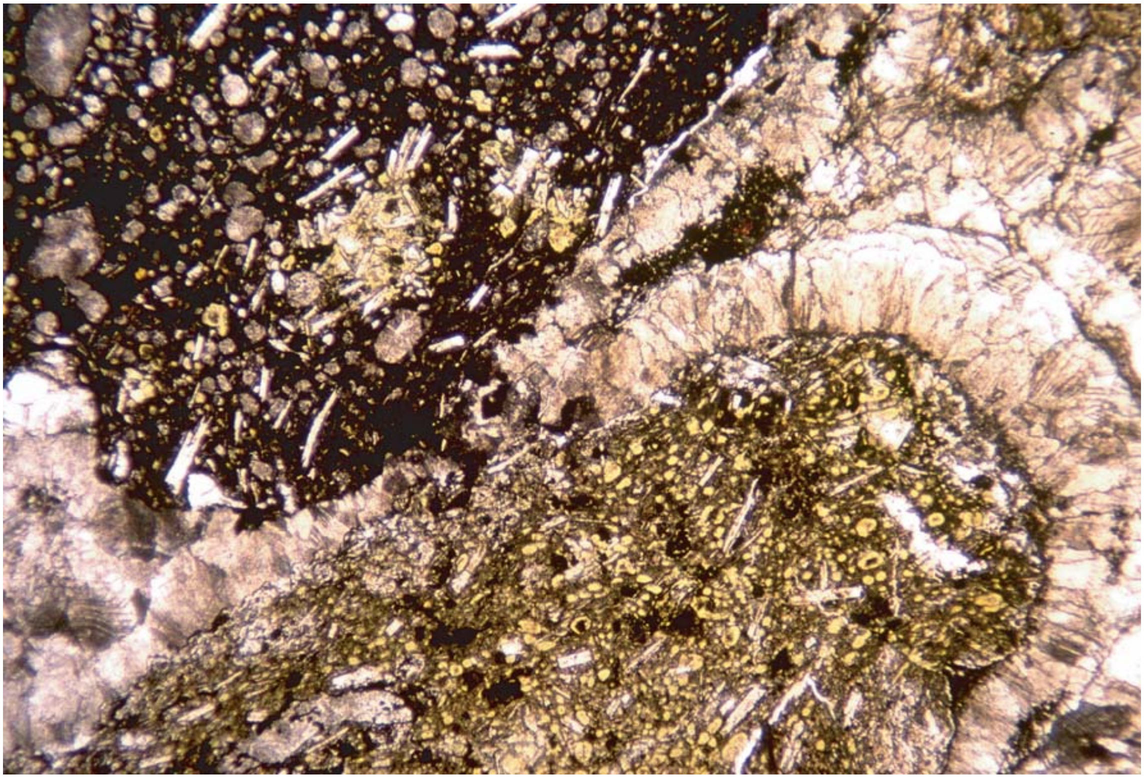


Fig. 2.8: Volcaniclastic conglomerate composed mainly of rounded clasts of amygdaloidal volcanic rocks, cemented by calcite grains that nucleated on the clasts and grew inwards towards the centres of the cavities between the clasts, forming a vugh-like microstructure. Plane-polarized light; base of photo 11 mm.

pores are filled, forming a mosaic of cement grains or 'overgrowth cement', the resulting new grains (fragments plus their overgrowths) tending to have polygonal shapes (Figs. 2.1, 2.13). The shapes of the original detrital grains are commonly outlined by minute inclusions (e.g. clay flakes or iron oxide particles). Quartz sandstones cemented in this way are called *orthoquartzites*, which should be distinguished from *metaquartzites*, which are formed by recrystallization of quartz sandstone under metamorphic conditions (Section 4.2), and which do not show overgrowths.

If the nucleation rate is low, the cement grains may be few but very large, each one cementing many clasts (Figs. 2.3, 2.7, 2.13). Common examples are calcite and gypsum in well-sorted quartz sandstones, but the situation may also occur in volcanic sandstones and conglomerates (Fig. 2.7). Commonly the cement crystals are large enough to show up in freshly broken hand specimens as shiny cleavage surfaces ('lustre mottling').

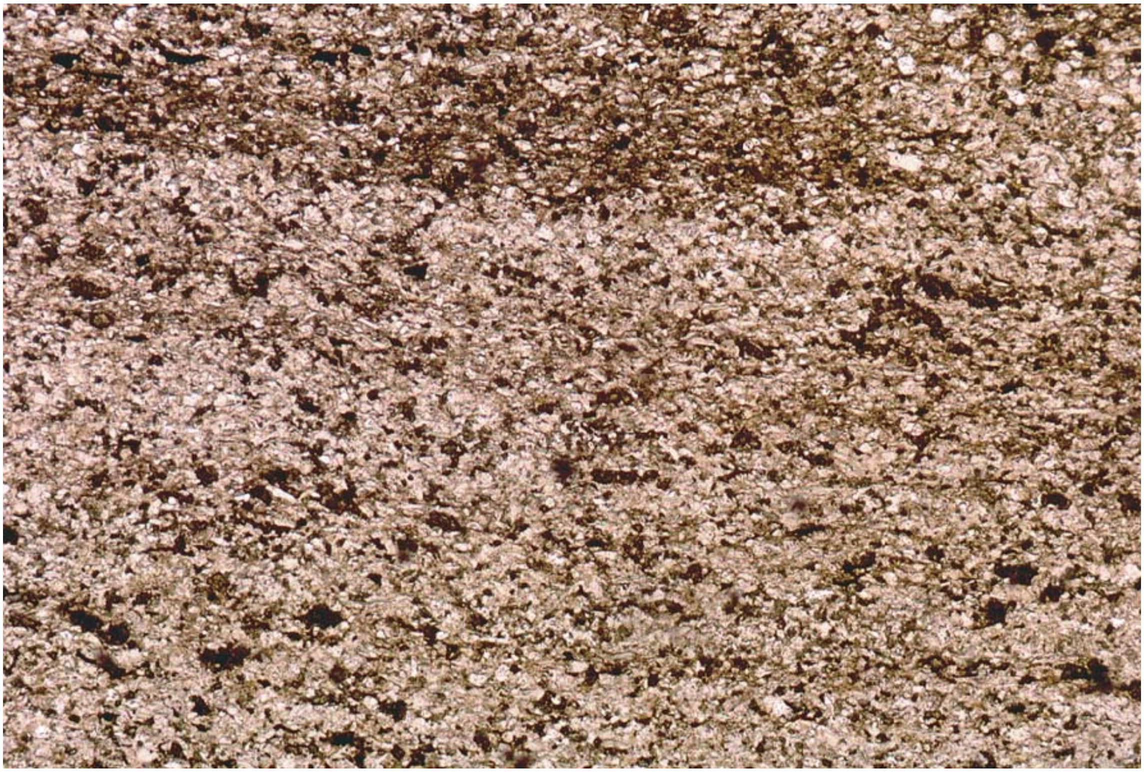


Fig. 2.9: Shale from the Wianamatta Group, Sydney area, New South Wales, Australia, showing flakes of clay minerals and/or mica aligned parallel to the bedding. Plane-polarized light; base of photo 2.8 mm.

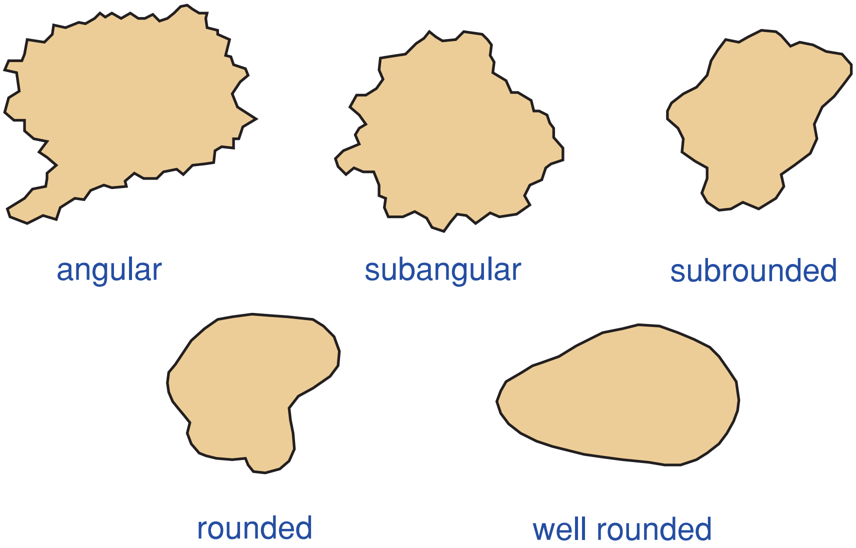
2.2.6 Source of sediment (provenance)

Pebbles in conglomerates and sand grains in sandstones may indicate what types of original rock were broken down to form the sediment. For example, the clasts in the sandstones of Figs. 2.3 and 2.12 and the conglomerates of Figs. 2.7 and 2.8 are mainly fragments of volcanic rocks, indicating that volcanoes were eroded to produce the fragments. Conglomerates are especially useful for determining provenance, because coarse-grained rocks, such as granites and high-grade metamorphic rocks, can be preserved as pebbles.

Sandstones containing fragments of quartz and feldspar imply coarse-grained quartzofeldspathic source rocks, such as granite or felsic gneiss, because weathering and abrasion can only make original grains smaller. Microcline (Fig. 2.2) suggests deformed granitic parent rocks (Sections 5.7.6, 5.7.9). Some microstructures of quartz clasts are distinctive enough to suggest broad parent rock-types. For example, embayed quartz grains (Section 3.13) indicate volcanic parent rocks (Fig. 2.6), which are best preserved in pyroclastic and

Fig. 2.10: Two dark mudstone chips (rip-up clasts) in a poorly sorted volcaniclastic sandstone with predominantly angular fragments and an altered clay (possibly former fragmental glass) matrix from the Abercrombie River, central western New South Wales, Australia. The incompletely consolidated nature of the mudstone clasts when they were incorporated in the sediment is indicated by their indentations by the other clasts. Plane-polarized light; base of photo 5.2 mm.

Fig. 2.11: Diagrams showing the main grain shape variations in clastic sedimentary rocks.



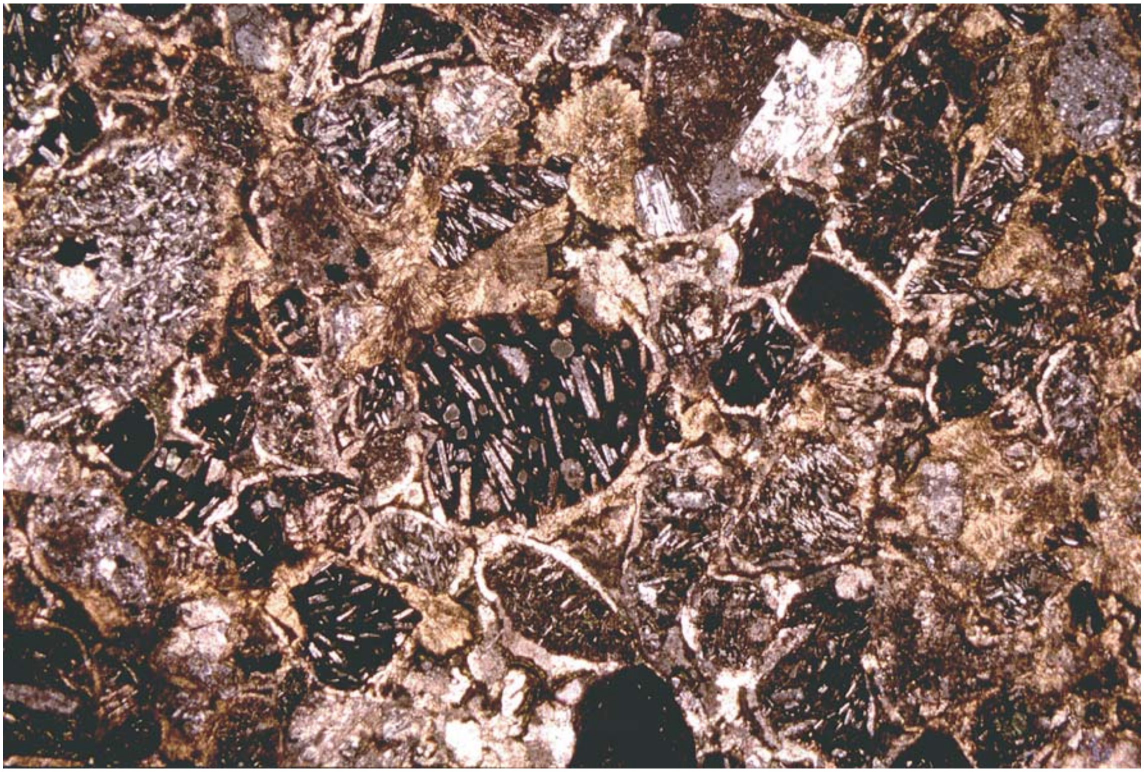


Fig. 2.12: Volcanic sandstone from the Illawarra area, south of Sydney, New South Wales, Australia, consisting of volcanic rock fragments (many of which show alignment of feldspar laths, reflecting flow of lava; Section 3.9) cemented by calcite, which locally occurs in radiating aggregates. Crossed polars; base of photo 5.2 mm.

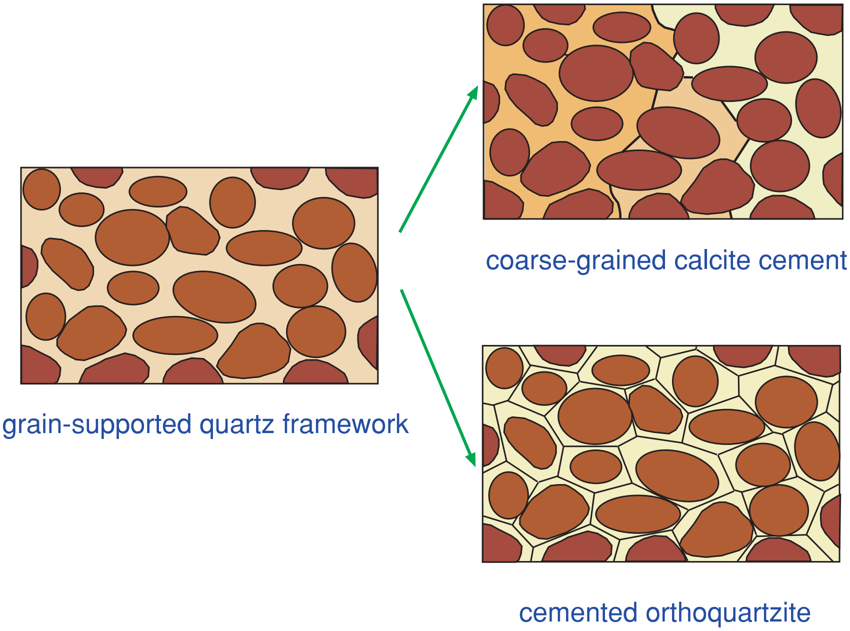


Fig. 2.13: Diagrams showing cementing overgrowths (secondary enlargement) on detrital grains, as with quartz in quartz sandstones (compare with Fig. 2.1) and cement formed by very large grains of calcite ('poikilitic microstructure', 'lustre-mottling').

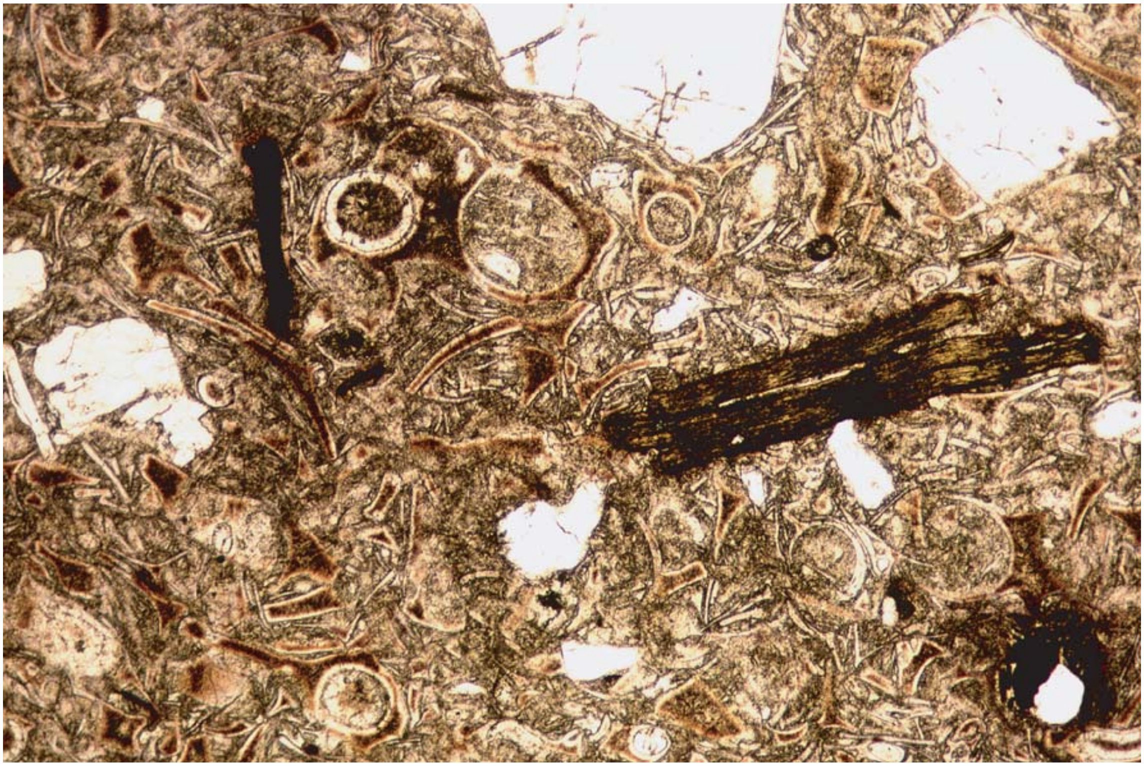


Fig. 2.14: Carboniferous ash-fall tuff from the Hunter Valley, New South Wales, Australia, composed mainly of glass and phenocryst fragments (quartz, feldspar and deformed biotite) ejected in an explosive eruption. The fragments (glass shards) were formed by fragmentation of pumice (Fig. 3.95), the curved edges, sharp corners and Y-shapes reflecting former vesicles in the pumice. A few small pumice fragments with one or two gas bubble holes are present. The glass has been replaced by hematite (reddish brown) and zeolite (colourless). Plane-polarized light; base of photo 4.4 mm.

volcaniclastic sediments, and strongly deformed quartz with the microstructural features described in Sections 5.3 and 5.4 (such as subgrains and recrystallization) indicates a regional metamorphic terrane (Fig. 2.4). Clasts made up of parallel, elongate quartz grains may have been derived from quartz veins in a regional metamorphic source region (Section 5.9.3).

Because weathering tends to break down minerals that are less stable in Earth's atmosphere, many sediments contain only a few of the minerals present in the source rocks. However, some of the minor minerals in the source rocks are relatively resistant to erosion and transport, and are deposited as dispersed grains or concentrated layers of 'heavy minerals', which may be general indicators of parent rock-types. For example, garnet and staurolite indicate metamorphic source

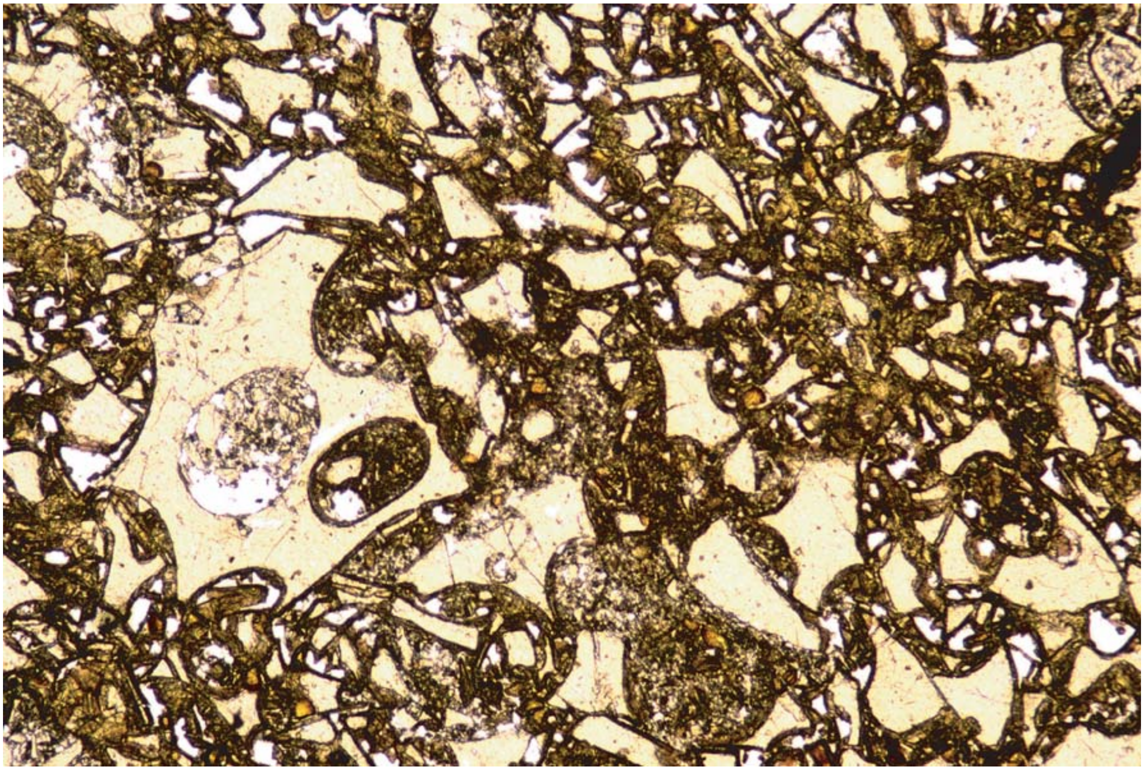


Fig. 2.15: Mafic tuff from Iceland, consisting of curved angular fragments (shards) of basalt glass that has been altered to submicroscopic chloritic aggregates. The larger clasts have gas bubble holes that have become filled or partly filled with small fragments. The sizes and shapes of the bubble holes in the large clasts have evidently controlled the shapes of the larger shards. Plane-polarized light; base of photo 4 mm.

rocks, chromite indicates ultramafic source rocks, and tourmaline and topaz are consistent with granitic or metamorphic source rocks.

2.2.7 Diagenetic changes

Minerals formed after deposition of the detrital (allogenic) materials are said to be authigenic. Some authigenic minerals (e.g. cement minerals) are deposited from aqueous solution, but many are formed from low-temperature reactions involving detrital material, the process bring called *diagenesis*. For example, volcanic glass clasts (Section 2.2) may alter ('neocrystallize') to clay minerals, chlorite or zeolites (Figs. 2.3, 2.5, 2.6), and detrital clay minerals may dehydrate and alter to chlorite and/or white mica. These reactions are very common in the matrix of sandstones. Another example of a post-depositional change

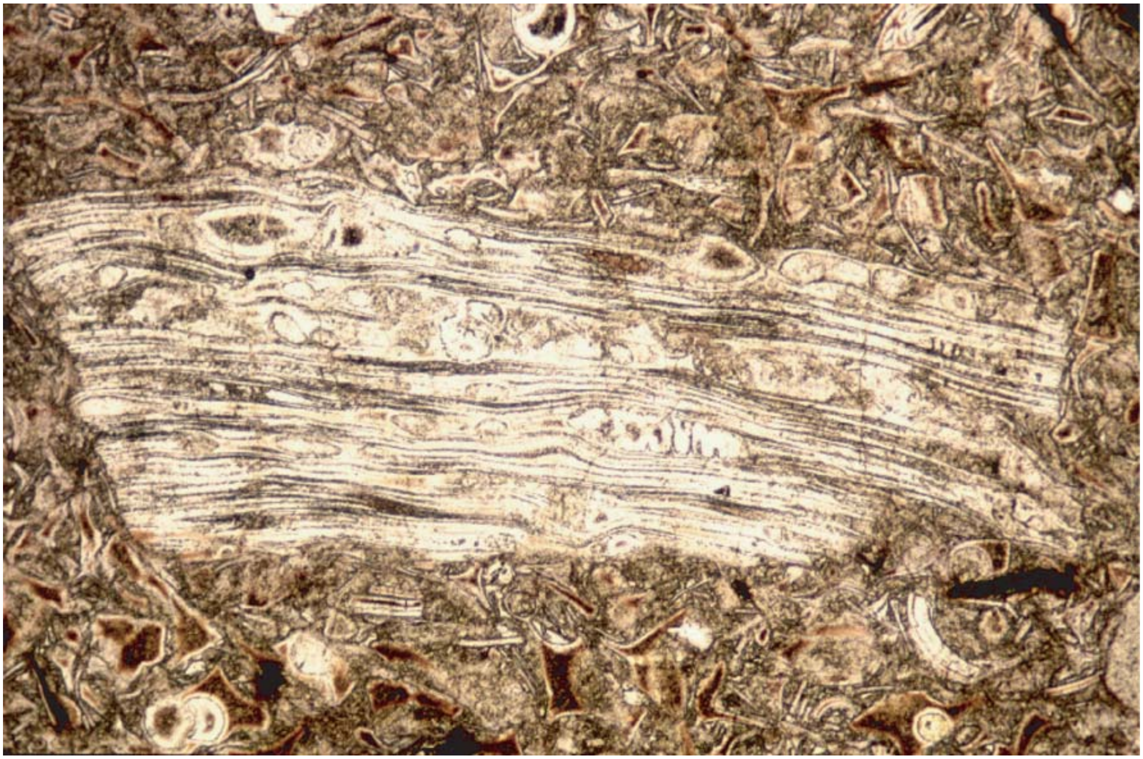


Fig. 2.16: Carboniferous ash-fall tuff from the Hunter Valley, New South Wales, Australia, similar to that shown in Fig. 2.14, but showing a large clast of highly vesicular glass with flow structure (Section 3.9). From Vernon (2000b, fig. 86). Plane-polarized light; base of photo 4.4 mm.

is the replacement of fragmental volcanic glass by opaline silica or silica gel, which are eventually replaced by very fine-grained quartz, forming a variety of *chert*.

Some diagenetic changes occur in sediments even before they are buried, for example recrystallization of aragonite to calcite and the alteration of some clay minerals to minerals such as glauconite, illite and chlorite. However, many diagenetic changes occur after burial, at considerable depths in the sedimentary succession, so that diagenesis grades into low-grade metamorphism (burial metamorphism). For example, zeolites replaced glass and detrital minerals in volcaniclastic sediments buried to depths of up to 15 km in Southland, New Zealand (Coombs, 1954; Turner, 1981). Identical minerals were formed at much shallower depth in volcaniclastic sediments in the Kiama area, south of Sydney, Australia, owing to geothermal heat in a subvolcanic region (Raam, 1968). These mineralogical changes are generally classified as belonging to the zeolite facies of regional metamorphism (Chapter 4).

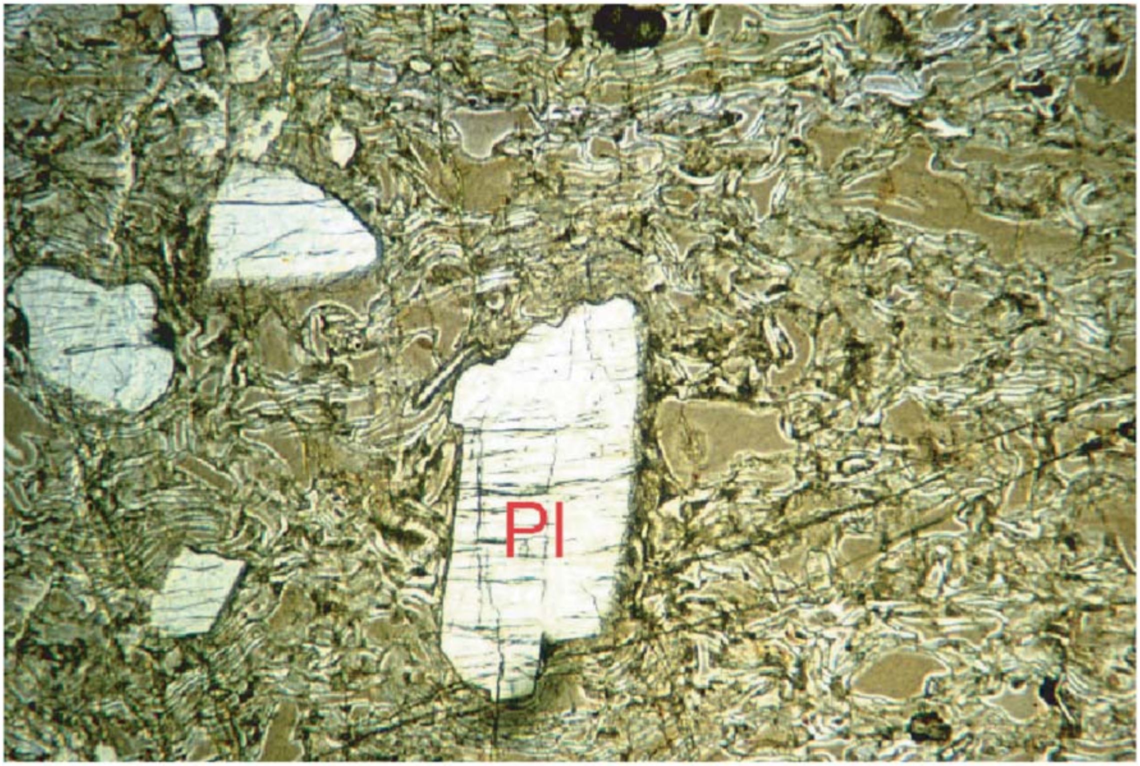


Fig. 2.17: Welded ash-flow tuff (ignimbrite) from Barraba, New South Wales, Australia, consisting of fragments of plagioclase (Pl) phenocrysts and abundant glass shards with rounded corners, owing to welding during cooling. From Vernon (2000b, fig. 92). Plane-polarized light; base of photo 3.5 mm.

Other possible products of diagenesis are pyrite framboids, which are rounded, porous, raspberry-shaped (hence the name) structures that occur in organic-rich sedimentary rocks of Precambrian to Recent age (see, for example, Chen, 1978). They commonly range from 1 to 20 μm in diameter. Many of these framboids in sedimentary rocks have been inferred to be of early diagenetic origin (see, for example, Love & Amstutz, 1966); this interpretation is supported by the observation of pyrite framboids in recent sediment at the water–sediment interface, under reducing conditions (Park, 1969). Some have suggested that they are fossils, whereas others have favoured biogenic deposition. Close relationships in some rocks between framboidal and colloform (colloid-like) pyrite suggest chemical precipitation (Chen, 1978), and both structures have been produced by inorganic laboratory synthesis (Berner, 1969). SEM and TEM investigations reveal that many pyrite aggregates in sedimentary rocks contain structures indicative of microbial remains, including some aggregates superficially similar to, though different in detail from, typical pyrite framboids (Schieber, 2002).

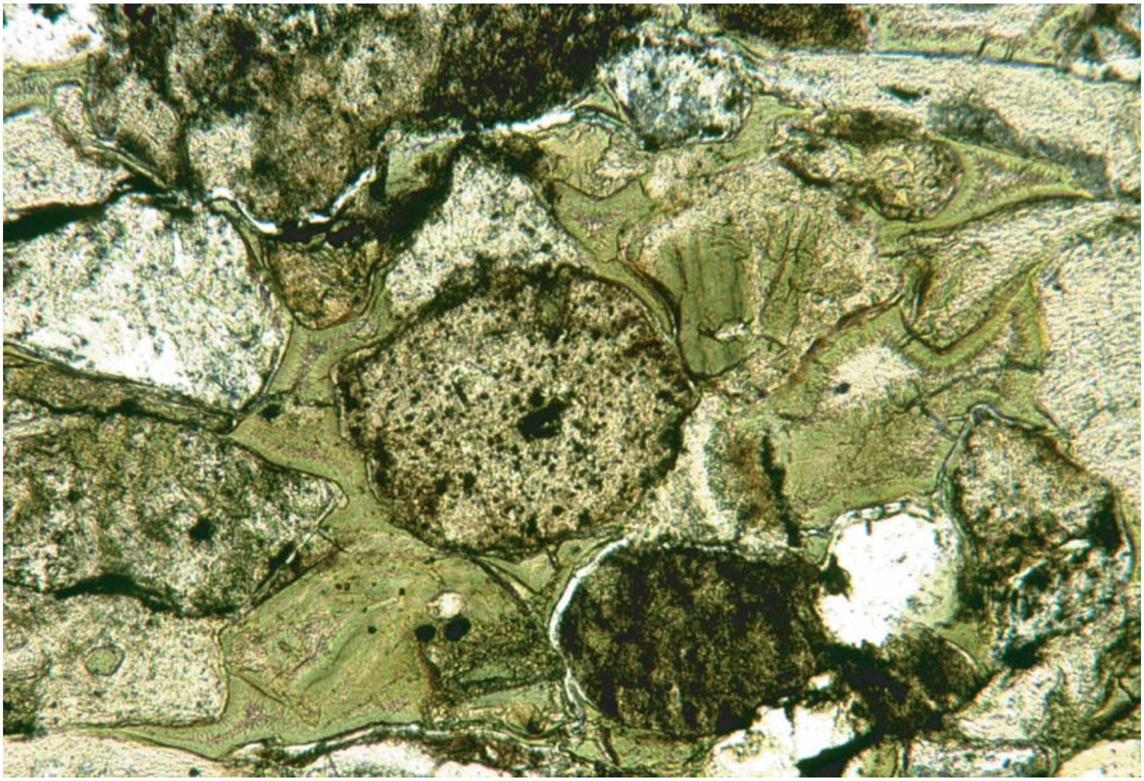


Fig. 2.18: Crystallization of chlorite as fibrous aggregates from the edges towards the centres of glass shards (axiolitic microstructure) in a tuff from Minchinbury, west of Sydney, New South Wales, Australia. Plane-polarized light; base of photo 1.3 mm.

2.3 Pyroclastic sedimentary rocks

Pyroclastic sediment is entirely of local volcanic provenance, and is deposited during explosive volcanic eruptions of predominantly felsic magma (Fig. 2.14), although intermediate and mafic magma may also produce pyroclastic rocks (Fig. 2.15). During these explosive eruptions, the lava solidifies and shatters into fragments, which may be scattered over large areas. The main reason for the explosive eruptions is that felsic magma is relatively rich in water (up to 10%). As the magma rises, cools and precipitates minerals (most of which don't contain water), the concentration of water in the melt increases, and begins to form bubbles of gas (superheated steam). The gas builds up pressure inside the magma trapped just below volcanoes, until the covering rocks fracture. The release of pressure allows the steam to escape suddenly.

During an explosive eruption, the escaping gas boils off, leaving the shapes of many small bubbles in the viscous lava, which cools and solidifies very rapidly, to form a glass froth called *pumice* (Fig. 3.95). The bubble holes may remain

A

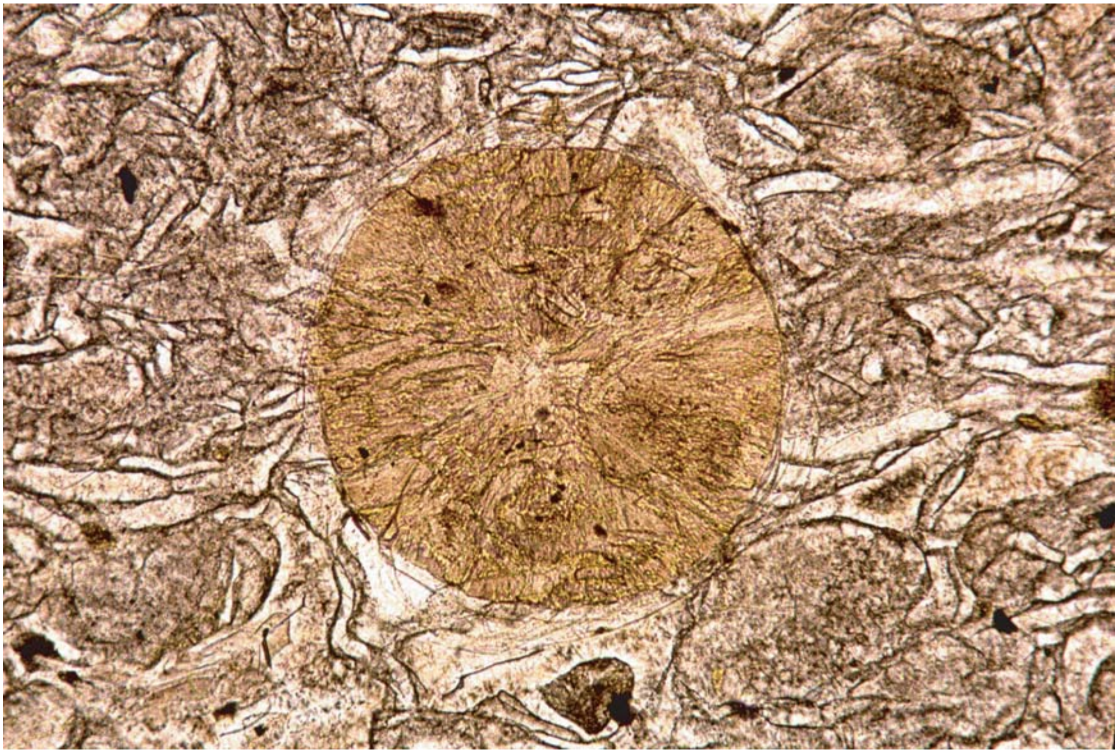


Fig. 2.19: (A) Spherulite (composed largely of fibres of alkali feldspar) that has replaced glass, preserving the shapes of the distorted shards, in an ignimbrite from the Hunter Valley, New South Wales, Australia. Presumably the spherulite fibres were able to grow at the same rate in all directions because all the fragments are of the same glass composition. The outlines of the shards are preserved probably because of minute inclusions or alteration products at their edges that were not needed by the growing feldspar fibres. Plane-polarized light; base of photo 1.5 mm. (B) Same field of view in crossed polars, showing the optical isotropism of the glass, as well as an extinction cross (Section 3.5.5) in the spherulite, caused by the optical extinction of feldspar fibres (each with one of its principal optical vibration directions parallel or nearly parallel to its length) that are approximately parallel to the vibration directions of the polarizer and analyser.

empty, but when hot watery solutions percolate along small cracks in the pumice for thousands or even millions of years, the holes may become filled with minerals precipitated from the solutions (Fig. 2.14). Filled vesicles are called amygdales or amygdules (Section 3.14.2).

During the eruption, the gas continues to escape so fast that it breaks the pumice into large lumps and small, sharp glass pieces or shards (Fig. 2.14). Crystals in the pumice that had crystallized from the magma when it was trapped below the surface and cooling slowly are also fragmented (Fig. 2.14). The high

B

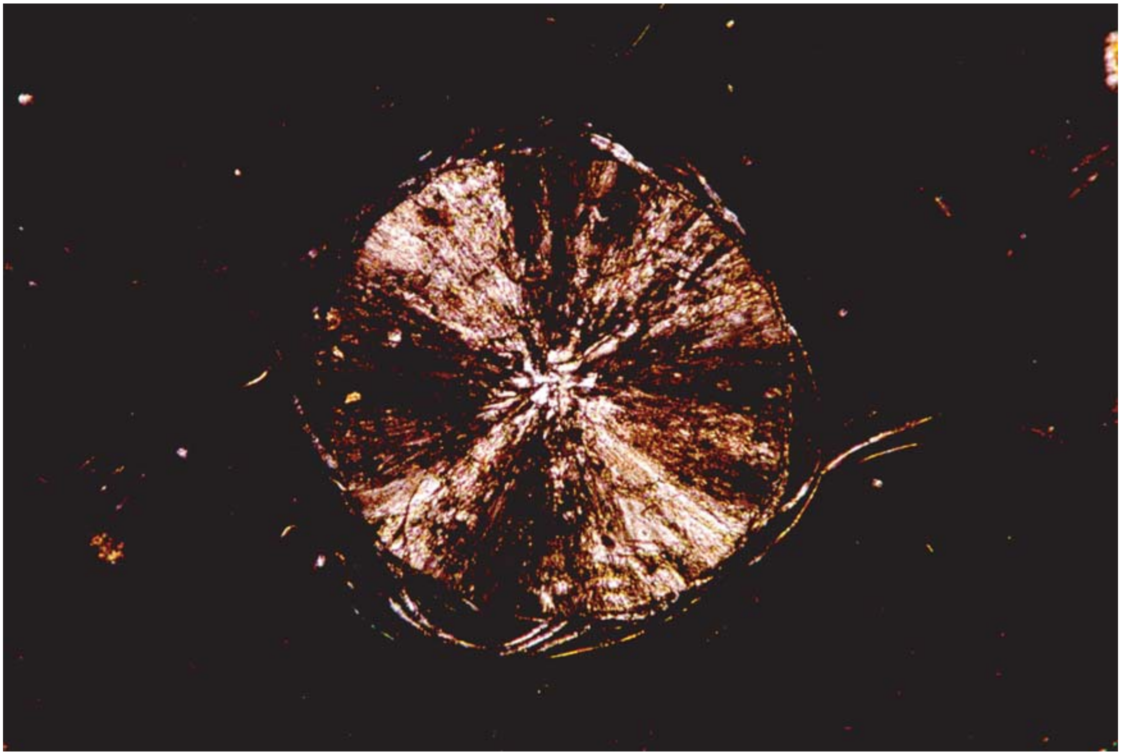


Fig. 2.19: (*cont.*)

gas pressure forces the mixture of glass fragments, pumice fragments, crystal fragments and gas high into the atmosphere. The glass and crystal fragments are mainly of sand size (Section 2.2.2), and are called volcanic ash. The falling ash, together with larger pumice fragments, forms a fragmental deposit called *ash-fall tuff*, in which the sharp-cornered, curved shapes of glass fragments can be seen with the microscope (Fig. 2.14). Some of the curved edges of these fragments may be the typical curved (conchoidal) fractures that glass produces when it breaks, but most of the Y-shaped curved surfaces probably represent glass walls at the junction of bubbles in the original pumice (see, for example, Ross & Smith, 1961), as suggested by Figs. 2.14 and 2.15. Ash-fall tuffs may show all stages from relatively large pumice or highly vesicular glass fragments, some of which have been stretched during the eruption (Fig. 2.16), through small chips in which only one or two bubble shapes are present, to individual shards (Fig. 2.14).

The high gas pressure may also blow out the side of the volcano, forcing the very hot mixture of glass fragments, pumice fragments, crystal fragments and gas to rush down the slopes at speeds of over 100 km h^{-1}. This is a fragmental flow, not a lava flow, and is called an *ash flow*. When the flow slows down, it deposits the ash as a thick deposit (ash-flow tuff, ignimbrite).

A

Fig. 2.20: (A) Volcanic sandstone (probably a reworked tuff) consisting of abundant clasts of former feldspar phenocrysts, many of which are euhedral, together with volcanic rock fragments and a chlorite-rich matrix showing shapes suggestive of former glass fragments. Plane-polarized light; base of photo 2.8 mm. (B) Same field of view in crossed polars, showing the almost optically isotropic matrix, which is due probably to the very fine-grained chlorite, not glass, although it may have been mainly glass originally.

Many ash flows are so hot when they come to rest that the still soft glass and pumice fragments become compressed and squashed together (Fig. 2.17). The flattened pumice fragments acquire distinctive lenticular shapes and are called fiamme, which is the Italian word for flames (the singular being fiamma). If the glass and pumice fragments become fused together by the heat and pressure, the rock is appropriately called a welded tuff. The formerly sharp glass fragments become rounded and squashed in the welding process (Fig. 2.17). Extreme flattening and slight flow may almost obscure the original clastic structure (Ross & Smith, 1961).

The glass and pumice fragments tend to crystallize (devitrify) with time, the glass being replaced by fine-grained crystal aggregates. Acicular (needle-like) crystals commonly grow perpendicular to the walls of the glass shards (a variety of axiolitic structure), as shown for a more mafic tuff in Fig. 2.18, and spherulites

B

Fig. 2.20: (*cont.*)

(Section 3.5.5) may also grow, preserving the shapes of the glass fragments (Fig. 2.19).

After initial accumulation, some reworking and redeposition of the pyroclastic material may occur, but provided this is not enough to change the shapes of the original particles (especially glass shards) or mix it with other sediment, the resulting sediment can still be called ‘pyroclastic’. Otherwise, the sediment becomes epiclastic, and the more general term ‘volcaniclastic’ is appropriate. The rock shown in Fig. 2.20 may well be a reworked tuff, as it is clast-rich, but has residual glass shapes in the chlorite matrix.

2.4 Organic and bioclastic sedimentary rocks

Rocks composed mainly of calcite ($CaCO_3$) are called *limestones*. The calcite can be deposited directly from seawater to form *pelagic limestone* (Fig. 2.21), but usually is deposited in the shells of marine animals. In many skeletons and shells, the $CaCO_3$ is precipitated as aragonite (the orthorhombic polymorph), but it is less stable than calcite (the trigonal polymorph), to which it eventually changes, and so it is not found in old limestones. Fossil-rich limestones (Fig. 2.22) may be

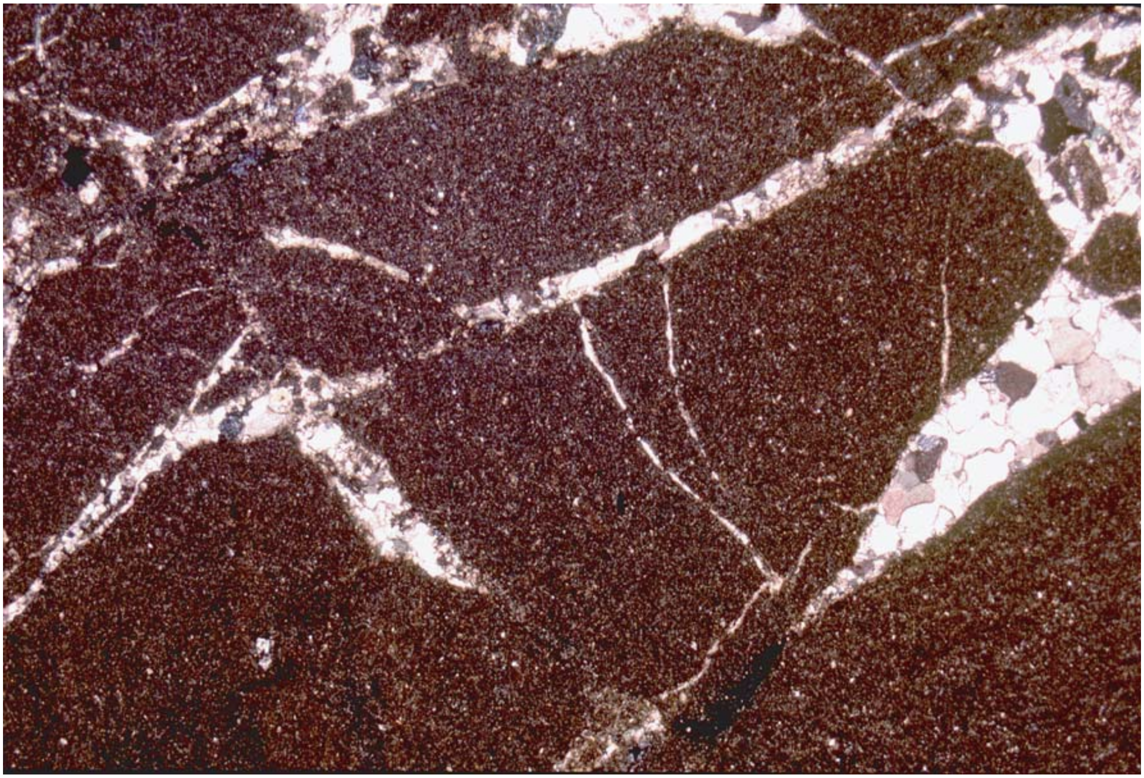


Fig. 2.21: Extremely fine-grained pelagic limestone (carbonate deposited directly from seawater) with later calcite veins. Crossed polars; base of photo 5.2 mm.

autochthonous, the skeletons remaining in the position of life and death without transport (e.g. in coral or algal reefs), or *allochthonous* (e.g. fragments eroded from reefs). In allochthonous limestones (*bioclastic limestones, calcarenites*), the skeletons or shells may be complete or broken into fragments (Fig. 2.22). Later the shells and fragments are cemented together, generally with more calcite. Fossiliferous limestones are named after the predominant fossil present, e.g. 'crinoidal limestone', 'coral limestone', or 'algal limestone'. Microstructural details of shells can be seen clearly in thin section, the calcite or aragonite crystals generally being fibrous and aligned perpendicular or approximately parallel to the walls of the shells. Cavities in shells are commonly filled with relatively coarse-grained mosaics of authigenic calcite.

Calcareous rocks composed of dolomite, $CaMg(CO_3)_2$, are called *dolomites* or *dolostones*, and rocks with both calcite and dolomite are called *dolomitic limestones* or *calc-dolomites*. Generally some ankerite, $CaFe(CO_3)_2$, component is present in the dolomite. Dolomite can be distinguished from calcite in thin section by staining techniques. Because dolomite is not precipitated by organisms, most dolomite is probably formed by the replacement of calcite ('dolomitization').

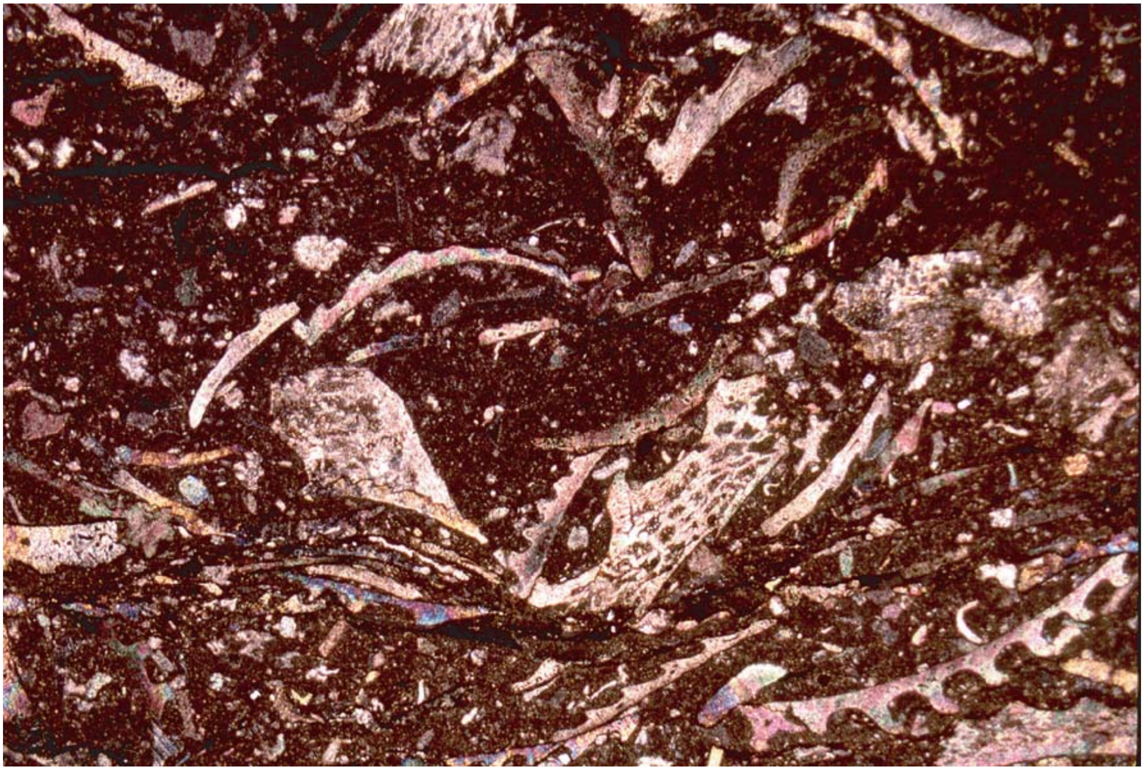


Fig. 2.22: Fossil limestone consisting mainly of fragments of mollusc shells in very fine-grained calcite mud. Crossed polars; base of photo 5.2 mm.

Some dolomite may be precipitated in highly saline lake and sea waters (e.g. in evaporite successions; Section 2.5), although replacement of earlier deposited calcite or aragonite is commonly a possible alternative interpretation. Much evidence of the replacement of calcite fossils by dolomite has been described, but dolomite filling cavities and veins is probably primary. Many dolomites consist mainly of polygonal aggregates or mosaics, without much microstructural evidence of their original nature.

Detrital minerals may also be mixed with the carbonate (Fig. 2.23), especially quartz and clay minerals, but also non-carbonate fossils, such as radiolarians, sponge spicules and diatoms, and pyroclastic fragments (Section 2.3). Limestones with high proportions of detrital material are called 'impure limestones', and at least 50% carbonate must be present for the term 'limestone' to apply. Non-carbonate authigenic minerals in limestones include quartz, chalcedony, gypsum, anhydrite, glauconite and pyrite. Very clay-rich carbonate rocks are called *marls*. Many limestones are very fine-grained, and form as chemically precipitated lime mud (Fig. 2.21), commonly mixed with introduced clay and silt.

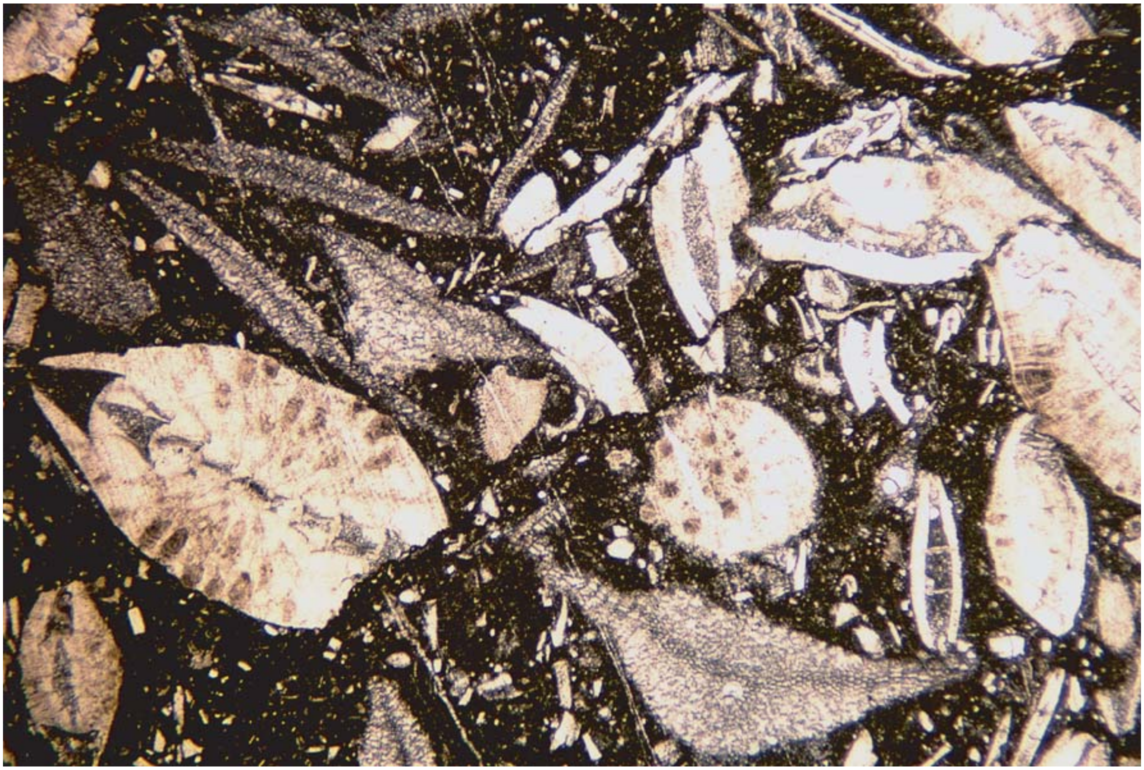


Fig. 2.23: Impure limestone (marl) from the Eocene of the Southern Pyrenees, Spain, consisting of fossil skeletons (mainly nummulites) in a clay–silt matrix. Many skeletons are truncated by dark seams (stylolites) formed by pressure-solution (Section 5.9.2). Sample by courtesy of David Durney. Plane-polarized light; base of photo 11 mm.

Impure (quartz and clay-bearing) limestones (Fig. 2.23) and dolomites produce spectacular metamorphic mineral assemblages when heated (Chapter 4).

Although limestones and dolomites consist mainly of authigenic carbonate, they have very variable and complex microstructures, owing to other contributing processes, such as fragmentation and detrital deposition of skeletons, addition of external detrital material, diagenetic recrystallization and organic accretion (e.g. the formation of ooids; Section 2.5).

2.5 Chemical sedimentary rocks

Chemical sediments are either (1) deposited directly from supersaturated aqueous solution, owing to chemical reactions or evaporation (e.g. in a land-locked lake), forming *evaporites*, such as halite-rock ('rock-salt') and gypsum-rock, or (2) formed by replacement of pre-existing sedimentary material (e.g. some cherts).

Fig. 2.24: Oolitic limestone, Lake Keepit area, north-western New South Wales, Australia. The limestone consists mainly of fossil shell and crinoid (sea-lily) fragments that probably accumulated in a lagoon near a coral reef being eroded by wave action. The dark rims of the ooids were formed by very fine-grained algal mud sticking to them as they were gently washed about in the lagoon. The shapes of the ooids are determined by the shapes of the original clasts. The fragments and ooids are cemented by calcite. From Vernon (2000b, fig. 111). Plane-polarized light; base of photo 1.8 cm.

Material formed in these ways is said to be *orthochemical*, but if it is moved and reorganized into new shapes by chemical, physical or biological processes within the depositional basin, it is said to be *allochemical*. Examples of allochemical sediment include *ooids* (ooliths), which are spherical accretions of chemically precipitated aggregates, and *faecal pellets*, which are rounded particles of carbonate mud produced by sediment-ingesting animals.

Limestones rich in ooid*s* are called *oolitic limestones* or *calcareous oolites* (Fig. 2.24). The ooids are spherical or ellipsoidal, and consist of calcite or aragonite deposited and accreted by chemical precipitation and/or physical plastering of carbonate mud in layers around nuclei, such as fossil fragments or sand clasts. Many ooids have shapes determined by the shapes of their nuclei (Fig. 2.24). They

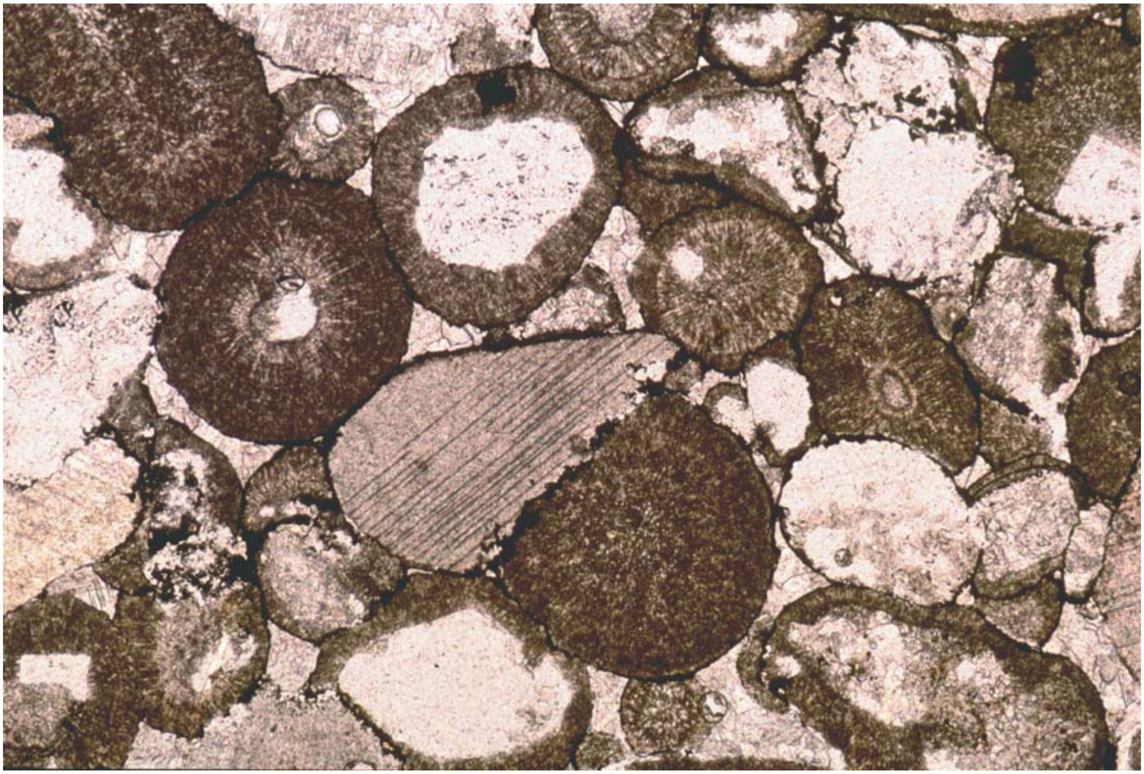


Fig. 2.25: Oolitic limestone, in which many of the ooids have a radial structure, cemented by calcite. Some calcite clasts are also present. Most of the ooids and the clasts meet along irregular solution surfaces, and some have been markedly truncated by stylolitic solution seams (Section 5.9.2). Plane-polarized light; base of photo 2.8 mm.

are generally well sorted. Some show radial microstructures (Fig. 2.25), which may result from recrystallization of ooids with concentric layering (Williams *et al.*, 1954, p. 344; Boggs, 1992, pp. 428–29). Chlorite may also occur as ooids (Fig. 2.26).

In contrast to the separate particles of clastic rocks, chemical precipitates typically form aggregates of interlocking grains and crystals (Fig. 2.27). Crystals precipitating from a supersaturated solution commonly settle to the bottom of a sea or lake, where they continue to grow until they impinge, forming laminated microstructures or mosaics of grains. Mosaics may also be formed when fine-grained, chemically precipitated calcite or aragonite is replaced by coarser-grained dolomite (Section 2.4), although rhombic crystals of dolomite in a calcite matrix may result when the replacement is incomplete. Mosaics are commonly formed when anhydrite ($CaSO_4$) is replaced by gypsum ($CaSO_4.2H_2O$), although pseudomorphs of gypsum after anhydrite may also be present. Replacement may

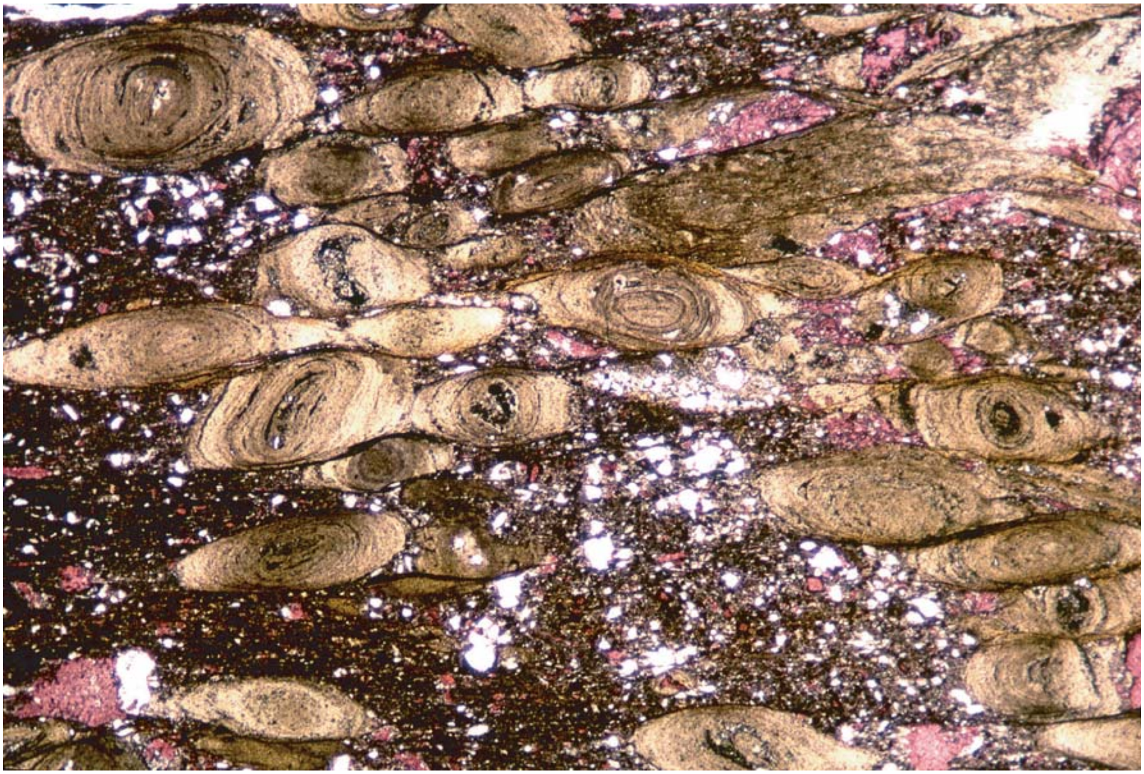


Fig. 2.26: Chlorite ooids in a calcareous shale from the Swiss Alps. The ooids have been distorted and elongated by deformation, and many have had their shapes truncated by solution seams (Section 5.9.2). The rock has been stained to reveal calcite (pink dye). Sample by courtesy of David Durney. Plane-polarized light; base of photo 4 mm.

also produce interpenetrating, radiating and even spherulitic aggregates of elongate crystals (Carozzi, 1960), as shown in Fig. 2.27.

The most common minerals in chemical sediments are carbonates, mainly calcite and dolomite (70–85%), quartz (10–15%), and mineral ‘salts’ (mainly gypsum, anhydrite, halite and sylvite). Gypsum in evaporites may be formed by replacement of anhydrite and *vice versa*. Many other mineralogical and microstructural diagenetic changes are typical of buried evaporites saturated with concentrated brine, because of the solubility and reactivity of the minerals involved. Some anhydrite and gypsum rocks may contain introduced sand, silt, clay and carbonate minerals.

Chemical processes can also contribute to residual (pedogenic) sediments, by precipitation of clay minerals, silica minerals, iron oxide minerals and carbonate minerals in small spherical accretions (*nodules*, *pisoliths*). Resulting rocks include pisolitic bauxite (composed of hydrated aluminium oxide minerals, as

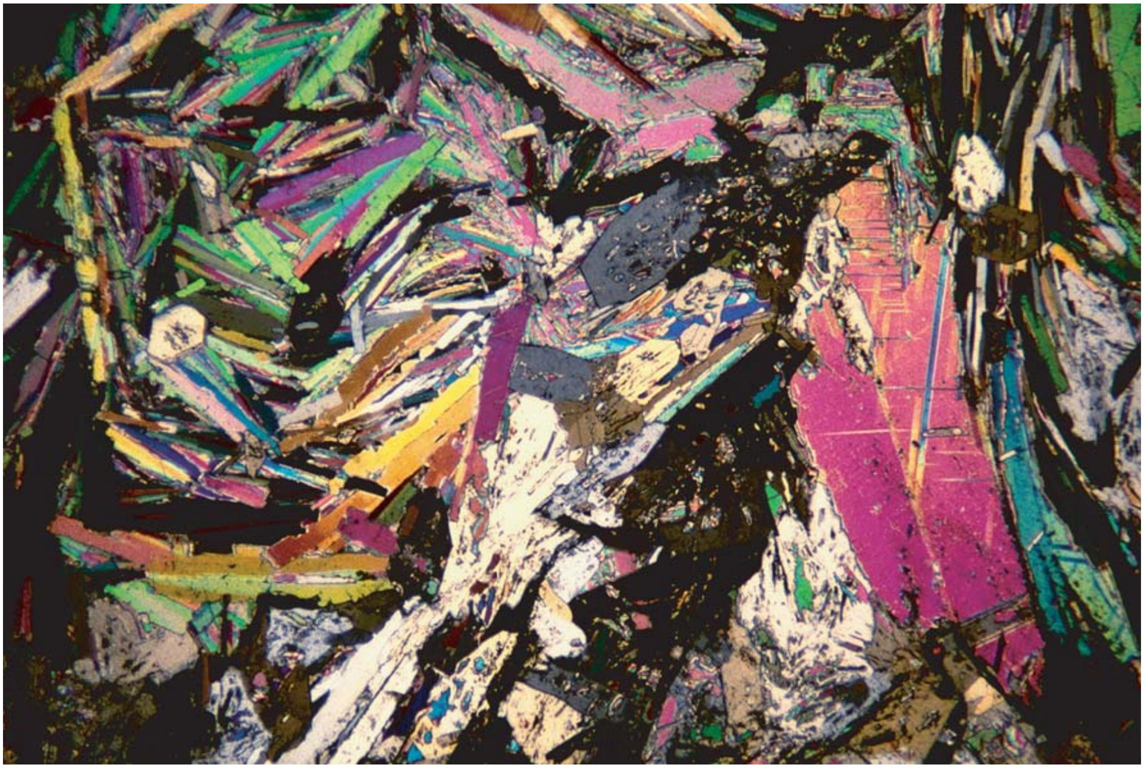


Fig. 2.27: Anhydrite, formed by displacive and replacive growth in clay–carbonate mud and calcareous rudite, Munta-1 exploration oil well, Tanana Formation, Ungoolya Group, Officer Basin, South Australia (Akouri *et al.*, 2000). Specimen by courtesy of Pat Conaghan. Crossed polars; base of photo 11 mm.

well as iron oxide and hydrated iron oxide minerals) and pisolitic caliche (carbonate minerals). The process involves fluctuating groundwater levels, leading to concentration and precipitation of dissolved chemical components by surface evaporation.

Other essentially chemically deposited sedimentary rocks are siliceous (cherts), ferruginous (ironstones) and phosphatic (phosphorites), with a wide variety of microstructures (including nodular, oolitic and finely granular) complicated by mixing with various other types of sedimentary material, as discussed by Pettijohn (1949) and Williams *et al*. (1954), among others.

Chapter 3
Microstructures of igneous rocks

3.1 Introduction

This chapter is concerned with the shapes of crystals and aggregates formed when melted rocks and glass crystallize, solid rocks melt, and melts boil. These processes are broadly grouped together as 'magmatic (igneous) processes'.

As discussed in introductory geology courses, partial melting of solid rock to form *magma* occurs in Earth's mantle, producing mainly *mafic* (basaltic, silica-poor) magma, and also in the deeper parts of Earth's crust, producing mainly *felsic* (granitic, silica-rich) magma. Magma may reach the surface as volcanic rocks or solidify in the crust as intrusive rocks. Magma bodies may encounter other magma bodies and mix or mingle with them. Microstructures may preserve evidence of these and other processes, such as magmatic flow and conditions of crystallization, and in this chapter I present this evidence and discuss its reliability.

When a liquid becomes *supersaturated* with a dissolved chemical component (i.e. the dissolved component reaches a concentration at which the liquid can hold no more of the component in solution), a new phase theoretically is produced. A phase is a chemically and physically homogeneous part of a *system* (e.g. a body of magma) that is bound by an interface with other phases (e.g. the melt phase and other crystal phases in a magma). Generally, the new phase is a crystalline solid (which in rocks we call a mineral), and the phase change is called *crystallization* or *freezing*. However, if a magma becomes supersaturated with water or another volatile component, the new phase is a gas, and the phase change is called *boiling* or *vesiculation* (Section 3.14). Although even slight supersaturation should theoretically produce a phase change, this is generally delayed until a higher degree of supersaturation occurs, as discussed in Section 3.3.1.

Magmas freeze (solidify by crystallization) as they cool, because the solubility of chemical components in the melt decreases at lower temperatures. Because natural silicate melts are complex solutions, various chemical components saturate the melt at different temperatures, so that some minerals crystallize earlier than others. This leads to an *order of crystallization*, which is determined by melting experiments on the magma composition concerned, but which generally cannot be determined by inspection with the microscope (Section 3.6). Magmas may also crystallize in response to sudden loss of water (Sections 3.4.4, 3.4.7) or changes in pressure.

The complexity of magmas and their individual histories — potentially involving changes in temperature, pressure and water content, as well as mixing with other magmas (Section 3.10.3) and contamination with foreign crystals (Section 3.10.2) — implies that most of the following explanations are only broadly applicable. Each body of magma requires individual, detailed investigation to work out its crystallization history.

3.2 Structure of silicate melts and glasses

The atomic structure of silicate melts determines several major features of igneous rocks, namely: (1) the rate of flow of magma, (2) the explosiveness of volcanic eruptions, (3) the rate of crystallization (and hence the grainsize), and (4) the extent to which glass is formed during rapid cooling (Section 3.5.6). Therefore, at least a basic idea of melt structure helps to understand many magmatic processes.

Theoretical and X-ray diffraction studies have indicated that the atoms in silicate melts (the liquid parts of natural magmas) are bound together in extensive networks (polymerized) to varying degrees, depending on the composition of the melt (see, for example, Hess, 1980, 1995; Mysen, 1988, 1999, 1980a,b, 1981a,b). This polymerization applies especially to O–Si and O–Al bonding. Silicate melts consist essentially of linkages of Si^{4+} and Al^{3+} ions with four O^{2-} ions. In view of the large amount of SiO_2 in most igneous rocks, the degree of polymerization depends largely on the O : Si ratio of the melt. Melts with higher O : Si ratios (as in mafic or basaltic magmas) tend to consist mainly of separate $[SiO_4]^{4-}$ groups (tetrahedra), without much polymerization, although X–ray evidence of isolated silicate chains has been obtained for melts of pyroxene composition. Such relatively silica-poor melts tend to be relatively fluid, as evidenced by extensive basalt lava flows, and gas escapes relatively easily from them, as evidenced by the relatively weakly explosive nature of basalt volcanic eruptions.

However, melts with lower O : Si ratios (as in felsic or granitic magmas) have more Si–O bonding, so that more continuous silicate networks are formed, increasing the viscosity (stiffness) of the melt. The result is that felsic magmas, being viscous, flow for relatively short distances or plug the vent of the volcano. Moreover, because gas escapes much less readily from viscous magmas, felsic volcanic eruptions are commonly very explosive, producing fragmental (pyroclastic) rocks (Section 2.3).

Melts of pure SiO_2 composition (i.e. with the lowest O : Si ratios of all) are strongly polymerized and show diffuse X-ray peaks similar to the main peaks of cristobalite, a high-temperature polymorph of SiO_2 (Hicks, 1967). Although the degree of linking of their $[SiO_4]^{4-}$ tetrahedra is very high, the arrangement of the tetrahedra is not as regular as in crystalline silica. When a pure silica melt changes to a glass by very rapid cooling (Section 3.5.6), very little rearrangement of the Si–O network is required (Brückner, 1983).

3.2.1 Effect of melt structure on crystallization

Poorly polymerized (Si-poor) melts crystallize more readily than polymerized (Si-rich) melts (see, for example, Kirkpatrick, 1983). In fact, experimental quenching of liquids of ultramafic composition, without the formation of at least some small crystals, is very difficult or impossible. Melts of basalt composition can be cooled quickly without crystallizing, as evidenced by natural occurrences of basaltic glass (*tachylite* or *tachylyte*). However, this is much less common than natural glass (Section 3.5.6) of SiO_2-rich composition (*obsidian*), and industrial glasses generally are SiO_2-rich.

Although glasses have the rigidity of solids, they have atom arrangements more like those of liquids. In this sense, they are supercooled (undercooled) liquids, having a high degree of polymerization, which accounts for their rigidity. The polymerized structure of Si-rich melts enables them to pass relatively easily from a liquid to a glassy state on rapid cooling, despite the fact that the crystalline state has a lower free energy than glass at low temperatures. In fact, glass is metastable, with respect to crystals of equivalent composition. Because of this, glass tends to change into the lower-energy crystalline state with time. However, this involves breaking strong Si–O and Al–O bonds, and can take place only if some other factor permits the necessary energy barriers to be surmounted. Possible factors include reheating of the glass (e.g. following burial of volcanic glass) and water, which tends to break strong Si–O bonds and which can be added to volcanic glass along cracks, after burial, especially in pyroclastic rocks (Section 2.3).

3.2.2 Effect of volatiles on viscosity and crystallization of magmas

Volatile components in the melt, especially water and fluorine, cause disruption of O–Si bonds and a reduction in polymerization. This is because ions such as $(OH)^-$ and F^- substitute for O^{2-} ions. Whereas each O^{2-} ion links to two Si^{4+} ions, $(OH)^-$ or F^- ions can link to only one. Thus, $(OH)^-$ and F^- ions break polymerized Si–O groups. This decreases the viscosity of the melt, because the remaining Si–O groupings become independent and so can move more easily. Therefore, volatile components, especially water (present as hydroxyl ions) can have a large effect on the viscosity and flow of silicate magmas. For example, although water dissolved in a felsic melt reduces the viscosity and so helps the magma to move through the crust, once the water escapes as gas (steam), owing to reduction in pressure at or near Earth's surface, the melt becomes very viscous, with the result that felsic lava flows with great difficulty. In fact, the rapid escape as steam of the relatively large amounts of water in felsic magmas contributes strongly to the explosive nature of most felsic volcanic eruptions.

Volatiles also decrease the rate of formation of crystals. Because the O–Si networks are disrupted, the atoms and atomic groupings can less easily coalesce to form crystal nuclei (Section 3.3.1). This applies especially to felsic melts, which are richer in water than mafic melts. The effect is particularly pronounced in felsic melts with unusually abundant water and other volatile components (e.g. F, B, Li), which commonly crystallize as rocks with very large grainsizes, called *pegmatites* (Section 3.4.6).

3.3 Crystallization (freezing) of magma: nucleation and growth

3.3.1 Nucleation

Crystallization of liquids does not take place by the co-operative action of all atoms present. Instead, local thermal, compositional or deformational fluctuations enable small aggregations of atoms of the new crystal to occur with a minimum expenditure of energy. These aggregates are called *nuclei*, which probably consist of 10 to 1000 atoms (Lasaga, 1998). So far, we cannot observe nuclei forming, and so the nucleation step in the crystallization of natural liquids is poorly understood.

Because most liquids have no long-range order of their atoms, and crystals do, the change from liquid to solid is generally a discontinuous process. The relatively sudden change of atomic structure at the liquid–solid interface produces a definite boundary with a specific *interfacial energy* (γ) per unit area. This energy is due to the fact that atoms at the surface of the crystal are not as well bound to the lattice (the array of linked atoms in the crystal) as those in the interior, and so have a higher energy.

Crystallization of liquids (e.g. melts) can occur only if a decrease in the total free energy of the system (G_t) results. At temperatures below the equilibrium freezing temperature, the crystal has a lower chemical free energy per volume (G_v) than the liquid. However, the formation of a nucleus (i.e. a *volume* of new crystal), which decreases the chemical free energy of the volume of the system converted from liquid to crystal (G_v), is accompanied by the formation of an *area* of interface, which increases the interfacial free energy (γ). Therefore, the free energy change (ΔG_t) involved in nucleation is:

$$\Delta G_t = -(4/3)\pi r^3 \Delta G_v + 4\pi r^2 \Delta\gamma,$$

where ΔG_v is the difference in free energy between the solid and the liquid (the volume free energy) and r is the radius of the embryonic nucleus (assumed to be spherical). Spontaneous nucleation can occur only if ΔG_t is negative. As shown in Fig. 3.1, a nucleus of less than a critical radius (r^*) is unstable and can dissolve, because its area is too large relative to its volume, and so $4\pi r^2 \Delta\gamma$ is

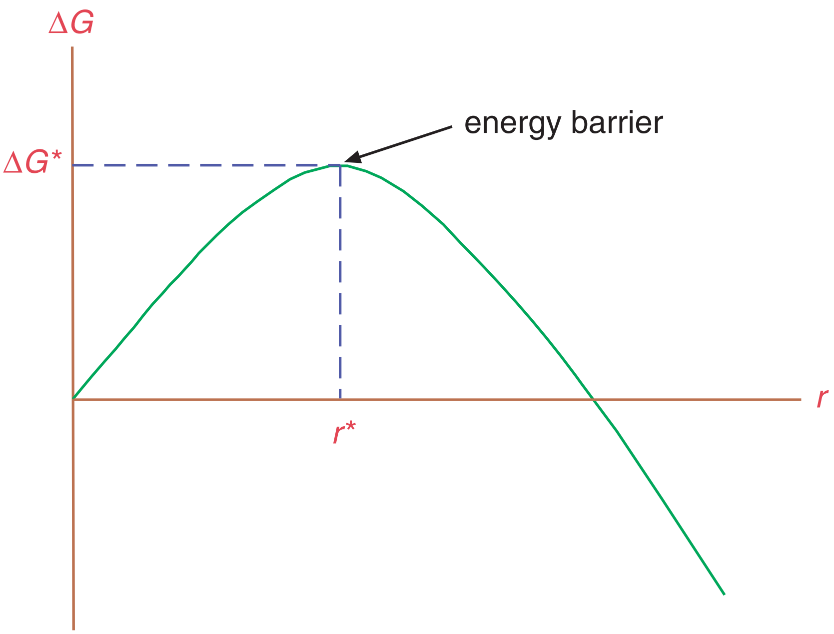


Fig. 3.1: Free energy of nucleation (ΔG) plotted against radius of nuclei (r). Stable nuclei must have radii greater than the critical radius (r^*), having surmounted the energy barrier (ΔG^*).

larger than $(4/3)\pi r^3 \Delta G_V$. Nuclei of radius larger than r^* are stable, and can grow into crystals. These nuclei have been called 'critical clusters' (see, for example, Walton, 1965). Smaller atomic clusters tend to dissolve in the melt.

The *energy barrier* to nucleation (Fig. 3.1) means that a certain amount of *undercooling* or *supercooling* (depression of temperature below the equilibrium freezing temperature) must occur before crystallization can take place (Fig. 3.2). In fact, all changes that involve a nucleation step (including solid-state metamorphic reactions, discussed in Chapter 4) involve some *overstepping* of reaction boundaries (e.g. with regard to temperature and/or pressure).

Nucleation is *homogeneous* if it occurs as a result of random thermal fluctuations in the liquid and *heterogeneous* if assisted by solid material, on which the new crystal can nucleate much more easily. The amount of undercooling required for heterogeneous nucleation is much less than for homogeneous nucleation. The solid material (which could be minute crystals, phenocrysts, foreign rock fragments, or the walls of a vein or magma chamber) forms a crystalline 'substrate', as shown in Fig. 3.3, which also shows the interfacial energies involved. A certain proportion of the area of the nucleus is a nucleus–substrate interface (γ_{NS}), the remainder being a nucleus–liquid interface (γ_{NL}), as in homogeneous nucleation. Heterogeneous nucleation is favoured because γ_{NS} is typically much less than γ_{NL} (especially where relatively good atomic registry occurs across the