Petrogenesis of Metamorphic Rocks

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Chapter 2
Metamorphic Rocks

This chapter deals with the descriptive characterization of metamorphic rocks. Metamorphic rocks are derived from other rocks of igneous, sedimentary or metamorphic origin. The chemical composition of this primary material (=protolith) determines the chemical and mineralogical composition of metamorphic rocks to a large degree. The compositional variation found in the primary material of metamorphic rocks is reviewed in Sect. 2.1.

The structure of metamorphic rocks is often inherited from the precursor material. In low-grade metasedimentary rocks, for example, the sedimentary bedding and typical structures of sedimentary rocks such as cross-bedding and graded bedding may be preserved. Ophitic structure, characteristic of basaltic lava, may be preserved in mafic metamorphic rocks. Very coarse-grained structures of igneous origin can occasionally be found even in high-grade metamorphic rocks. Most metamorphic rocks, however, exhibit structures that are of distinct metamorphic origin. The structures typically result from a combination of deformation and recrystallization. Deformation normally accompanies large-scale tectonic processes causing metamorphism. Some descriptive terms of metamorphic structures are defined in Sect. 2.2. Classification principles and nomenclature of metamorphic rocks are explained in Sect. 2.3.

Large scale tectono-thermal processes move rocks along unique paths in pressure–temperature–time (P–T–t) space. Rocks may undergo continuous recrystallization and new minerals may replace old ones in a complex succession. Earlier-formed minerals and groups of minerals often experience metastable survival because of unfavorable reaction kinetics. This happens particularly if an aqueous fluid phase is absent. Metamorphism may proceed episodically. The study of metamorphic rocks aims at the correct identification of the group of minerals that may have coexisted in chemical equilibrium at one stage during the evolutionary history of the metamorphic rock. This group of minerals is the **mineral assemblage**. The total succession of mineral assemblages preserved in a metastable state in the structure of a metamorphic rock is designated **mineral paragenesis**. Some aspects regarding the mineral assemblage and the mineral paragenesis are discussed in Sect. 2.4.

Discussion and analysis of phase relationships in metamorphic rocks is greatly facilitated by the use of composition phase diagrams. Their construction is explained

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in the last section of this chapter (Sect. 2.5) on graphical representation of mineral assemblages. However, the quantitative computation of equilibrium composition phase diagrams is not discussed in this book.

2.1 Primary Material of Metamorphic Rocks

All metamorphic rock-forming processes make rocks from other rocks. The precursor rock or **protolith** determines many attributes of the new metamorphic rock.

Metamorphism results from the addition (or removal) of heat and material to discrete volumes of the crust or mantle by tectonic or igneous processes. Metamorphism, therefore, may affect all possible types of rock present in the Earth's crust or mantle. Protoliths of metamorphic rocks comprises rocks of all possible chemical compositions and include the entire range of sedimentary, igneous and metamorphic rocks.

Metamorphic processes tend to change the original composition of the protolith. Addition of heat to rocks typically results in the release of volatiles (H_2O, CO_2, etc.) that are stored in hydrous minerals (e.g. clay, micas, amphiboles), carbonates and other minerals containing volatile components. Therefore, many metamorphic rocks are typically depleted in volatiles relative to their protoliths. Metamorphism that releases only volatiles from the protolith is, somewhat illogically, termed isochemical. On a volatile-free basis, the chemical composition of protolith and product rock is identical in isochemical metamorphism. In truly isochemical metamorphism, protolith and product rocks are of identical composition including the volatile content. Recrystallization of sedimentary calcite to coarse grained calcite marble is an example of isochemical metamorphism in a strict sense. It affects the structure of the rock, no new minerals are formed.

Many, if not most, metamorphic processes also change the cation composition of the protolith. This type of metamorphism is termed allochemical metamorphism or metasomatism. The aqueous fluid released by dehydration reactions during metamorphism generally contains dissolved solutes. These are then carried away with the fluid and lost by the rock system. It has been found, for example, that many granulite facies gneisses are systematically depleted in alkalis (Na and K) relative to their amphibolite facies precursor rocks. This can be explained by alkali loss during dehydration. Silica saturation is a general feature of nearly all metamorphic fluids. Pervasive or channelled regional scale flow of silica-saturated dehydration fluids may strongly alter silica-deficient rocks (ultramafic rocks, dolomite marbles) that come in contact with these fluids. Unique metamorphic rock compositions may result from metasomatism on a local or regional scale. Efficient diffusion and infiltration metasomatism requires the presence of a fluid phase. Metasomatism is a process of fluid–rock interaction combined with transfer of material between the systems at elevated temperature and pressure. Fluid–rock interaction processes are dominant in chemical weathering, in sedimentary, diagenetic, groundwater, hydrothermal and other near surface environments, but they play an important role also in the metamorphic domain. In fact, isochemical metamorphism of rocks may be viewed as interaction of rocks with an internally derived fluid that makes dissolution of unstable and precipitation of more stable mineral assemblages possible.

Interaction of rocks with externally derived fluids is referred to as allochemical metamorphism. The volatile composition of the fluid may not in equilibrium with the mineral assemblage of the rock and, consequently, the rock may be altered. Some examples: flushing of rocks with pure H_2O at high-P–T conditions may initiate partial melting, it may form mica and amphibole in pyroxene-bearing rocks, it may induce wollastonite or periclase formation in marbles. CO_2 metasomatism is particularly common in very high-grade rocks. Metasomatism can create rocks of extreme composition which, in turn, may serve as a protolith in subsequent metamorphic cycles. Metasomatic rocks of unusual composition are widespread in regional metamorphic terrains and contact aureoles. However, the total volume of such types of rocks is negligible. Although interesting petrologically, these exotic rocks will not be discussed in Part II where we present a systematic treatment of prograde metamorphism of the most important types of bulk rock compositions.

2.1.1 Chemical Composition of Protoliths of Metamorphic Rocks

The average composition of crust and mantle is listed in Table 2.1. The mantle constitutes the largest volume of rocks on planet Earth. From geophysical and petrophysical evidence and from mantle fragments exposed at the surface, we know that the mantle consists predominantly of ultramafic rocks of the peridotite family. The bulk of the mantle is in a solid state and experiences continuous recrystallization as a result of large-scale convective flow in the sub-lithospheric mantle and tectonic processes in the lithospheric mantle. Therefore, nearly all mantle rocks represent metamorphic rocks. The composition of the mantle (Table 2.1) is representative for the most prominent type of metamorphic rock of this planet. However, mantle rocks can only be transported through the lid of crust to the surface of the Earth by active tectonic or igneous processes. Although outcrops of ultramafic

Table 2.1 Composition of the Earth's crust and mantle (After Carmichael 1989)

	Peridotite mantle	Continental crust	Oceanic crust	Basalt	Tonalite
SiO_2	45.3	60.2	48.6	47.1	61.52
TiO_2	0.2	0.7	1.4	2.3	0.73
Al_2O_3	3.6	15.2	16.5	14.2	16.48
FeO	7.3	6.3	8.5	11.0	5.6
MgO	41.3	3.1	6.8	12.7	2.8
CaO	1.9	5.5	12.3	9.9	5.42
Na_2O	0.2	3.0	2.6	2.2	3.63
K_2O	0.1	2.8	0.4	0.4	2.1
H_2O	<0.1	1.4	1.1	<1.0	1.2
CO_2	<0.1	1.4	1.4	<1.0	0.1

rocks are common and widespread, particularly in orogenic belts, the total volume of ultramafic rocks exposed on continents is small.

Crustal rocks may be divided into rocks from oceanic and continental environments. Characteristic compositions of continental (tonalite) and oceanic crust (basalt) are listed in Table 2.1. It is evident that the average composition of oceanic crust is well represented by an average basalt composition, and the average composition of continental crust can be described by an average tonalite composition. The composition of the continental crust is more heterogeneous than the oceanic crust which is ~99% basaltic. Table 2.2 lists abundances of types of rocks that make up typical crust and are the predominant protoliths of metamorphic rocks. Igneous rocks of mafic composition (basalts, gabbros of Mid Ocean Ridge Basalt [MORB] affinity) form the oceanic crust which covers much larger areas than continental crust, and constitute an important chemical group of metamorphic rocks (greenschist, amphibolite, granulite, eclogite).

Granite and related rocks such as granodiorite and quartz–diorite (typical granite and tonalite compositions given in Table 2.3) dominate the continental crust. They make up the family of metamorphic rocks termed meta-granitoids (= quartzo-feldspathic rocks) and represent 33% of all igneous rocks of the Earth's crust (Table 2.2).

Table 2.2 Abundance of rocks (vol%) in the crust (After Carmichael 1989)

Igneous rocks	64.7		
Sedimentary rocks	7.9		
Metamorphic rocks	27.4		
Igneous rocks (64.7)		*Sedimentary rocks (7.9)*	
Granites	16	Shales	82
Granodiorites/diorites	17	Sandstones, arkoses	12
Syenites	0.6	Limestones	6
Basalts/gabbros	66		
Peridotites/dunites	0.3		

Table 2.3 Chemical composition of sedimentary and igneous rocks (After Carmichael 1989)

	Sandstones, graywackes	Shales (platforms)	Pelites, pelagic clays	Carbonates (platforms)	Tonalite	Granite	Basalt MORB
SiO_2	70.0	50.7	54.9	8.2	61.52	70.11	49.2
TiO_2	0.58	0.78	0.78	–	0.73	0.42	2.03
Al_2O_3	8.2	15.1	16.6	2.2	16.48	14.11	16.09
Fe_2O_3	0.5	4.4	7.7	1.0	–	1.14	2.72
FeO	1.5	2.1	2.0	0.68	5.6	2.62	7.77
MgO	0.9	3.3	3.4	7.7	2.8	0.24	6.44
CaO	4.3	7.2	0.72	40.5	5.42	1.66	10.46
Na_2O	0.58	0.8	1.3	–	3.63	3.03	3.01
K_2O	2.1	3.5	2.7	–	2.1	6.02	0.14
H_2O	3.0	5.0	9.2	–	1.2	0.23	0.70
CO_2	3.9	6.1	–	35.5	0.1		
C	0.26	0.67	–	0.23			

On a global basis, sedimentary rocks are dominated by shales and clays of pelagic and platform (shelf) environments (82% of all sediments, Table 2.2). Compositions of typical shales from continental and oceanic settings are listed in Table 2.3. Pelagic clays represent the typical sediments of the deep oceans and these extremely fine-grained clay-rich sediments are designated as pelites. They form the most important type of metamorphic rock of sedimentary origin (metapelites).

Shales deposited on continental shelfs typically contain carbonate minerals as reflected in the chemical composition of the average analysis given in Table 2.3. Such carbonate-rich shales are usually referred to as marls and their metamorphic equivalents, e.g. calcareous mica-schists are important types of metasediments in orogenic belts.

With reference to Table 2.2, sandstones (e.g. graywacke, arkose) and carbonate rocks (limestone) are the remaining important groups of rocks. The former dominate the sedimentary rocks in continental basins and subduction related trenches, whereas carbonate sequences (including marls) are characteristic of continental shelf (platform) depositional environments. Characteristic compositions of graywacke sandstone and platform carbonate are listed in Table 2.3.

If metamorphic rocks become a protolith in a new cycle of metamorphism, their composition will comprise those of the most common sedimentary and igneous rocks as given in Table 2.3. Thus, all metamorphic rocks can be grouped into seven classes of characteristic bulk rock composition.

2.1.2 Chemical Composition Classes of Metamorphic Rocks and Their Protoliths

The seven classes are arranged according to increasing chemical complexity.

1. *Ultramafic rocks*. Usually mantle-derived, very Mg-rich family of rocks (typical peridotite composition in Table 2.1). Metamorphism of ultramafic rocks produces hydrous and non-hydrous Mg-silicate minerals and is discussed in Chap. 5.
2. *Carbonate rocks*. These are chemical sedimentary rocks modally dominated by carbonate minerals (calcite, dolomite) (Table 2.3). Their metamorphosed equivalents, marble, calc-silicate rocks or metacarbonate, are dominated by Ca–Mg and Ca-silicate minerals in addition to calcite and are described in Chap. 6.
3. *Pelites (shales)*. Pelitic rocks such as shale are the most common type of sedimentary rock. Pelagic clays (true pelites) are poor in calcium compared with shales from continental platforms (Table 2.3). Pelites constitute a separate composition group and their metamorphic equivalents are termed metapelites (metapelitic slate, phyllite, schist and gneiss) characterized by the formation of K–Al and Al-rich silicate minerals as detailed in Chap. 7.

4. *Marls*. Marls are shales containing a significant proportion of carbonate minerals (usually calcite) (Table 2.3), and metamorphic equivalents are dominated by Ca–Al silicate minerals as discussed in Chap. 8.
5. *Mafic rocks*. Metamorphic mafic rocks (e.g. greenstone, mafic schist and gneiss, greenschist, amphibolite) are derived from mafic igneous rocks, mainly basalt (but also andesite) and, of lesser importance, gabbro (Tables 2.1 and 2.3). Based on total volume, basalts are the most important group of volcanic rocks, and metabasalts or metabasites are common in many metamorphic terrains. Metamorphic assemblages in mafic rocks are dominated by Ca–Mg–Fe–Al silicate minerals that define the intensity of metamorphism in the metamorphic facies concept (Chap. 4). Prograde metamorphism of mafic rocks is described in Chap. 9.
6. *Quartzo-feldspathic rocks*. Metamorphic rocks of sedimentary (arkose, sandstone, siltstone, greywacke) or igneous origin (granite, granodiorite, tonalite, monzonite, syenite, etc.) which are modally dominated by quartz and feldspar (Table 2.3) are described in Chap. 10. Gneisses derived from igneous rocks of the granite family may be designated as metagranitoids.
7. *Other bulk compositions*. All remaining compositions of protoliths of metamorphic rocks are of minor and subordinate importance. They include manganese sediments and chert, volcanogenic sediments, ironstones, laterites, evaporites and alkaline igneous rocks, to name only a few.

2.2 The Structure of Metamorphic Rocks

Most metamorphic rocks originate from large-scale tectonic processes (deformation) and associated changes in pressure and temperature. Metamorphism involves chemical reactions in rocks that replace minerals and mineral assemblages in the original material by new minerals or groups of minerals. The orientation and geometric arrangement of new inequant metamorphic minerals is largely controlled by the anisotropic pressure field associated with the tectonic processes.

Therefore, metamorphic rocks display not only a characteristic metamorphic mineral content they are also characterized by distinctive **metamorphic structures**. The structure of metamorphic rocks is used for the classification of the rocks. The structures of metamorphic rocks inherit essential information about the type of tectonic setting in which they formed and about the nature of metamorphism. Structural and chemical (petrological) aspects of metamorphic rocks are of equal importance in the study of metamorphism and the characterization of metamorphic rocks requires a description of their structure. Some important and typical structures of metamorphic rocks are defined by descriptive terms. The definition of terms given in this chapter partly follows recommendations of the “International Union Commission of Geological Sciences Subcommission on the Systematics of Metamorphic Rocks” (Fettes and Desmonds 2007).

Structure. The arrangement of parts of a rock mass irrespective of scale, including geometric interrelationships between the parts, their shapes and internal features. The terms micro-, meso- and mega- can be used as a prefix to describe the scale of the feature. Micro- is used for a thin-section scale, meso- for hand-specimen and outcrop scale, mega- for larger scales.

Fabric. The kind and degree of preferred orientation of parts of a rock mass. The term is used to describe the crystallographic and/or shape orientation of mineral grains or groups of grains, but can also be used to describe meso- and mega-scale features.

Layer. One of a sequence of near parallel tabular-shaped rock bodies. The sequence is referred to as being layered (equivalent expressions: bands, banded, laminated).

Foliation. Any repetitively occurring or penetrative planar structural feature in a rock body. Some examples:

- Regular layering on a cm or smaller scale
- Preferred planar orientation of inequant mineral grains
- Preferred planar orientation of lenticular (elongate) grain aggregates

More than one kind of foliation with more than one orientation may be present in a rock. Foliations may become curved (folded) or distorted. The surfaces to which they are parallel are designated **s-surfaces** (Fig. 2.1).

Schistosity. A type of foliation produced by deformation and/or recrystallization resulting in a preferred orientation of inequant mineral grains. It is common practice in phyllosilicate-rich rocks to use the term slaty cleavage instead of schistosity when individual grains are too small to be seen by the unaided eye (Fig. 2.1).

Cleavage. A type of foliation consisting of a regular set of parallel or sub-parallel closely spaced surfaces produced by deformation along which a rock body will usually preferentially split. More than one cleavage may be present in a rock.

Slaty cleavage. Perfectly developed foliation independent of bedding resulting from the parallel arrangement of very fine-grained phyllosilicates (Fig. 2.1).

Fracture cleavage. A type of cleavage defined by a regular set of closely spaced fractures.

Crenulation cleavage. A type of cleavage related to microfolding (crenulation) of a pre-existing foliation. It is commonly associated with varying degrees of metamorphic segregation.

Gneissose structure. A type of foliation on hand-specimen scale, produced by deformation and recrystallization, defined by:

- Irregular or poorly defined layering
- Augen and/or lenticular aggregates of mineral grains (augen structure, flaser structure)
- Inequant mineral grains which are present, however, only in small amounts or which display only a weak preferred orientation, thus defining only a poorly developed schistosity

Lineation. Any repetitively occurring or penetrative visible linear feature in a rock body (Fig. 2.1). This may be defined by, for example:

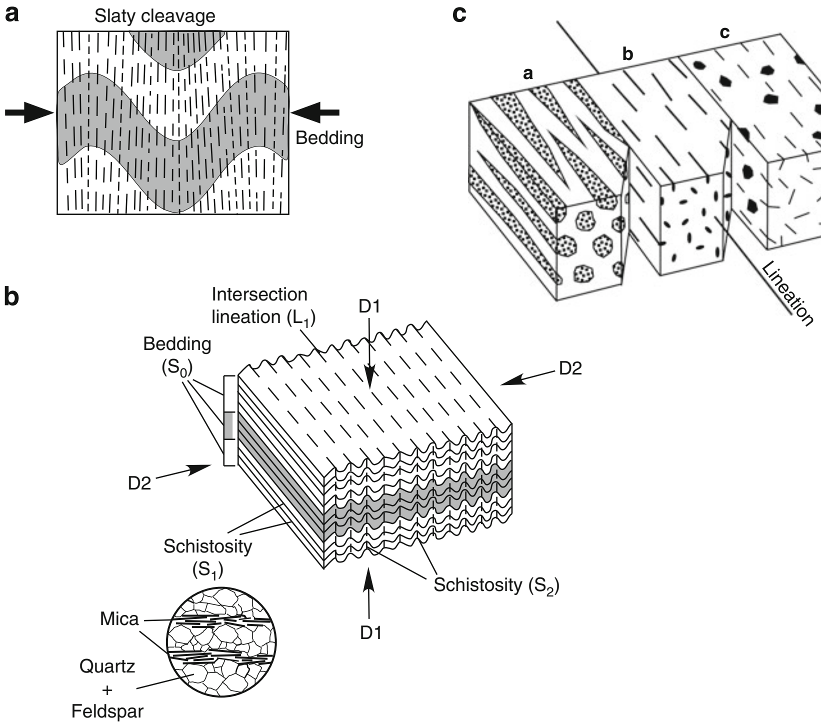

Fig. 2.1 Diagrams to illustrate some structure features in metamorphic rocks. (**a**) Folded sedimentary rocks and slaty cleavage; (**b**) Schist with alternating mica and quartz–feldspar foliation displaying two schistosity planes (S_1 and S_2) related to deformation events D_1 and D_2, respectively; The first schistosity, S_1, is parallel to sedimentary bedding planes (S_0); intersection of S_1 and S_2 planes produce an intersection lineation (L_1); (**c**) Examples of metamorphic lineation, a = alignment of mineral aggregates; b = alignment of elongate mineral grains (e.g. hornblende); c = alignment of tabular mineral grains (e.g. mica)

- Alignment of the long axes of elongate mineral grains (= mineral lineation)
- Alignment of elongate mineral aggregates
- Alignment of elongate objects, bodies (e.g. strongly deformed pebbles in a meta-conglomerate)
- Common axis of intersection of tabular mineral grains (or bodies)
- Intersection of two foliations (intersection lineation)
- Parallelism of hinge lines of small scale folds
- Slickenside striations
- Striations due to flexural slip

More than one kind of lineation, with more than one orientation, may be present in a rock. Lineations may become curved or distorted. The lines to which they are parallel are called **L-lines**. Reference to a lineation is incomplete without indication of the type concerned.

Joint. A single fracture in a rock with or without a small amount (<1 cm) of either dilatational or shear displacement (joints may be sealed by mineral deposits during or after their formation).

Cataclasis. Rock deformation accomplished by some combination of fracturing, rotation, and frictional sliding producing mineral grain and/or rock fragments of various sizes and often of angular shape.

Metamorphic differentiation. Redistribution of mineral grains and/or chemical components in a rock as a result of metamorphic processes. Metamorphic process by which mineral grains or chemical components are redistributed in such a way to increase the modal or chemical anisotropy of a rock (or portion of a rock) without changing the overall chemical composition.

Textural zones. Regional geological mapping of metamorphic terranes is typically based on criteria such as lithologic associations, metamorphic zones and structural zones. Field and petrographic subdivision of metamorphic rocks has also been made on the basis of textural zones that subdivide rocks in terms of the degree of recrystallization, e.g. foliation, mineral segregation, increasing grain size, with increasing metamorphism. For example, four textural zones have been established for quartzofeldspathic rocks (metagreywacke) that comprise the Otago Schist, New Zealand (Fig. 2.2) in which the boundaries of the 10–50 km wide zones are gradational and are termed **isotects** (Bishop 1972; Turnbull et al. 2001). The macroscopic/microscopic criteria used to distinguish textural zones are essentially a mapping tool and cannot be used as a mineralogical or isochemical *P*–*T* indicator.

2.3 Classification and Names of Metamorphic Rocks

The names of metamorphic rocks are usually straightforward and self-explanatory. The number of special terms and cryptic expressions is relatively small. Nevertheless in order to be able to communicate with other geologists working with metamorphic rocks it is necessary to define commonly used names and expressions and to briefly review currently used classification principles for metamorphic rocks.

There is not one sole classification principle used for the description of metamorphic rocks, which consequently means that all metamorphic rocks may have a series of perfectly correct and accepted names. However, **modal mineral composition and mesoscopic structure are the main criteria for naming metamorphic rocks**. In addition, the **composition and the nature of the protolith** (original material) is an important classification criterion. Finally, some well-established **special names** are used also in metamorphic geology.

The names of metamorphic rocks consist of a **root** name and a series of **prefixes**. The root of the name may be a special name (e.g. amphibolite) or a name describing

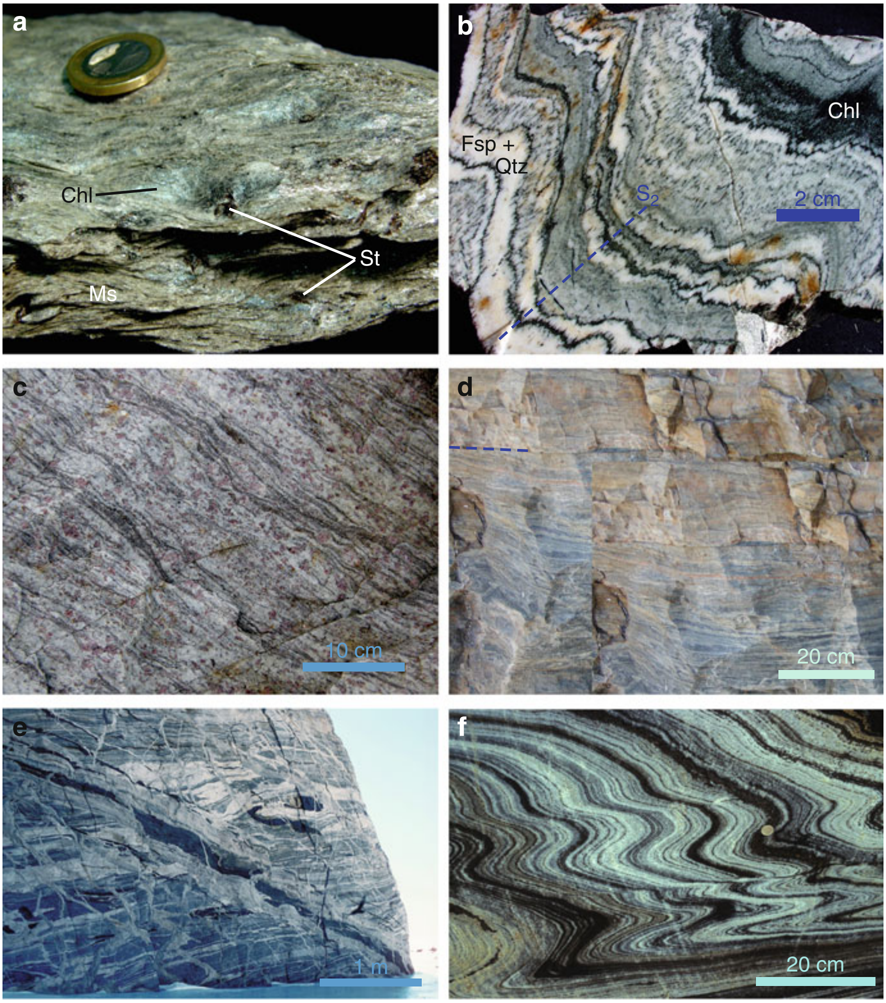

Fig. 2.2 Photos illustrating metamorphic rock fabrics. (**a**) St–Chl micaschist and schistosity (Saltdalen, Norway); (**b**) Banded and folded lower greenschist facies granite–gneiss; deformation and low grade metamorphism produced a banded orthogneiss from a Variscan granite, metamorphic segregation associated with deformation resulted in alternating white bands of Fsp + Qtz and green Chl bands, note Chl sheets oriented parallel to axial plains of folds (Arolla Gneiss, Matterhorn, Alps); (**c**) Gneiss and gneissosity, high-grade Spl–Sil–Bt–Grt gneiss (Jøkelfjorden, Norway); (**d**) Mylonite at the basal thrust of the Helvetic nappe complex (Glarus thrust, Piz Sardona, Swiss Alps); (**e**) Amphibolite facies banded orthogneiss with granite veins, *dark band in the upper part of picture* is a amphibolitized mafic dyke (Thor Range, Antarctica); (**f**) Upper amphibolite facies banded and folded marble, note extreme boudinage of bands of silicate rock (Engabreen, Svartisen, Norway)

the structure of the rock (e.g. gneiss). The root name always embraces some modally dominant metamorphic minerals (amphibolite is mainly composed of amphibole + plagioclase; gneiss is mainly composed of feldspar ± quartz).

The rock may be further characterized by adding prefixes to the root name. The prefixes may specify some typical structural features of the rock or may give some additional mineral information (e.g. banded epidote-bearing garnet–amphibolite, folded leucocratic garnet–hornblende gneiss). The prefixes are optional and the name may consist of the root only.

2.3.1 Rock Names Referring to the Structure

The structure of a rock concerns the characteristic distribution of its constituents (minerals, aggregates, layers etc.). It results from the geometrical arrangement of minerals, mineral aggregates with non-equant crystal shapes and other structural features, as discussed above. This structure is largely controlled by mechanical deformation and chemical segregation processes which are almost always associated with metamorphism. Expressions that mainly characterize the structure of metamorphic rocks are often used as root names. Metamorphic rocks are named primarily by using **descriptive structural terms**. The most important of these terms are:

Gneiss. A metamorphic rock displaying a gneissose structure (Fig. 2.2). The term gneiss may also be applied to rocks displaying a dominant linear fabric rather than a gneissose structure, in which case the term lineated gneiss may be used. This term gneiss is almost exclusively used for rocks containing abundant feldspar (±quartz), but may also be used in exceptional cases for other compositions (e.g. feldspar-free cordierite–anthophyllite gneiss). Examples: garnet–biotite gneiss, granitic gneiss, ortho-gneiss, migmatitic gneiss, banded gneiss, garnet–hornblende gneiss, mafic gneiss.

Schist. A metamorphic rock displaying on the hand-specimen scale a pervasive, well-developed schistosity defined by the preferred orientation of abundant inequant mineral grains (Fig. 2.2). For phyllosilicate-rich rocks the term schist is usually reserved for medium- to coarse-grained varieties, whilst finer-grained rocks are termed slates or phyllites. The term schist may also be used for rocks displaying a strong linear fabric rather than a schistose structure. Examples: epidote-bearing actinolite–chlorite schist (=greenschist), garnet–biotite schist, micaschist, calcareous micaschist, antigorite schist (=serpentinite), talc-kyanite schist (=whiteschist).

Phyllite. A fine-grained rock of low metamorphic grade displaying a perfect penetrative schistosity resulting from parallel arrangement of phyllosilicates. Foliation surfaces commonly show a lustrous sheen.

Slate. A very fine-grained rock of low metamorphic grade displaying slaty cleavage (Fig. 2.1).

Granofels. A metamorphic rock lacking schistosity, gneissose structure, and mineral lineations.

Fig. 2.3 Map showing distribution of textural zones (TZ's) delineating transition from non-schistose to schistose metagreywacke rocks, Otago, South Island of New Zealand (inset map shows distribution of Otago, Alpine and Marlborough schist) (Redrawn from Turnbull et al. 2001, Fig. 1D). TZI. Non foliated or spaced fracture cleavage. TZIIA. Weak-moderate, anastomosing, penetrative foliation that is stronger in fine-grained rocks. Sedimentary bedding dominant over cleavage. Rocks break into wedge-shaped blocks. Lowest grade semischist (sandstone = phyllite; mudstone = slate). TZIIB. Strong, penetrative foliation. Cleavage dominant over bedding which is transposed. No foliation parallel segregation of quartz and mica. Rocks break into parallel-sided slabs. Highest grade semischist (sandstone = phyllite; mudstone = slate). TZIII. Strong, penetrative foliation undulating on millimeter-scale. Distributed foliation-parallel quartz segregation lenses <1 mm thick. Psammitic-pelitic contacts sharp on millimeter scale. Lowest grade schist. TZIV. Strong, penetrative foliation undulating on millimeter- to centimeter-scale. Segregation banding (<1 mm thick) of quartz and mica. Psammitic–pelitic contacts blurred on millimeter-scale but distinguishable on centimeter-scale. Higher grade schist

2.3.2 Names for High-Strain Rocks

Metamorphism may locally be associated with an extremely high degree of rock deformation. Localized high strain in metamorphic terrains produces rocks with distinctive structures. Some widely used special names for high-strain rocks are defined below:

Mylonite. A rock produced by mechanical reduction of grain size as a result of ductile, non-cataclastic deformation in localized zones (shear zones, fault zones), resulting in the development of a penetrative fine-scale foliation, and often with an associated mineral and stretching lineation (Fig. 2.2).

Ultramylonite. A mylonite in which most of the megacrysts or lithic fragments have been eliminated (>90% fine-grained matrix).

Augen mylonite (blastomylonite). A mylonite containing distinctive large crystals or lithic fragments around which the fine-grained banding is wrapped.

Cataclasite. A rock which underwent cataclasis.

Fault breccia. Cataclasite with breccia-like structure formed in a fault zone.

Pseudotachylite. Ultra-fine-grained vitreous-looking material, flinty in appearance, occurring as thin veins, injection veins, or as a matrix to pseudo-conglomerates or -breccias, which seals dilatancy in host rocks displaying various degrees of fracturing.

2.3.3 Special Terms

Some commonly used and approved special terms (names, suffixes, prefixes) include:

Mafic minerals. Collective expression for ferro-magnesian minerals.

Felsic minerals. Collective term for quartz, feldspar, feldspathoids and scapolite.

Mafic rock. Rock mainly consisting of mafic minerals (mainly $\equiv$ modally >50%).

Felsic rock. rock mainly consisting of felsic minerals.

Meta-. If a sedimentary or igneous origin of a metamorphic rock can be identified, the original igneous or sedimentary rock term preceded by "meta" may be used (e.g. metagabbro, metapelite, metasediment, metasupracrustal). Also used to generally indicate that the rock in question is metamorphic (e.g. metabasite).

Ortho- and para-. A prefix indicating, when placed in front of a metamorphic rock name, that the rock is derived from an igneous (ortho) or sedimentary (para) rock respectively (e.g. orthogneiss, paragneiss).

Acid, intermediate, basic, ultrabasic. Terms defining the SiO_2 content of igneous and metamorphic rocks (respectively, >63, 63–52, 52–45, <45 wt% SiO_2).

Greenschist and greenstone. Schistose (greenschist) or non-schistose (greenstone) metamorphic rock whose green color is due to the presence of minerals such as chlorite, actinolite, and epidote (greenschist e.g. epidote-bearing actinolite-chlorite schist; greenstone, e.g. chlorite-epidote granofels).

Blueschist. Schistose rock whose bluish color is due to the presence of sodic amphibole (e.g. glaucophane schist). However, the "blue" color of a blueschist will not easily be recognized by a non-geologist (i.e. it is not really blue, although very rare outcrops of really blue glaucophanites do exist). Blueschists are schistose rocks containing amphibole with significant amounts of the M(4) cation position in the amphibole structure occupied by Na (glaucophane, crossite).

Amphibolite. Mafic rock predominantly composed of hornblende (>40%) and plagioclase.

Granulite. Metamorphic rock in or from a granulite facies terrain exhibiting characteristic granulite facies mineral assemblages. Anhydrous mafic minerals are modally more abundant than hydrous mafic minerals. Muscovite is absent in such rocks. Characteristic is the occurrence of metamorphic orthopyroxene in both mafic and felsic rocks. The term is not used for marbles and ultramafic rocks in granulite facies terranes.

Charnockite, mangerite, jotunite, enderbyite. Terms applied to orthopyroxene-bearing rocks with igneous texture and granitic (charnockite), monzonitic (mangerite, jotunite), and tonalitic (enderbyite) composition, irrespective of whether the rock is igneous or metamorphic.

Eclogite. A plagioclase-free mafic rock mainly composed of omphacite and garnet, both of which are modally adundant.

Eclogitic rock. Rocks of any composition containing diagnostic mineral assemblages of the eclogite facies (e.g. jadeite–kyanite–talc granofels).

Marble. A metamorphic rock mainly composed of calcite and/or dolomite (e.g. dolomitic marble).

Calc-silicate rock. Metamorphic rock which, besides 0–50% carbonates, is mainly composed of Ca-silicates such as epidote, zoisite, vesuvianite, diopside–hedenbergite, Ca-garnet (grossular–andradite), wollastonite, anorthite, scapolite, Ca-amphibole.

Skarn. A metasomatic Ca–Fe–Mg–(Mn)-silicate rock often with sequences of compositional zones and bands, formed by the interaction of a carbonate and a silicate system in mutual contact. Typical skarn minerals include, wollastonite, diopside, grossular, zoisite, anorthite, scapolite, margarite (Ca skarns); hedenbergite, andradite, ilvaite (Ca–Fe skarns); forsterite, humites, spinel, phlogopite, clintonite, fassaite (Mg skarns); rhodonite, tephroite, piemontite (Mn skarns).

Blackwall. A chlorite- or biotite-rich rock developed by metasomatic reaction between serpentinised ultramafic rocks and mafic rocks or quartzo-feldspathic rocks, respectively.

Rodingite. Calc-silicate rock, poor in alkalis and generally poor in carbonates, generated by metasomatic alteration of mafic igneous rocks enclosed in serpentinized ultramafic rocks. The process of rodingitization is associated with oceanic metamorphism (serpentinization of peridotite, rodingitization of enclosed basic igneous rocks such as gabbroic/basaltic dykes). Metarodingite is a prograde metamorphic equivalent of rodingite produced by oceanic metamorphism.

Quartzite or metachert. A metamorphic rock containing more than about 80% quartz.

Serpentinite. An ultramafic rock composed mainly of minerals of the serpentine group (antigorite, chrysotile, lizardite), e.g. diopside–forsterite–antigorite schist.

Hornfels. Is a non-schistose very fine-grained rock mainly composed of silicate ± oxide minerals that shows substantial recrystallization due to contact metamorphism. Hornfelses often retain some features inherited from the original rock such as graded bedding and cross-bedding in hornfelses of sedimentary origin.

Migmatite. Composite silicate rock, pervasively heterogeneous on a meso- to megascopic scale, found in medium- to high-grade metamorphic terrains (characteristic rocks for the middle and lower continental crust). Migmatites are composed of dark (mafic) parts (**melanosome**) and light (felsic) parts (**leucosome**) in complex structural association. The felsic parts formed by crystallization of locally derived partial melts or by metamorphic segregation, the mafic parts represent residues of the inferred partial melting process or are formed by metamorphic segregation. Parts of the felsic phases may represent intruded granitic magma from a more distant source.

Restite. Remnant of a rock, chemically depleted in some elements relative to its protolith. The depletion is the result of partial melting of that rock, e.g. emery rock.

2.3.4 Modal Composition of Rocks

The mineral constituents of metamorphic rocks are classified as follows: (1) *Major constituent*: present in amounts >5 vol%. To account for a major constituent not included in the definition of a rock name, the mineral name is put in front of the rock name (e.g. muscovite-gneiss, epidote–amphibolite). It follows from this that epidote–amphibolite is a rock which is mainly composed of epidote, hornblende and plagioclase. A garnet–staurolite gneiss is a metamorphic rock which is mainly composed of feldspar and quartz (included in the root name "gneiss"). In addition, it contains modally more staurolite than garnet as major constituents. (2) *Minor constituent*: present in amounts <5 vol%. If one wishes to include a minor constituent mineral in the rock name it is connected with "-bearing", e.g. rutile–ilmenite-bearing garnet–staurolite gneiss (contains less rutile than ilmenite). (3) *Critical mineral (or mineral assemblage)*: indicating distinctive conditions for the formation of a rock (or distinct chemical composition of the rock) based on the presence or absence of a critical mineral or mineral assemblage. The critical mineral(s) may be present as a major or/and minor constituent(s).

It is up to the geologist to decide how many and which of the minerals that are not included in the definition of the rock name need to be prefixed. It is also possible to use abbreviations of mineral names in naming rocks, e.g. Bt–Ms gneiss for biotite–muscovite gneiss. Recommended standard abbreviations for mineral names are listed in the *Appendix* and are used throughout this book.

2.3.5 Names Related to the Origin of the Protolith

Metamorphic rocks can be grouped into seven chemical composition classes according to the most common rock types in the crust and mantle.

Examples of such rock names include: metapelite, metabasite, metabasalt, metapsamite, metamarl (e.g. calcareous micaschist), metagranite, metagabbro, semi-pelitic gneiss, metaeclogite.

2.4 Mineral Assemblages and Mineral Parageneses

The petrogenesis of metamorphic rocks is often a long-lasting complicated process in time and space which perhaps may endure for 30 Ma or so (see Chap. 3). The geologists collects the end product of such a long-drawn-out process at the Earth's surface (0.1 MPa and 25°C). The geologist is, of course, interested in finding out something about the evolution, the petrogenesis of the collected rock at this sample locality, to compare it with rocks from other localities and finally draw some logical conclusions on the large-scale processes that caused the metamorphism of the observed rock.

One of the important and successful methods to decipher the petrogenesis and evolution of metamorphic rocks is the application of chemical thermodynamics to the heterogeneous chemical systems exemplified by rocks. Rocks are usually composed of a number of different minerals that represent chemical subspaces of the rock, so-called phases. Most metamorphic rocks are also composed of a number of different chemical constituents that are designated system components. The chemical constituents that make up the bulk composition of rocks are distributed among a group of mineral species or phases, each with a distinct composition (heterogeneous system). The number, composition and identity of the minerals is uniquely defined by the laws of thermodynamics and depends exclusively on the prevailing intensive variables such as pressure and temperature. The group of minerals that make up a rock at equilibrium is designated as the **equilibrium mineral assemblage** or equilibrium phase assemblage. The succession of mineral assemblages that follow and replace one another during the metamorphic evolution of a given terrain are designated **mineral parageneses**.

In practical work with metamorphic rocks it is impossible to demonstrate that a given mineral assemblage once coexisted in chemical equilibrium. Therefore, one uses a less rigorous definition: a **mineral assemblage** is an association of mineral species in mutual grain contact. The assemblage occurs in a chemically homogeneous portion of a rock. With thin section examination, it is convenient to use a matrix table for marking all observed two-phase grain contacts. An example: in a thin section of a metamorphic rock showing a coarse-grained mosaic structure, the following mafic minerals are identified: staurolite, garnet, biotite and kyanite. All observed grain contacts between two of these minerals are marked with an "X" in Table 2.4. The thin-section observations summarized in Table 2.4 imply that the four minerals indeed represent a mineral assemblage.

In rocks displaying clear disequilibrium features (compositional zoning of minerals, reaction rims, symplectitic structures, replacement structures, exsolution

Table 2.4 Practical determination of a mineral assemblage

Mineral	Staurolite	Garnet	Biotite	Kyanite
Staurolite	X	X	X	X
Garnet		X	X	X
Biotite			X	X
Kyanite				X

A cross in the Grt–St cell means; garnet and staurolite have mutual grain contacts in thin section

structures), it is often difficult to determine mineral assemblages that represent a particular stage in the evolutionary history of the rock.

One has to keep in mind that mineral assemblages are usually identified in thin sections which represent a two-dimensional section through a volume of rock. In such sections, only two and a maximum of three minerals may be in mutual grain contact. In the three-dimensional rock a maximum of four minerals can be in contact at a point in space, three minerals form contacts along a line, and two minerals contact along surfaces (that show as lines in thin sections). It is therefore strongly recommended to study **more than one thin section** from the most significant and interesting samples. As an example, in a series of 20 thin-sections of a single hand specimen of a coarse-grained sapphirine–granulite from the Western Alps, unique assemblages and unique structures were found in nearly all of the sections. The problem is especially acute in coarse-grained samples, where the scale of chemical homogeneity may considerably exceed the size of a thin section.

In extremely fine-grained samples the mineral assemblage must be determined by X-ray techniques. In this case it is not possible to maintain the requirement of mutual grain contact in the definition of an assemblage. However, this shortcoming may be overcome by using the high magnification backscatter image mode of an electron probe microanalyser (EPMA).

Rocks should always be examined by X-ray techniques in order to identify minerals which are difficult to distinguish under the microscope (e.g. muscovite, paragonite, talc, pyrophyllite, also quartz and untwinned albite).

Staining techniques help in distinguishing some important rock-forming minerals with similar optical properties such as calcite from dolomite.

Minerals occurring as inclusions in refractory minerals such as garnet but not in the matrix of the rock do not belong to the main matrix assemblage. In the example of the staurolite–garnet–biotite–kyanite assemblage cited above, garnet may show small composite two-phase inclusions of chlorite and chloritoid. In this case, chlorite–chloritoid–garnet constitutes another, earlier assemblage of the rock.

During metamorphism some earlier-formed minerals may become unstable and react chemically to form a new, more stable assemblage. However, metastable relics of the early assemblage may partly survive. Great care must be taken in the study of metamorphic micro-structures in order to avoid mixing up mineral assemblages. The correct identification of a successive series of mineral assemblages, i.e. the paragenesis of a metamorphic rock, represents the “great art” of metamorphic petrology. It can be learned only by experience.

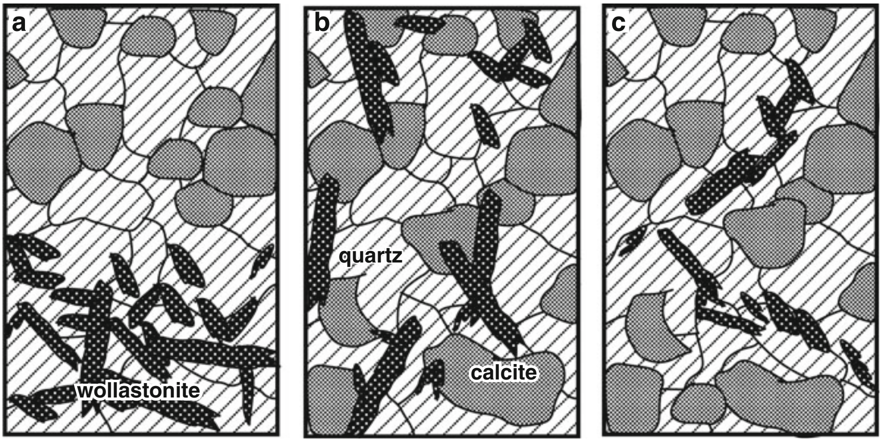

Fig. 2.4 (**a**–**c**) Mineral assemblages in three different rocks all containing wollastonite, calcite and quartz

Figure 2.4 shows some of the aspects related to the recognition of mineral assemblages. Three fictive rocks all contain the minerals, quartz, calcite and wollastonite on the scale of a thin section. The general micro-structure of the three sections (Fig. 2.4) shows the distribution of the three minerals in their respective rocks. Rock A is clearly heterogeneous on the scale of a thin section, the upper part contains the assemblage Qtz + Cal, the lower half of the section is free of carbonate and contains the assemblage Qtz + Wo. The rock does not contain the assemblage Qtz + Cal + Wo. The two parts of the section are different in overall composition. Note, however, on a volatile-free basis, the two domains of the rock may have very similar or identical compositions (same Ca/Si ratio). In rock B, obviously all three minerals can be found in mutual grain contact, and Qtz + Cal + Wo constitutes the mineral assemblage. In rock C, which appears to be compositionally homogeneous on thin-section scale, Qtz + Cal and Qtz + Wo form common grain boundaries. However, no Wo + Cal grain boundaries can be observed. Thus, although the three phases Qtz + Cal + Wo do not strictly represent a mineral assemblage, many petrologists would probably approve it being labeled as an equilibrium assemblage. [Comment: after the publication of the first edition of this book Fig. 2.4 (originally Fig. 2.1), inspired a lively discussion among petrologists on the geo-metamorphism mailbase. A number of comments were made and it was suggested that the figure be modified. We prefer to keep it essentially in its original form and some aspects of the debate will be discussed in Sect. 3.7. We strongly recommend that the reader and anyone interested in rock metamorphism subscribe to the mailbase geo-metamorphism where one sometimes captures a stimulating dispute on questions related to geo-metamorphism. The present address is: geo-metamorphism@jiscmail.ac.uk: you need to send an e-mail with only the following in the body of the message: *subscribe geo-metamorphism*].

2.5 Graphical Representation of Metamorphic Mineral Assemblages

Once the mineral assemblage of a rock has been identified, it is convenient or even necessary to represent graphically the chemical composition of the minerals that constitute the assemblage. Such a figure is called a **chemograph** and represents a **composition phase diagram**. The geometric arrangement of the phase relationships on such a phase diagram is called the **topology**. Composition phase diagrams can be used only to document the assemblages found in rocks of a given metamorphic terrain or outcrop. However, such diagrams are an indispensable tool for the analysis of metamorphic characteristics and evolution of a terrain. They can be used to deduce sequences of metamorphic mineral reactions. Finally, composition phase diagrams can also be calculated theoretically from thermodynamic data of minerals which permits the quantitative calibration of field-derived chemographs.

Composition phase diagrams display the chemical composition of minerals and the topologic relationships of mineral assemblages. Variables on the diagrams are concentrations or amounts of chemical entities. All other variables that control the nature of the stable mineral assemblage such as pressure and temperature must be constant. Thus, composition phase diagrams are isothermal and isobaric. Also, not more than two composition variables can conveniently be displayed on a two-dimensional xy-diagram (sheet of paper, computer screen).

2.5.1 Mole Numbers, Mole Fractions and the Mole Fraction Line

It is useful to change the scale for the compositional variables from wt% to mol%, mole fractions or mole numbers. Most chemographs use mol% or mole fraction as units for the composition variables. The mineral forsterite (Fo), for example, is composed of 42.7 wt% SiO_2 and 57.3 wt% MgO. Mole numbers and mole fractions for this mineral are calculated as follows:

SiO_2: 42.7/60.1 (molecular weight SiO_2) = **0.71** (number of moles of SiO_2 per 100 g Fo)

MgO: 57.3/40.3 (molecular weight MgO) = **1.42** (number of moles of MgO per 100 g Fo)

Forsterite has a MgO/SiO_2 ratio of 1.42/0.71 = 2 and thus has 2 mol MgO per 1 mol SiO_2. The composition is reported as Mg_2SiO_4 or ($2MgO\ SiO_2$) or 66.66% MgO + 33.33% SiO_2 or $2/3MgO + 1/3SiO_2$. This is all equivalent. However, the last version has many advantages ⇒ mole fraction basis.

The mole fraction is defined as follows:

$$X_{MgO} = \frac{(\text{Number of moles of MgO})}{(\text{Number of moles of MgO}) + (\text{Number of moles of SiO}_2)}$$

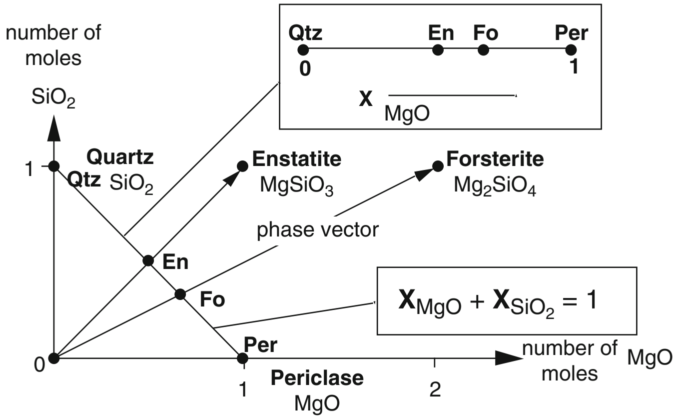

Fig. 2.5 Composition space and the two-component MgO–SiO_2 system

For the example above: X_{MgO} = 2/(2 + 1) = 1.42/(1.42 + 0.71) = 0.66 (dimensionless quantity).

Figure 2.5 shows a graphical representation of the two component system MgO and SiO_2 in a rectangular coordinate system. The two components MgO and SiO_2 define the two-dimensional **composition space**. The graphical representation of the compositions of forsterite, enstatite, quartz and periclase is a point in the Cartesian coordinate system with the axes: numbers of moles MgO and numbers of moles SiO_2. It is apparent in Fig. 2.5 that the composition of forsterite, for example, can also be viewed as a vector in the composition space with the elements (2,1). Mineral compositions are unequivocally characterized by a vector, the **phase vector**, in the composition space.

Mineral compositions can also be represented in terms of mole fractions rather than mole numbers. The mole fractions of the two **system components** MgO and SiO_2 sum to unity in all four minerals represented in Fig. 2.5. This can be expressed by the general relationship:

$$X_{\mathrm{MgO}} + X_{\mathrm{SiO_2}} = 1$$

This equation represents a straight line in Fig. 2.5 connecting X_{MgO} = 1 with X_{SiO_2} = 1. The phase vectors of forsterite and enstatite intersect the mole fraction line at unique positions (En and Fo, respectively; Fig. 2.5). It is, therefore, not necessary to represent phase composition in a two-component system on a two-dimensional diagram. The topologic information is contained on the mole fraction line. The dimension of the graph can be reduced from two to one. If the concentration of one component is known in a phase that is composed of two components, the second concentration is known as well. In an n-component system there are n − 1 independent compositional variables, the remaining concentration is given be the equation (or: n composition variables and one equation relating them):

$$\sum_{i=1}^{n} X_i = 1$$

Note also, for example, that enstatite compositions expressed as $MgSiO_3$, $Mg_2Si_2O_6$, $Mg_4Si_4O_{12}$ all have the same intersection point with the mole fraction line (multiplying the phase vector for enstatite with a scalar preserves its position on the mole fraction line). Phase compositions in a two-component system can be represented on a mole fraction line as shown in Fig. 2.5.

2.5.2 The Mole Fraction Triangle

The mineral talc (Tlc) is composed of three simple oxide components and its composition can be written as: $H_2Mg_3Si_4O_{12}$. Based on a total of 12 oxygens per formula unit, talc consists of 1 mol H_2O, 3 mol MgO and 4 mol SiO_2. A graphical representation of the talc composition in the three-component system is shown in Fig. 2.6. The three components, MgO, SiO_2 and H_2O, define a Cartesian coordinate system with numbers of moles of the components displayed along the coordinate axes. The unit vectors of the system components span the composition space. The talc composition is represented by a phase vector with the elements; ×3 the unit vector of MgO, ×4 the unit vector of SiO_2 and ×1 the unit vector of H_2O.

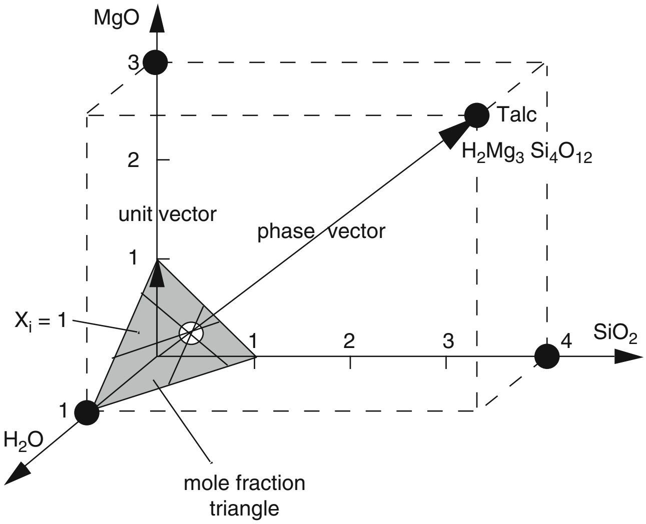

Fig. 2.6 Composition space and the three-component $MgO–H_2O–SiO_2$ system

The total number of moles of system components is 8 and the composition of talc can be normalized to a total number of 8 mol of system components. In this case the talc composition will be expressed by ⇒ mole fractions: $X_{H_2O} = 1/8$, $X_{MgO} = 3/8$, $X_{SiO_2} = 4/8$, or: $X_{H_2O} = 0.125$, $X_{MgO} = 0.375$, $X_{SiO_2} = 0.50$; in mol%: $H_2O = 12.5\%$, $MgO = 37.5\%$, $SiO_2 = 50\%$. Graphic representation of the talc composition on a mole fraction basis is given by the intersection of the talc phase vector in Fig. 2.6 and the plane $\Sigma X_i = 1$. The mole fraction plane is a regular triangle with the corners $X_i = 1$. This triangle is called the **mole fraction triangle**.

A representation of the mole fraction triangle is shown in Fig. 2.7. The lines of constant X_i are parallel with the base lines of the triangle. This follows from Fig. 2.6, where it can be seen that the planes X_i = constant and $X_{j \neq i} = 0$ intersect the mole fraction plane along lines parallel to the base line $X_i = 0$. Three rulers with $X_i = 0.1$ increments are also shown in Fig. 2.7 for the three components. Triangular coordinate paper is available on the market.

A further example of a mineral composition in a three-component system $MgO–SiO_2–H_2O$ is the chemical formula of chrysotile (one of many serpentine minerals) that can be written as, $H_4Mg_3Si_2O_9$. The sum of moles of system components is 7; ⇒ mole fractions: $X_{H_2O} = 2/7$, $X_{MgO} = 3/7$, $X_{SiO_2} = 2/7$, or equivalent: $X_{H_2O} = 0.29$, $X_{MgO} = 0.42$, $X_{SiO_2} = 0.29$; expressed in mol%: $H_2O = 29\%$, $MgO = 42\%$, $SiO_2 = 29\%$.

Because triangular coordinate drawing paper is not always at hand for plotting composition data, the formulae for recalculation into Cartesian coordinates are given in Fig. 2.8 together with the position of the serpentine composition. The value of the scaling factor "f" depends on the desired size of the figure. The scaling factor must be multiplied by 100 when using mole fractions rather than mol%.

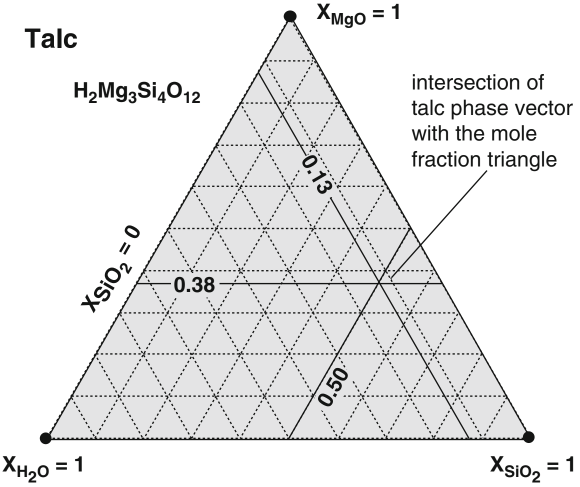

Fig. 2.7 Mole fraction triangle $MgO–H_2O–SiO_2$ and intersection coordinates of the talc phase vector with the mole fraction triangle

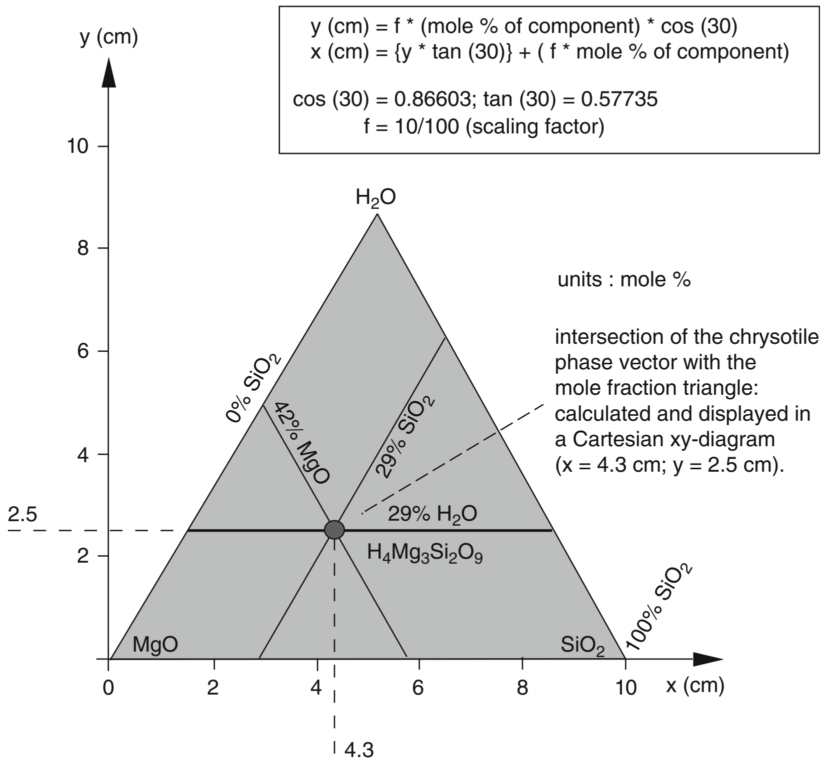

Fig. 2.8 Mole fraction triangle MgO–H_2O–SiO_2 in the Cartesian coordinate system

2.5.3 *Projections*

2.5.3.1 Simple Projections

On a mole fraction triangle two compositional variables of a three-component system can be depicted. Most rocks, however, require more than three components to describe and to understand the phase relationships. Graphical representation of an eight component system requires a seven-dimensional figure. Projection phase diagrams are graphical representations of complex n-component systems that show two composition variables at a time while keeping the other n − 3 composition variables constant. The remaining variable is given by the mole fraction equation as outlined above.

In a suite of samples of similar bulk composition (e.g. 20 samples of metapelite), one often finds certain mineral species that are present in many of them. This circumstance permits projection of the phase compositions from the composition of one such mineral which is present in excess. For instance, in many metapelitic rocks quartz is modally abundant whereas calcite is the main mineral in most marbles.

Composition phase diagrams for metapelites can therefore be constructed by projecting through SiO_2 onto an appropriate mole fraction triangle, and for marbles by projection the phase compositions from $CaCO_3$.

The system $MgO–SiO_2–H_2O$ (MSH system) can be used to explain the basic principle of making projections. Some phase compositions in the MSH system are given in Table 2.5. This table represents a **composition matrix** with oxide components defining unit vectors of the composition space and the mineral compositions used as column vectors.

Figure 2.9 shows the chemographic relationships in the ternary system $H_2O–MgO–SiO_2$. The corners of the triangle represent $X_i = 1$ or 100 mol% of the system components. The lines connecting them are the binary subsystems of the three-component system. The compositions of the black circles occur as stable phases in nature (phase components). The ternary system has three binary subsystems ($MgO–SiO_2$, $MgO–H_2O$, and $SiO_2–H_2O$). Some phase compositions in the ternary system fall on straight lines such as Tlc–Ath–En and Brc–Atg–Tlc. This means that the phase compositions along a straight line, e.g. Tlc–Ath–En, are linearly dependent; one of these compositions can be expressed by the other two (4En + Tlc = Ath). Therefore, there are only two components required to describe the compositions of the other phases on the straight line ⇒ **pseudobinary join.** The colinearity is also said to be a **compositional degeneracy** in the system.

Now, one might wish to analyze and discuss phase relationships among the minerals shown in Fig. 2.9 for geologic situations in which an aqueous fluid phase (H_2O) is present in all rocks and in equilibrium with the solid phase assemblage. The presence of excess water in all the considered rocks permits projection of the other phase compositions through water onto the $MgO–SiO_2$-binary (in principle on any pseudobinary as well). Imagine that you are standing in the H_2O corner of Fig. 2.9. What you will see from there is shown at the bottom of Fig. 2.9. The

Table 2.5 Phase compositions in the MSH-system

(a) Composition matrix (moles); columns are phase vectors, composition space defined by the system components SiO_2, MgO, H_2O

	Fo	Brc	Tlc	En	Ath	Qtz	Per	Atg	Fl
SiO_2	1	0	4	2	8	1	0	2	0
MgO	2	1	3	2	7	0	1	3	0
H_2O	0	1	1	0	1	0	0	2	1
Sum	3	2	8	4	16	1	1	7	1

(b) Composition matrix (mole fractions); columns are normalized phase vectors, values are coordinates on the molefraction triangle

SiO_2	0.33	0.00	0.50	0.50	0.50	1.00	0.00	0.29	0.00
MgO	0.67	0.50	0.38	0.50	0.44	0.00	1.00	0.43	0.00
H_2O	0.00	0.50	0.13	0.00	0.06	0.00	0.00	0.29	1.00
Sum	1	1	1	1	1	1	1	1	1

(c) Composition matrix (mole fractions); projected through H_2O, columns are normalized phase vectors, values are coordinates on the SiO_2–MgO-binary (mole fraction line)

SiO_2	0.33	0.00	0.56	0.50	0.53	1.00	0.00	0.40	0.00
MgO	0.67	1.00	0.43	0.50	0.47	0.00	1.00	0.60	0.00

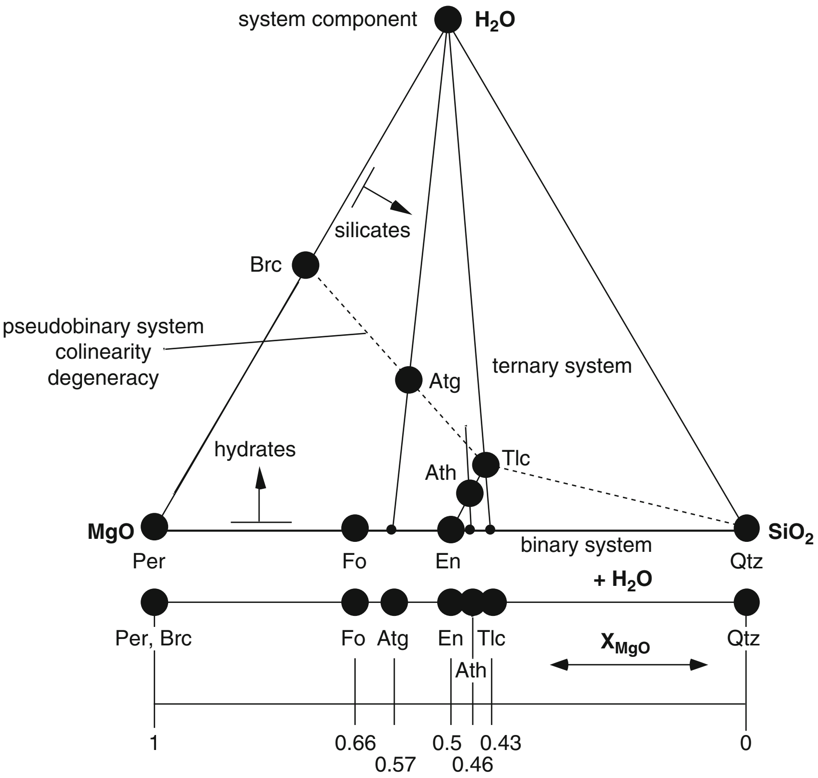

Fig. 2.9 Projection of phase compositions in the $MgO–H_2O–SiO_2$ system through H_2O onto the $MgO–SiO_2$ binary

chemography on the $MgO–SiO_2$ binary is a projected chemography and the positions of the mineral compositions on this binary are expressed in terms of mole fractions X_{Mg}. They can be calculated from the composition matrix (Table 2.5b) by first; deleting the row containing the component one wishes to project from (in this case H_2O) and second; renormalizing the column vectors to unity (Table 2.5c). It follows that projections from quartz, periclase and H_2O can be prepared using the composition matrix of Table 2.5b. It must be stressed again, however, that the projections are meaningful only if the projection component is present as a phase of fixed composition in all rocks of interest. The projected chemography in Fig. 2.9 cannot be used in H_2O-absent situations. Likewise, an analogous projection through MgO onto the $SiO_2–H_2O$-binary requires that periclase is present in all assemblages, and projection through SiO_2 onto the $MgO–H_2O$-binary requires that quartz is present in all assemblages. It is therefore necessary to write the projection compositions on any projected composition phase diagram. In simple projections, the projecting phase component is identical with one of the simple oxide system components.

2.5.3.2 Projection of Solid-Solution Phases

In nature, the magnesian minerals shown in Fig. 2.9 have variable substitution of Mg by Fe. If the restriction of water saturation is maintained, phase compositions in the FeO–MgO–SiO_2–H_2O system can be projected from H_2O onto the mole fraction triangle FeO–MgO–SiO_2, as shown in Fig. 2.10. In certain applications one does not wish to consider the complexity arising from the Fe–Mg substitution. In such cases, one projects the compositions onto the MgO–SiO_2–binary parallel to the direction of the substitution vector $MgFe_{-1}$. The resulting projection is identical to the binary chemography shown in Fig. 2.9. All solid-solution phases must be projected along exchange vectors. More examples will be given below.

2.5.3.3 Complex Projections

The technique of projecting phase compositions in the m-dimensional space from one system component onto a subspace with m − 1 dimensions is one of the two key steps in producing geologic composition phase diagrams. In the MSH system, used as an example here, the projection was made from the H_2O component in the three-component space onto the MgO–SiO_2 binary. However, often one wishes to project from a composition other than the original system components. Most commonly, it is necessary to find a projection from a composition of a phase that is present in excess. In order to make this possible, the composition matrix has to be

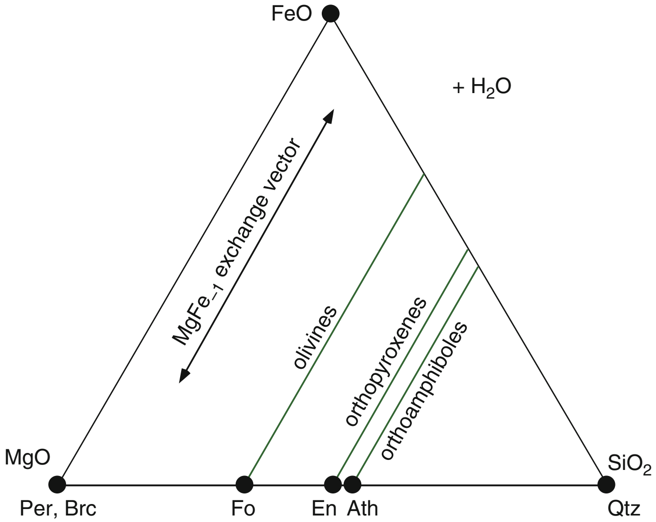

Fig. 2.10 Projection of phase compositions in the FeO–MgO–H_2O–SiO_2 system through H_2O and along the $MgFe_{-1}$ exchange vector onto the MgO–SiO_2 binary

rewritten in terms of a set of new system components. One of these new components must be the desired new projection composition.

For example, we would like to prepare a figure representing the phase compositions in the MSH system on a mole fraction triangle with the corners Mg_2SiO_4 (Fo), $Mg(OH)_2$ (Brc), and $Mg_3Si_4O_{10}(OH)_2$ (Tlc). Secondly, we would like to study phase relationships in rocks which contain excess forsterite and need a projection from Mg_2SiO_4 (Fo) onto the Brc–Tlc-binary.

The solution to the problem is shown in Table 2.6. In Table 2.6a the phase compositions are expressed in terms of the new system components Mg_2SiO_4 (Fo), $Mg(OH)_2$ (Brc), and $Mg_3Si_4O_{10}(OH)_2$ (Tlc). As an example, the composition of enstatite can be expressed by 2Fo − 1Brc + 1Tlc, which is equivalent to 6MgO + $6SiO_2$. Table 2.6b gives the coordinates of the mineral compositions in the mole fraction triangle Fo–Brc–Tlc. The algebraic operation which transforms the composition space expressed in terms of simple oxide components (Table 2.5a) to the composition space expressed in terms of Mg_2SiO_4 (Fo), $Mg(OH)_2$ (Brc), and $Mg_3Si_4O_{10}(OH)_2$ (Tlc) components (Table 2.6a) is also given in Table 2.6. It can be seen that the operation is a pre-multiplication of Table 2.5a by the inverse of the leading 3 × 3 square matrix in Table 2.5a. The result of the operation is the composition matrix (Table 2.7a) with the mineral compositions expressed by the new set of system components. Today, any standard commercial spreadsheet

Table 2.6 Phase compositions in the Fo–Brc–Tlc-system

(a) Composition matrix in terms of moles; columns are phase vectors, composition space defined by the system components

	Fo	Brc	Tlc	En	Ath	Qtz	Per	Atg	Fl
Fo	1.00	0.00	0.00	2.00	4.00	−1.00	4.00	0.00	−4.00
Brc	0.00	1.00	0.00	−1.00	−2.00	−1.00	1.00	3.00	5.00
Tlc	0.00	0.00	1.00	1.00	5.00	1.00	−1.00	1.00	1.00

(b) Composition matrix in terms of mole fractions; columns are normalized phase vectors, values are coordinates on the mole fraction triangle

Fo	1.00	0.00	0.00	1.00	0.57	1.00	1.00	0.00	−2.00
Brc	0.00	1.00	0.00	−0.50	−0.29	1.00	0.25	0.75	2.50
Tlc	0.00	0.00	1.00	0.50	0.71	−1.00	−0.25	0.25	0.50

Initial composition matrix (old system components): $[A][B_{OC}]$ (Table 2.5a)
New matrix (new system components) $[I][B_{NC}]$ (Table 2.6a)
Matrix operation: $[A^{-1}][A][B_{OC}] \Rightarrow [I][B_{NC}]$

Old basis [A]			Inverse of old basis [A−1]			Identity matrix [I]		
1	0	4	−0.33	0.67	−0.67	1	0	0
2	1	3	−0.33	0.17	0.83	0	1	0
0	1	1	0.33	−0.17	0.17	0	0	1

(c) Composition matrix in terms of mole fractions; projection through Mg_2SiO_4 (Fo), columns are normalized phase vectors, values are coordinates on the mole fraction line

	Fo	Brc	Tlc	En	Ath	Qtz	Per	Atg	Fl
Brc	0.00	1.00	0.00	−1.00	−0.69	1.00	0.00	0.75	0.83
Tlc	0.00	0.00	1.00	1.00	1.69	−1.00	0.00	0.25	0.17

Table 2.7 Composition matrix for minerals in the KFMASH-system

(a) Composition matrix in terms of oxide components

	Ky	Qtz	Ms	Fl	FeO	MgO	Alm	Prp	Ann	Phl	FEs	Es	St	Cld	Crd	OPX	Spl	Chl
SiO_2	1	1	6	0	0	0	3	3	6	6	5	5	8	2	5	1.8	0	3
Al_2O_3	1	0	3	0	0	0	1	1	1	1	2	2	9	2	2	0.2	1	1
FeO	0	0	0	0	1	0	3	0	6	0	5	0	4	2	0	0.5	0	0
MgO	0	0	0	0	0	1	0	3	0	6	0	5	0	0	2	1.3	1	5
K_2O	0	0	1	0	0	0	0	0	1	1	1	1	0	0	0	0.0	0	0
H_2O	0	0	2	1	0	0	0	0	2	2	2	2	2	2	0	0.0	0	4

(b) Inverse of leading 6 × 6 square matrix (A-matrix ⇒ A^{-1})

0.00	1.00	0.00	0.00	−3.00	0.00
1.00	−1.00	0.00	0.00	−3.00	0.00
0.00	0.00	0.00	0.00	1.00	0.00
0.00	0.00	0.00	0.00	−2.00	1.00
0.00	0.00	1.00	0.00	0.00	0.00
0.00	0.00	0.00	1.00	0.00	0.00

(c) Composition matrix in terms of Ky, Qtz, Ms, H_2O, FeO and MgO

	Ky	Qtz	Ms	Fl	FeO	MgO	Alm	Prp	Ann	Phl	Fe-Es	Es	St	Cld	Crd	OPX	Spl	Chl
Ky	1.00	0.00	0.00	0.00	0.00	0.00	1.00	1.00	−2.00	−2.00	−1.00	−1.00	9.00	2.00	2.00	0.20	1.00	1.00
Qtz	0.00	1.00	0.00	0.00	0.00	0.00	2.00	2.00	2.00	2.00	0.00	0.00	−1.00	0.00	3.00	1.60	−1.00	2.00
Ms	0.00	0.00	1.00	0.00	0.00	0.00	0.00	0.00	1.00	1.00	1.00	1.00	0.00	0.00	0.00	0.00	0.00	0.00
Fl	0.00	0.00	0.00	1.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	2.00	2.00	0.00	0.00	0.00	4.00
FeO	0.00	0.00	0.00	0.00	1.00	0.00	3.00	0.00	6.00	0.00	5.00	0.00	4.00	2.00	0.00	0.50	0.00	0.00
MgO	0.00	0.00	0.00	0.00	0.00	1.00	0.00	3.00	0.00	6.00	0.00	5.00	0.00	0.00	2.00	1.30	1.00	5.00

(d) Renormalized composition matrix, column vectors are coordinates of mineral compositions in the mole fraction triangle Ky (A), FeO (F) and MgO (M) (⇒ AFM-diagram), projection through Qtz, Ms and H_2O

	Ky	Qtz	Ms	Fl	FeO	MgO	Alm	Prp	Ann	Phl	Fe-Es	Es	St	Cld	Crd	OPX	Spl	Chl
Ky	1.00	0.00	0.00	0.00	0.00	0.00	0.25	0.25	−0.50	−0.50	−0.25	−0.25	0.69	0.50	0.50	0.10	0.50	0.17
FeO	0.00	0.00	0.00	0.00	1.00	0.00	0.75	0.00	1.50	0.00	1.25	0.00	0.31	0.50	0.00	0.25	0.00	0.00
MgO	0.00	0.00	0.00	0.00	0.00	1.00	0.00	0.75	0.00	1.50	0.00	1.25	0.00	0.00	0.50	0.65	0.50	0.83

program running on a PC or MAC will do these algebraic operations for you (e.g. Excel). We are now ready for the construction of the desired forsterite projection. Just as in simple projections, delete the Fo-row in Table 2.6b and renormalize to constant sum. Table 2.6c shows the coordinates of the phase compositions along the Brc–Tlc binary. Some of the compositions project to the negative side of talc, and periclase cannot be projected onto the Brc–Tlc binary at all. This is apparent also from Fig. 2.9, where it can be recognized that, seen from forsterite, periclase projects away from the Brc–Tlc binary. It is also clear from the procedure outlined above, that the compositions which one chooses to project from or which one wants to see at the apexes of the mole fraction triangle must be written as column vectors in the original A-matrix.

With the two basic operations, projection and redefinition of system components, one can construct any thermodynamically valid composition phase diagram for any geologic problem.

2.5.3.4 AFM Projections

A classical example of a composite projection is the AFM projection for metapelitic rocks (see also Chap. 7). Many of the phase relationships in metapelitic rocks can be described in the six-component system K_2O–FeO–MgO–Al_2O_3–SiO_2–H_2O (KFMASH system). A graphical representation of the system requires projection from at least three fixed compositions. Many of the metapelitic rocks contain excess quartz and some aspects of metamorphism can be discussed for water-present conditions. Therefore, a projection from SiO_2 and H_2O can be easily prepared. However, none of the remaining four components is present as a phase in such rocks. On the other hand, in low- and medium-grade metapelites muscovite is normally present as an excess phase, and in high-grade rocks muscovite is replaced by K-feldspar. Therefore, a useful diagram could be prepared by projecting through muscovite or K-feldspar onto the plane Al_2O_3–FeO–MgO. Furthermore, under the condition of excess quartz an Al_2SiO_5 polymorph is always more stable than corundum. This, in turn, requires that the composition matrix for minerals in metapelitic rocks is rewritten in terms of the new system components $KAl_3Si_3O_{10}(OH)_2$–Al_2SiO_5–FeO–MgO–SiO_2–H_2O as shown in Table 2.7. The coordinates of the phase compositions in the AFM mole fraction triangle can then be represented in an AFM diagram (Fig. 2.11). All Mg–Fe-free Al-silicates project to the A apex, pure enstatite, talc and anthophyllite are located at the M apex, ferrosilite and Fe-anthophyllite project to the F apex. Biotites have negative A-coordinates. The iron-magnesium substitution ($FeMg_{-1}$ exchange) in the ferromagnesian minerals is parallel to the AM binary, the Mg-tschermak substitution ($2AlSi_{-1}Mg_{-1}$ exchange) is parallel to the AM binary. Minerals such as staurolite, chloritoid, garnet, cordierite and spinel do not show any tschermak variation, their compositional variation is exclusively parallel to the $FeMg_{-1}$ exchange vector. Other minerals such as chlorites, biotites, orthopyroxenes and orthoamphiboles show both $FeMg_{-1}$ exchange and $2AlSi_{-1}Mg_{-1}$ exchange. The compositions of

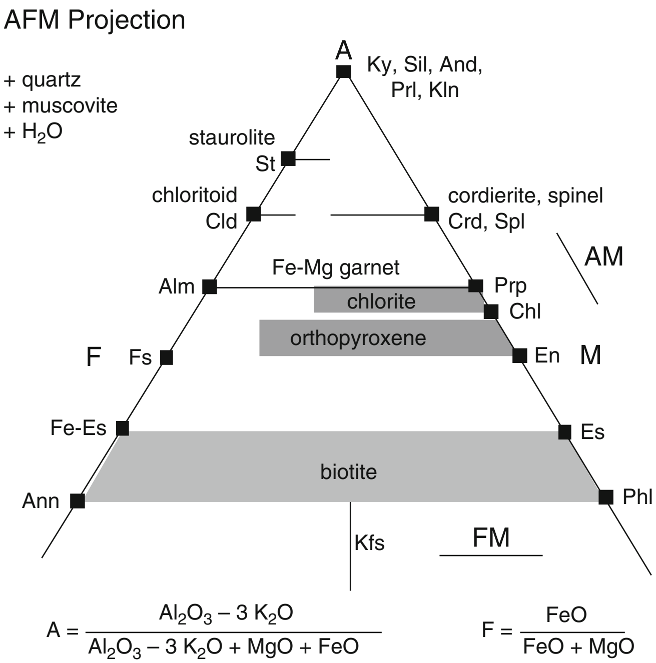

Fig. 2.11 Coordinates of phase compositions in the K_2O–FeO–MgO–Al_2O_3–SiO_2–H_2O system projected through muscovite, quartz and H_2O onto the plane Al_2O_3, FeO and MgO (AFM projection)

these minerals are represented by fields in an AFM diagram. The AFM projection (Fig. 2.11) can be used exclusively for rocks with excess quartz and muscovite and for water-present conditions. Similar projections can readily be prepared from the composition matrix in Table 2.7c if one wants to analyze phase relationships for rocks containing excess muscovite, quartz and Al-silicate but treating H_2O as a compositional variable on the diagram. Such a WFM diagram can be constructed by deleting the rows Ky, Qtz and Ms and renormalizing the remaining three rows. A QFM diagram (Qtz–FeO–MgO) projecting through Ms, Ky and H_2O can be constructed from Table 2.7c for discussing rocks with excess Al-silicate and muscovite but not excess quartz. To project from K-feldspar rather than muscovite, the Ms column in Table 2.7a has to be replaced by the column vector of K-feldspar. The subsequent procedure is the same as that described above. If, for example in low-grade rocks, muscovite contains a significant tschermak component (phengite), one may replace the Ms column vector (endmember muscovite) in Table 2.7a by the analyzed mica composition.

Figure 2.11 shows the general projection coordinates of AFM phases. However, composition phase diagrams are used to represent phase relationships at a given pressure and temperature. A specific example of an AFM diagram at a distinct P and T is shown in Fig. 2.12. Present in excess is quartz, muscovite and H_2O. The AFM surface is divided into a number of sub-regions. Three types can be distinguished:

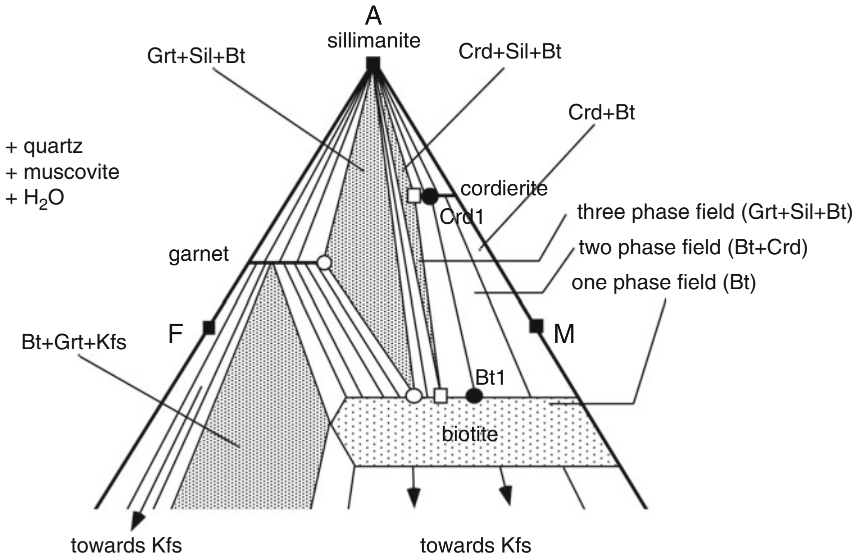

Fig. 2.12 AFM-diagram at a specified pressure and temperature showing phase relationships among some typical AFM phases

1. *One-phase fields*. If the total rock composition projects inside, for example, the one phase field for biotite, it will be composed of quartz, muscovite and biotite. The composition of biotite is given by the composition of the rock. One-phase fields are divariant fields because the composition of the mineral freely changes with the two composition variables of the bulk rock.
2. *Two-phase fields*. If the total rock composition projects inside the two-phase field biotite + cordierite it will contain the assemblage Ms + Qtz + Crd + Bt. The composition of coexisting biotite and cordierite can be connected with a tie line (isopotential line) passing through the projection point of the bulk rock composition (e.g. Bt1–Crd1 in Fig. 2.12). A tie line can be drawn through any bulk composition inside the two-phase field which connects the two coexisting minerals (tie line bundle). It follows that all bulk rock compositions on a specific Crd–Bt tie line (at a given P and T) contain minerals of identical composition. If the rock composition projects close to cordierite it will contain abundant cordierite and little biotite (and vice versa). The mineral compositions are controlled by the Fe–Mg exchange equilibrium between cordierite and biotite (see Chap. 3). Also the tschermak exchange in biotite is controlled by the assemblage. Consider, for example, a rock composition which projects in the garnet + biotite field. For a given biotite composition, the composition of the coexisting garnet is fixed by the tie line passing through the given biotite composition and the projection point of the bulk rock.
3. *Three phase fields*. If the rock composition projects inside the three-phase field biotite + cordierite + sillimanite it will contain the assemblage Ms + Qtz + Crd + Bt + Sil. Here the compositions of all minerals are entirely controlled by the

assemblage (open squares in Fig. 2.12). All bulk rock compositions which project in the three-phase field Crd + Bt + Sil will be composed of these three minerals of identical composition in modal proportions depending on the composition of the rock. Three-phase fields are invariant fields at constant P and T, because the mineral compositions do not vary with rock composition. The cordierite coexisting with biotite and sillimanite has the most Fe-rich composition of all cordierite at the P–T conditions of the AFM diagram. Note that, in general, all three-phase fields on an AFM diagram must be separated from each other by two-phase fields. In the three-phase field Grt + Sil + Bt, the mineral compositions are controlled by the assemblage as well (open circles in Fig. 2.12). It follows that biotite in this assemblage must be more Fe-rich than biotite in the assemblage Crd + Sil + Bt. The garnet coexisting with Bt + Sil has the most Mg-rich composition of all garnets at the given pressure and temperature.

AFM-type diagrams will be extensively used for discussing metamorphism in metapelitic rocks in Chap. 7.

2.5.3.5 ACF Projections

Composition phase diagrams for assemblages in marbles, calcsilicate rocks, calcareous metapelites and other rocks with calcic phases such as mafic rocks, can be prepared, for example, for a simple CFMAS system. Calcic phases may include amphiboles, plagioclase, epidote, diopside and carbonate minerals. A graphic representation of the phase relationships can, for instance, be made by projecting from quartz (if present in excess) and from a CO_2–H_2O fluid phase of constant composition onto the mole fraction triangle Al_2O_2–CaO–FeO (ACF diagram, Fig. 2.13). The ACF diagram is also a projection parallel to the $FeMg_{-1}$ exchange vector. All minerals of the AFM system can also be represented on ACF diagrams provided that one also projects through muscovite or K-feldspar. The coordinates of mineral compositions can be calculated as explained above for the MSH and KFMASH systems, respectively. Figure 2.14 is a typical ACF diagram, showing phase relationships at a certain pressure and temperature. The tschermak variation is parallel to the AF binary and affects mainly chlorite and amphibole (and pyroxene which is not present at the *P–T* conditions of the figure). The three-phase fields Pl + Am + Grt and Ky + Grt + Pl are separated by a two-phase field Grt + Pl because of $CaMg_{-1}$-substitution in garnet (grossular component). Both garnet and plagioclase show no compositional variation along the TS exchange vector. Any mineral which can be described by the components K_2O–CaO–FeO–MgO–Al_2O_3–SiO_2–H_2O– CO_2 can be represented on an ACF diagram such as Fig. 2.13. However, the consequences of $FeMg_{-1}$ substitution in minerals cannot be discussed by means of ACF diagrams. Therefore, any discontinuous reaction relationship deduced from an ACF diagram is in reality continuous and dependent on the Fe–Mg variation (if it involves Fe–Mg minerals). For example, replacement of the Pl–Grt tie line by a more stable tie line between kyanite and amphibole can be related to the

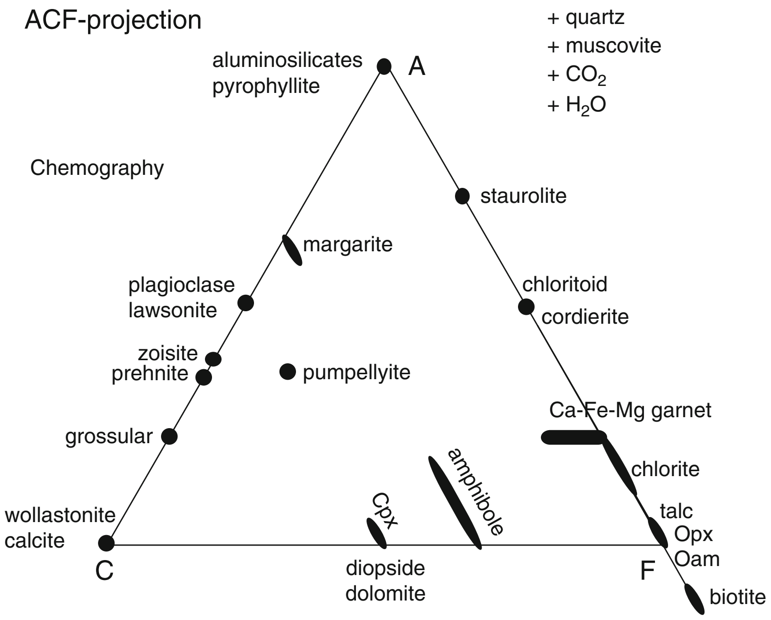

Fig. 2.13 Coordinates of phase compositions in the (K_2O)–CaO–FeO–MgO–Al_2O_3–SiO_2–H_2O–CO_2 system projected through (muscovite), quartz, CO_2 and H_2O and parallel to the $MgFe_{-1}$ exchange vector onto the plane Al_2O_3, FeO and MgO (ACF projection). *Upper* coordinates of some ACF phase compositions; *lower* phase relationships among some typical ACF phases at a specified pressure and temperature

reaction: Pl + Grt ⇒ Am + Ky ± Qtz ± H_2O. Equilibrium of the reaction, however, depends not only on pressure and temperature but also on the Fe–Mg variation in garnet and amphibole. The projection coordinates of a selection of mineral compositions is shown in the chemograph above Fig. 2.13.

In Fig. 2.14, the same mineral chemography is shown together with two protolith compositions listed in Table 2.3. As the *P–T* conditions shown on Fig. 2.14 are those of the amphibolite facies (Barrovian kyanite zone) conditions, it can be seen that metamorphosed MOR basalt, would consist of Pl + Am + Grt and would be rich in amphibole. The platform carbonate plots in the Cal + Am + Zo triangle.

2.5.3.6 Other Projections

Any other graphic representation of phase relationships on composition phase diagrams can be prepared by the procedure outlined above. The type of graphic representation of assemblages is entirely dictated by the material one is working with and by the problem one wants to solve. The following steps are a guide to the production of adequate phase composition diagrams.

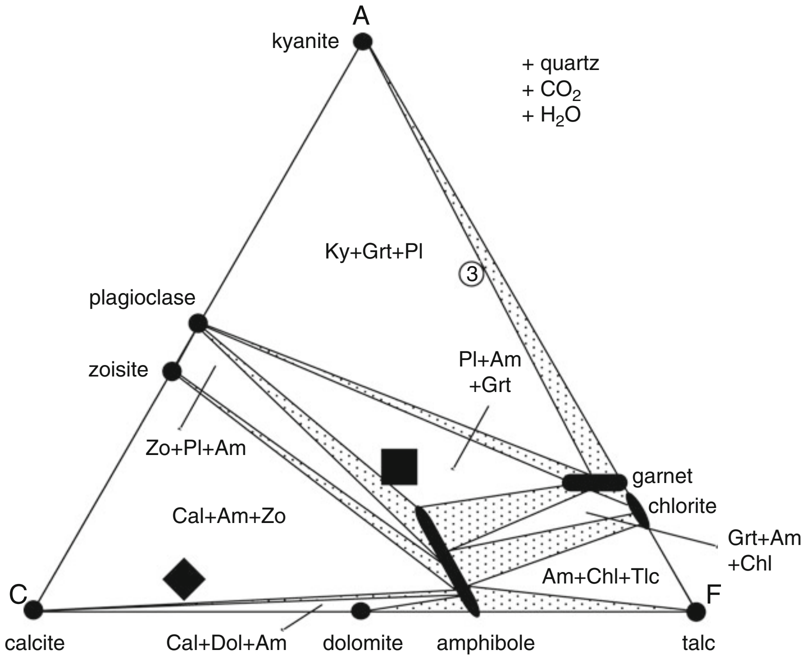

Fig. 2.14 ACF projection of phase relationships of typical ACF phases at a specified pressure and temperature (same as lower diagram in Fig. 2.13). MORB composition projected at *black square*, platform carbonate at *black diamond* (rock compositions listed in Table 2.3)

1. Group the collected rocks in populations with similar bulk compositions, e.g. "normal" metapelites, metabasalts, etc.
2. Identify minerals which are present in the majority of a given group of rocks (e.g. muscovite and quartz in metapelites). Special assemblages require a special treatment, e.g. in quartz-absent corundum-bearing metapelites one may project through corundum onto an QFM plane.
3. If the excess minerals are not composed of simple oxide components, rewrite the composition matrix in terms of the compositions of the excess phases and the desired compositions at the corners of the mole fraction triangle selected as the new projection plane.
4. Delete rows in the composition matrix containing the excess phases and renormalize the column vectors. Draw the diagram and keep in mind the proper distribution of one- two- and three-phase fields. Do not forget to write the compositions of the projection phases on the diagram (without this information your figure is meaningless).

In Part II we will make extensive use of various kinds of composition phase diagrams.

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