1 Introduction

1.1 What Is Rock Physics?

The meaning of the term assumes that "rock physics" is an application of physical methods to the study of rock properties. From the geological and mineralogical point of view, rocks may be distinguished by their macroscopic properties studied in the field and by the microscopic properties studied by mineralogists and petrologists in labs. Rocks also possess some very variable physical properties, such as density, elastic modulus, permeability, porosity, magnetic susceptibility, resistivity, etc., just like any other solid material. From the geophysical point of view, rocks are an environmental medium whose properties need to be known in order to provide an adequate interpretation of geophysical measurements. Thus, *petrophysics* or rock physics is a link between the branches of geoscience knowledge, such as geophysics, lithology, petrography, hydrogeology and rock mechanics.

Rock, in geology, is the general term for any naturally occurring mineral aggregate with one or more mineral phases. Rocks build up not only the Earth's crust and mantle but also celestial bodies such as terrestrial planets, moons and asteroids. Rocks are traditionally subdivided into sedimentary rocks (sedimentites), magmatic rocks (plutonites, volcanic effusive and intrusive rocks) and metamorphic rocks after their typical formation process. *Sediments* are formed by deposition of weathering or erosive material of other rocks or by chemical or organic deposition processes. *Magmatic rocks* solidify from magmas, that is, from melts or partially molten rocks at depth or on the Earth's surface. *Metamorphic rocks* form in the Earth's crust after transformations of other types of rocks under the influence of high temperatures and pressures or in contact with fluid phases. According to the degree of mechanical consolidation, which means the way of combining and binding individual components of rocks, a distinction is also made between loose or soft rocks (e.g. sand) and hard rocks (e.g. sandstone). For example, ores are rocks in which metals have been accumulated. If they are available in economically sufficient concentrations and quantities, they are called deposits.

The Earth, as well as other terrestrial planets and many celestial bodies, consists of rocky shells and a core. Approximately 5 billion years ago, it was time for a rotating disk-shaped body to emerge from a cosmic molecular cloud consisting of volatile elements as well as of heavier dust particles. This may have been caused by the explosion of a nearby star (called a supernova [SN]). The explosion waves caused a squeezing of the cosmic cloud, which started to collapse. The gas and dust pulled together due to gravity and formed a solar nebula. Density and temperature inside the nebula increased, as did the spinning velocity. The energy of contraction in the center led to the ignition of nuclear fusion in the proto-Sun. Solar wind and centrifugal forces blew the light elements of hydrogen and helium away from the Sun, while the heavy ones remained close to the Sun. The innumerable individual bodies in the solar system came together in orbits and formed the inner rocky and outer gas planets over the next billion years. According to one of the plausible core accretion scenarios, the cores of protoplanets have to achieve critical masses, and only after that are the planets able to retain a gas atmosphere. This theory is a good explanation of the origin of the terrestrial planets (the four planets closest to the Sun –

Mercury, Venus, Earth and Mars) but not the gas giants (Jupiter, Saturn, Uranus and Neptune). The difficulties in attributing the formation of the solar system directly to a single SN explosion are as follows (Pfalzner et al., 2015). (1) Prior to SN explosion, stars having various starting masses undergo different stages of radionuclide burning, ending in explosions in stages from hydrogen, through helium–carbon–neon–oxygen, to silicon. Each explosion burning fingerprints a specific pattern of element abundances, which differ from the solar system abundances (Arnett & Clayton, 1970). (2) The interstellar medium around a massive star prior to SN explosion is depleted of molecular clouds. It is very unlikely to find a protoplanetary disk or a dense core that close to an SN. (3) The composition of meteorites (chondrites, achondrites, and magmatic and nonmagmatic iron meteorites) indicates the past presence of short-lived radionuclides with half-lives ($\tau_{1/2}$) $\lesssim$ 10 Ma. The age of differentiated meteorites that experienced a heating effect from these radionuclides, largely caused by ^{26}Al decay, triggering their differentiation into metal core and silicate mantle, is only 1–4 Ma after the formation of the solar system. The age of nondifferentiated meteorite parental bodies is only 1–2 Ma greater. (4) The abundances of the short-lived radionuclides ^{26}Al and ^{60}Fe and their ratio in the solar system are higher than the theoretical estimates from a silicon SN explosive bursting (Arnett & Clayton, 1970). An alternative theory, which tries to explain these contradictions, suggests that the Sun is one of a second generation of stars, whose formation was mediated by the “runaway” of a massive star due to an SN explosion. The most likely origin for ^{26}Al is local; that is, the wind of a single massive star enriched a shell around it, and the subsequent gravitational collapse of the shell resulted in the formation of a new ^{26}Al-rich star and other stars, including the Sun. The origin of ^{60}Fe is due to the enhanced galactic background as a result of stars being born and dying in the same molecular cloud as the Sun but in a prior generation. The protoplanetary disk of the massive star, in which dust and gas were already bound together in clumps in the early stages of the solar system’s development, existed before the Sun’s formation. The clumps slowly compacted and formed the giant planets. Gas planets can be formed faster than terrestrial ones; it may take about 1,000 years to trap gases from the disk. The orbits of massive gas planets were rapidly stabilized, preventing them from falling down back into the Sun. A new view of the origin of the solar system and the planets suggests that solid materials or proto-rocks existed at an early stage in the form of *chondrules* and *calcium-aluminum rich inclusions* (CAIs; Figure 1.1a). CAIs were formed from molten gas droplets at high temperatures (>1,300 K), whereas chondrules were the collected dust clumps melted and rapidly cooled at lower temperatures (<1000 K). The fast spinning of material caused the disk to flatten, unevenly distributing these two sources of proto-rocks into planes along the protoplanetary disk. Local heating and remelting of cosmic proto-rocks may be caused by shock waves after material collapses onto the protoplanetary disk. Recently, it was suggested that rocks existed even at an early stage of terrestrial planet formation due to melting of chondrules and CAIs caused by spikes of electrical current in the protoplanetary disk and subsequent rapid cooling (Howell, 2012).

Metallic and nonmetallic elements had differing evolution in the solar nebula. Metallic elements condensed almost as soon as the spinning disk accreted (4.55–4.56 billion years ago according to isotope measurements of certain meteorites), and the solid nonmetallic material condensed a bit later (between 4.4 and 4.55 billion years ago). As a consequence, different kinds of meteorites were formed, and their participation in the bombardment of protoplanets resulted finally in distinct shell structures of the planets.

1.2 Shell Structure of the Earth

The age of the Earth is $4.54 \pm 0.05 \times 10^9 a$, based on the evidence from radiometric age-dating of meteorite materials, which is consistent with the radiometric ages of the oldest known terrestrial and lunar rock samples. Thus, at this age the first separation of substances had been completed, in which rock-forming elements such as silicon, iron, aluminum and oxygen were delivered to the overall

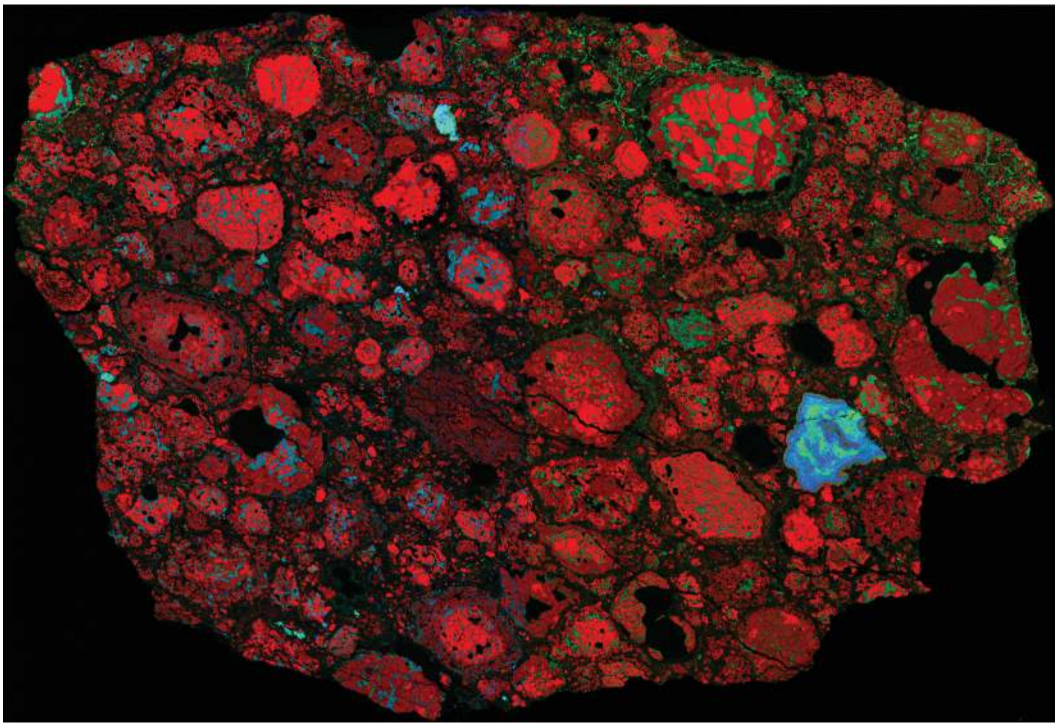

Figure 1.1a Proto-solar system rock material: combined elemental X-ray map of CR (Renazzo type) carbonaceous chondrite Northwest Africa 801. Legend: Mg (red), Ca (green) and Al (blue). The size of the largest CAI is about 400 μm. Round, almost spherical-shape fragments are chondrules, CAI are of more irregular form, and the space between chondrules and CIA is filled with matrix material. (Courtesy of A. Krot, University Hawaii.)

budget. With the aggregation of *planetesimals* (orbital cosmic objects with a diameter of about 1 km), the first celestial bodies in the solar system were formed, and terrestrial planets like the Earth grew to their present size. The birth of the Earth was accompanied by intensive meteorite impacts for 70–80 million years. As the Earth's melting progressed, intense magmatism and a magma ocean developed. This ensured that light silicate-rich material was transported toward the surface and heavier iron-rich material sank down. From an initially homogeneous protoplanet emerged a strictly differentiated body with a spherical shell structure (Figure 1.1b). Volatile components were transported to the outside and escaped the surface, forming an early proto-atmosphere. The solid part of the Earth differentiated further and, as in a blast furnace, a layer of slag was formed above the metal core, which mainly contained iron. This slag layer was quite powerful and massive, and included the Earth's mantle and crust. Basically, these slag shells consisted of silicates – the Earth's rocks.

With the emergence of the Earth's shell structure a further separation of matter continued, by which some specific light elements have been accumulated in the crust, the so called *lithophile elements* (λίθος means stone in Greek), whereas others disappeared almost completely inside the Earth's core, the so called *siderophile elements* (σίδερο = iron).

Only 8 out of about 100 natural elements account for over 99 percent of the Earth's crust composition (Figure 1.2). All heavy elements tend to be stuck in the Earth's interior beyond human reach and are found only in rare cases in the crust. Such elements are "noble" to us and are eagerly sought as ore deposits. This starts with iron, the Earth's most abundant element, which forms only 6 percent of the crust by weight.

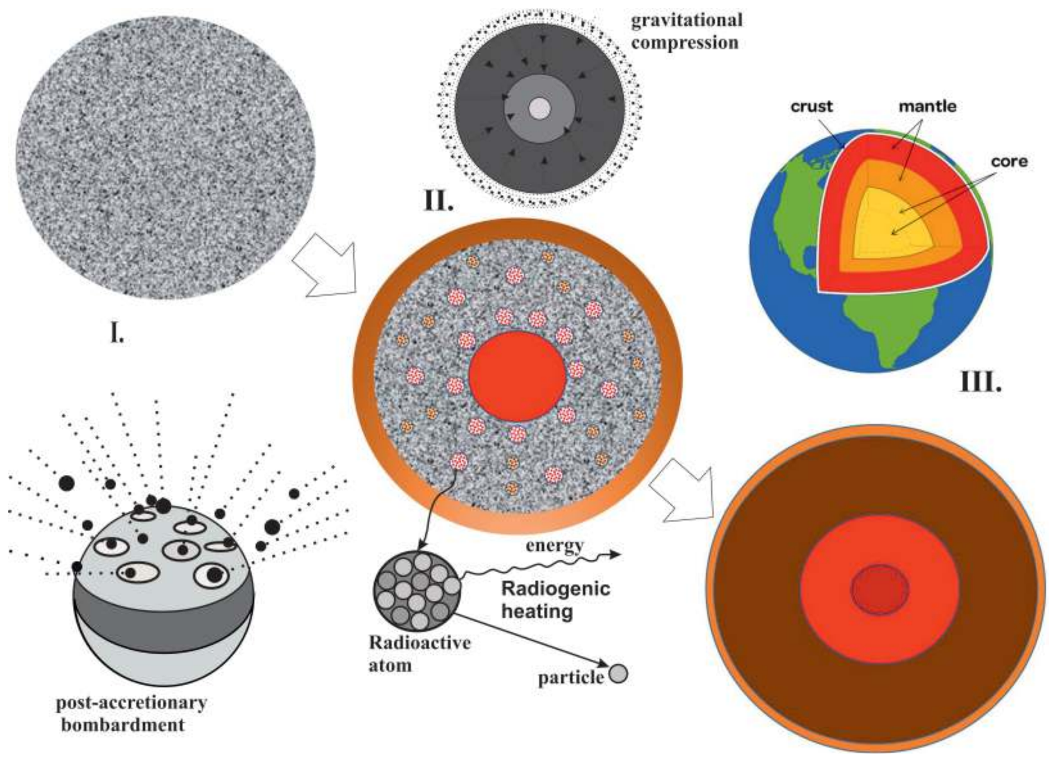

Figure 1.1b Shell structure of the Earth. Starting as a homogeneously accreted body, the proto-Earth developed into a differentiated shell-structured planet with an iron-nickel core and a silicate-rich mantle. The uppermost part of the silicate shell, where plate tectonics is active, represents the lithosphere ("foam" of the Earth), a realm of rocks.
I. Homogeneous accretion stage: Heat builds up as a planet accretes due to meteorite bombardment; II. Differentiation stage: A core is formed by Fe-Ni alloy melting, accompanied by other chemical transformations, and heat is produced due to gravitational energy release and material contraction under pressure in the center as well as by radiogenic disintegration of nonstable isotopes; III. The mantle overturn during the core formation is over: solid inner core 5,150–6,370 km, liquid core 2,891–5,150 km, silicate mantle 40–2,891 km and crust 5–40 km are built.

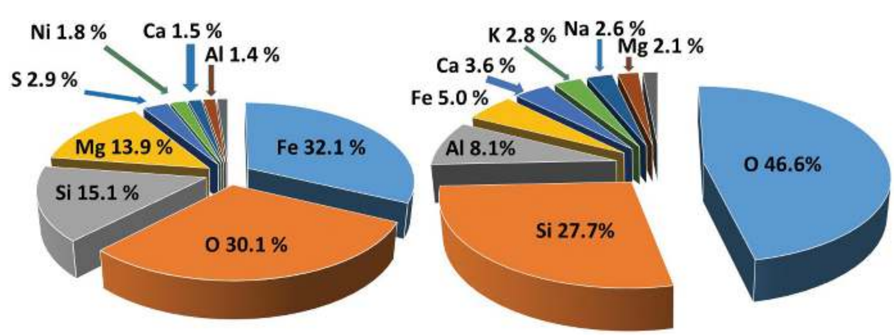

Figure 1.2 Chemical element distribution in the bulk Earth and in the crust, weight %. (a) represents the bulk Earth composition; (b) is the crust composition. (Replotted from Sebastian, 2009)

1.3 Rock Properties and Geophysical Methods

The basic concept of geophysical measurements is to categorize the structure of the Earth or other terrestrial planets and to discriminate the materials composing them. Thus, geophysical methods should be sensitive enough to reveal physical anomalies in rocks under varying temperature and

Table 1.1 Comparison of petrological parameters and some applied geophysics methods

Geophysical procedure	Measured petrological parameter
Gravimetric methods: Measurements of differences, gradients and anomalies of gravitational fields of the Earth and planets	Density: Measurements of density or variations of density
Seismic and seismology methods: Using natural and controlled generation of elastic waves in order to obtain an image of subspace or subsurface	Propagation velocity V and absorption coefficients of elastic waves, reflection, transmission and absorption coefficients of mechanical discontinuities
Geoelectric resistance methods: Use of controlled electrical potential source to detect, discriminate and map conductive, nonconductive and polarizable subspaces and subsurfaces	Specific electrical resistance ρ: Measurements of electrical impedance of bulk rocks and of mineral grain boundaries under controlled environmental conditions
Geomagnetic methods: Measurements of the Earth's magnetic field intensity and its components	Magnetic susceptibility κ: Measurement of magnetization and demagnetization as a function of the content of individual minerals in rocks
Electromagnetic methods: Use of controlled or natural source of time-varying electromagnetic fields exciting electrical currents in the Earth through electromagnetic induction	Dielectric constant ε: Characterization of external electromagnetic field interaction with electric dipole moment in rocks or minerals, quantification of storage and dissipation of electromagnetic field in geomaterials
Geothermal methods: Measurements of heat flux, geothermal gradient, surface temperature and emitted radiation	Thermal conductivity λ, heat capacity C_p, heat generation A: Transient and storage heat measurements, radioactive heat generation in rocks
Radiometric methods: Natural spectral γ-measurements, radon emanations, borehole neutron transmission measurements, air-borne and space-borne radiometry	Activity of natural radiation: Radiometric-spectroscopy, fissure tracks in rocks, radioactive isotope concentration, age-dating measurements
Nuclear physics methods: Spectrometry and accumulation of total counts of natural radioactivity and cosmic radiation, registration of X-ray, γ-particle scattering, measurements with neutron sources and positron emission tomography in boreholes	Parameters that govern interactions of radiation with rocks: γ-fluorescence, neutron–particle reactions, neutron backscattering; nuclear magnetic resonance in rocks

Source: adapted from Schön, 2011.

pressure, and at varying degrees of fluid phase saturation. This puts a certain condition on the methods in use, which is called *resolution*, or the achievement of the maximum degree of structural and material detail at the lowest distance between spatially separated objects. In this case, knowledge of the physical properties of rocks is a prerequisite. The geophysical methods aiming to measure particular physical parameters are listed in Table 1.1.

The physical background of each of the measured petrophysical parameters in the right-hand column of Table 1.1 will be presented in the following chapters of this book.

1.4 Texture and Fabric of Rocks

The texture of rocks is the sum of all spatial data inside a considered area; that is, the texture includes above all the information on division, sequence and spatial arrangements of constituent parts of rocks. The individual, contrasted, larger-scale features of rocks are called "structures." The term "fabric" encompasses the terms "structure" and "texture"; it can be considered as an element of structure on a macro-scale or an element of textures on a micro-scale. Under the term "texture," clearly one understands the spatial arrangement of crystal building blocks (ions, atoms, defects) and their reproduction in a certain geometric arrangement. In some cases the term "texture" is used for sedimentary and igneous rocks, meaning a directional feature, whereas with metamorphic rocks the term used is a "fabric." Fabrics often mean observable characteristics of rocks, which can be studied in outcrop hand specimens and under a microscope. Thus, rock fabrics refer to the mutual relationship between grains. In contrast, rock textures refer to small-scale features in a rock, including grain size and grain shape, intergranular relations, and degree of crystallinity (amount of crystal and glass). Hard rocks are usually characterized by a *holocrystalline texture*, in which all the rock material is crystallized. Holocrystalline rocks are subdivided into a group having *phaneritic texture*, in which the minerals may be distinguished with the naked eye, and a group with *aphanitic texture*, in which the minerals may be discerned only under a microscope. Holocrystalline textures are subdivided according to their grain size into *fine-grained* (crystal size<1 mm), *medium-grained* (1–5 mm), *coarse-grained* (5–10 mm) and *very coarse-grained* (grain size >10 mm). Textures also depend on the shapes and habitus of constituting minerals. In some cases, minerals have ideal growth according to their crystallographic axis (*habitus*) and form *idiomorphic crystals*, while in other cases ideal mineral growth is suppressed and their shapes are distinct from an idiomorphic one; these are called *allotriomorphic* or *xenomorphic*. Minerals may be idiomorphic in relation to some sorts of minerals (their shape is closer to the ideal) and xenomorphic (their shape is more distinct from the ideal) in relation to others. The idiomorphic and allotriomorphic crystal shapes are recognized by two distinct morphologies, according to whether they grow by a diffusional mechanism from melt or from solution: (1) allotriomorphic crystals nucleate at the grain boundaries, while (2) idiomorphic crystals usually form intragranularly. Toughness and other mechanical properties of rocks are strongly affected by their microstructure. Microstructures formed by crystals nucleated at preexisting grain boundaries increase the resistance to cleavage crack propagation. If the shapes of all crystallized minerals are close to idiomorphic, the textures of hard rocks are described as *panidiomorphic-granular* (pyroxenites, peridotites, dunites). Textures that are represented by a combination of minerals having differing degrees of idiomorphism are called *hypidiomorphic-granular* (granites, syenites, diorites). Thus, rocks composed of *euhedral crystals*, those that are well formed with sharp, easily recognized faces, are opposite to rocks with *anhedral crystal* texture, that is, rocks with mineral grains having no well-formed crystal faces or cross-section shapes in thin sections. An example is the simultaneous crystallization from melt of feldspar and quartz crystals in granites, which produces *pegmatitic* or *graphic* textures with intergrowths of minerals. Depending on the size distribution of minerals, a distinction is made between *equigranular* and *inequigranular textures* (Figure 1.3).

Individual minerals composing rocks may or may not be bounded by their own crystal faces. Those that are bounded are termed *idioblastic*, while those that are not bounded are called *xenoblastic*. Different minerals exhibit specific tendencies to be idioblastic. A high tendency to be idioblastic is termed *porphyroblastic*. A distinction is made between uniformly granular (*homeoblastic*) and nonuniformly granular (*heteroblastic*) textures. A special case of heteroblastic texture is seen in the porphyroblastic texture, characterized by the presence of large mineral crystals (porphyroblasts) within a fine-grained mass of rock.

Depending on the volume ratio of glass to small crystals (*microlites*), the groundmass texture of volcanic rocks is classified as glassy (or *vitrophyric*), semicrystalline (for example, *hyalopilitic* texture) or *microlitic*. The minerals in metamorphic rocks are classified according to the shape of grains as *granoblastic* or granular (quartzites, marbles), *lepidoblastic* or foliated, a category that is characteristic of rocks containing mineral grains with foliated forms (mica schists, phyllites), or *lepidogranoblastic* or granular-foliated. In metamorphic rocks retaining relicts of the initial rock texture, the texture receives the name of *primary texture* with the prefix "blasto-," for example, *blastoporphyritic* and *blastopsammitic* (ψαμμίτης is sandstone in Greek).

Figure 1.3 Classification of main rock textures: a – crystalline; b – clastic-round grain, c – clastic-angular; d – cubic; e – columnar; f – parallelepiped; g – uniform; h – continuous nonuniform; i – discontinuous nonuniform; k – porphyritic; l – conglomerate; m – breccia.

1.5 Structure of Rocks

Structure identifies the individual parts of rocks in terms of their size, their shape and, if necessary, their crystal development and spatial arrangements. If there are no individual parts available, not even any that are discernible with optical aids, for example, recognizable under a magnifying glass or microscope, that rock is called *amorphous* (examples are volcanic and impact glasses, such as obsidian, pseudotachylites, etc.). At the beginning of each determination and quantification of rock structure, a statement should be made about the grain size of any individual mineral component. The first category of hard rock structures is massive or *homogeneous structures*, in which minerals are uniformly distributed throughout rock mass that has approximately the same mineral composition and texture throughout a rock. *Heterogeneous* (*taxite*) structures are also very common, for example, banded and fluidal structures with minerals oriented in a particular fashion that have arisen under conditions of movement or deformation of melted, partially melted or ductile rocks. Taxite structures may result from a nonuniform distribution of minerals (hornblende, biotite) or from an alternating segment arrangement of different granularity.

Structures of hard rocks that have experienced deformation under flow are classified as massive or *fluidal* or as structures exhibiting *flow banding*, which results from a parallel arrangement of variously colored bands, for example, of volcanic glass, phenocrysts and microlites. Depending on the quantity of gas bubbles in lavas, distinctions are made between *porous*, *vesicular* and *pumiceous* structures (pumices, reticulates). *Amygdaloidal* structures are formed by the filling of cavities with secondary minerals (quartz, opal, zeolites, carbonates). Some metamorphic rocks demonstrate a structure called *foliation*, caused by a preferred orientation of sheet silicate minerals. Rocks that show *banded structures* without a distinct foliation are termed *gneisses*. Rocks that show no foliation are called *hornfels*, if the grain size <1 mm, or *granulites*, if the grain size is >1 mm (see Table 1.2). In metamorphic rocks, *flaccid* (shattered and disintegrated as a result of a paleo landslide or earthquake) and *schistose* (mostly due to lineation of schist minerals along specific schistosity planes) structures occur (Figure 1.4).

The structure of clastic rocks depends on the mutual arrangement of grains, which can be *random*, *laminar* or *fluidal*. In random structures, particles do not have any ordered arrangement. This structure is characteristic of very coarse-grained rocks, such as gravel, shingle and sands, but it may also be present in certain fine-grained rocks. Random structures arise in regions of sedimentation characterized by an abundant and continuous supply of homogeneous detrital material or by a continuous rolling of particles. Laminar structures are formed by the alternating arrangement of layers, either of rocks having varying mineralogical compositions or of chemogenic layer-forming rocks (anhydrites, gypsums, rock salt, potassium salts). In laminar structures, the individual layers are distinguished from one another not only by composition but also by grain size. Fluidal structures result from deformations of the original laminar structure or the influence of water flows and landslides, or an intensive wave action, or a collapse of laminar structures. In chemical sediments, *oölitic structures* are very common, characterized by rounded grains or grain aggregates; these are typical for carbonate rocks (limestones, dolomites), iron, manganese, and phosphate ores and bauxites. Biogenic sediment structures in some cases result from organism build-up, the so-called *growth structures*. Structures of this type are

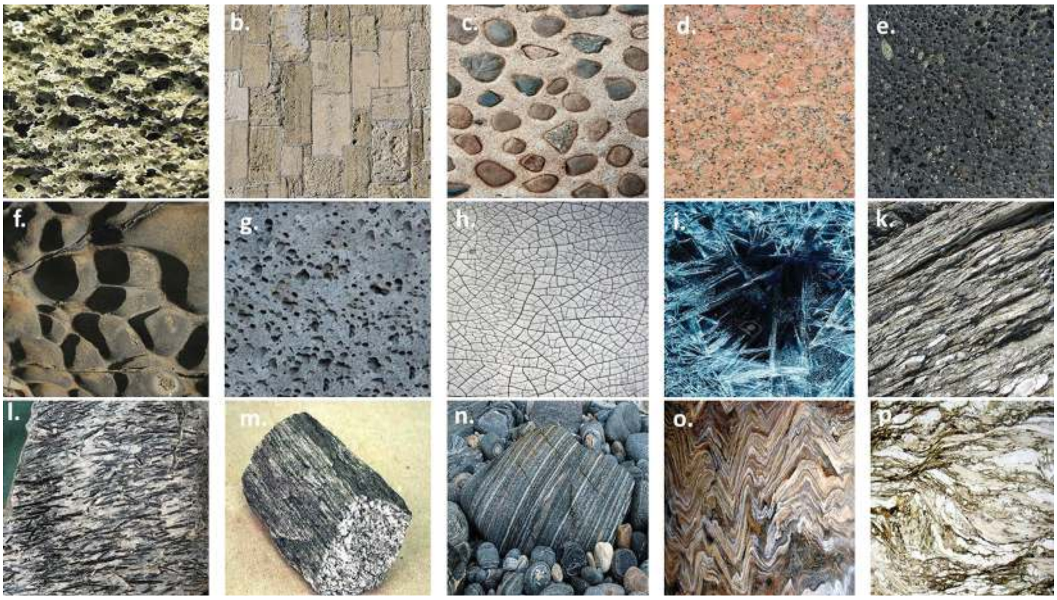

Figure 1.4 Principal rock structures classified according to the type of grain contacts (a–c), void space geometry (d–h), fabric orientation (i–m) and flatness of foliation (n–p): (a) cemented pore space; (b) cemented grain contacts; (c) cemented basal bandage (structure where grains float in cement); (d) dense isotropic; (e) porous; (f) cavernous; (g) vesicular; (h) fractured; (i) chaotic orientation of fabrics; (k) parallel planar arrangement; (l) dimensionally linear fabric; (m) linear fabric; (n) flat foliation; (o) wavy foliation; (p) shear band S-C fabric.

characteristic of corals, bryozoans ("moss animals"), calcareous algae and hydractinians (gastropod shells). The build-up of organisms produces either a flat body lying along the bottom with a gently undulating surface (*stromatolite*) or a small, oval mass resembling concretions (*oncolite*). Bodies that grow in the shape of hills or high knolls are called *bioherms* (ancient organic reefs with a predominance of fossil calcareous algae). Coral reefs are usually a combination of stromatolites (columns formed by the growth of layer upon layer), oncolites (layered spherical structures) and bioherms (shapes built by marine invertebrates).

Rocks whose mineral grains are larger than 0.2 mm are grained rocks (Tables 1.2 and 1.3). In this case, it must be distinguished whether mineral grains are predominantly monodispersed or are more or less nonuniform in size. In addition to size, the shape of individual grains and minerals in rocks is an important diagnostic feature of their physical properties. Shapes of crystals newly formed and grown from solutions or melts, or from recrystallization, are determined by the shape of their neighbors. In any case, they are straight-edged or irregularly angular on margins. In contrast, mechanically transferred or transformed grains or fragments of rocks are rounded more or less well, depending on the original grain shape, transport distance or degree of strain, and hardness of the material. This is the case of many sedimentary rocks, due to the significant influence of their formation environment. Grain binding in rocks is also used to determine their mechanical properties. One distinguishes between indirect and immediate grain binding. In the case of indirect grain binding, individual mineral grains are cemented by a – usually crystalline – binder (cement). This can be of the same material as the grains themselves, or it may be a foreign material. Indirect grain binding is mainly observed in sedimentary rocks. In rocks formed by recrystallization from solutions or from melts, the recrystallization predetermines the immediate grain binding. Single crystals directly border each other here, i.e. without being cemented by an intermediary. Cohesion of rocks is preserved by interfacial forces and by a close interlocking of individual grains. The texture elements are illustrated in Figure 1.5.

1.6 Global Rock Cycle

No rock exists forever, even though in continental formations the rocks have a better "chance of survival" than the oceanic ones. In the course of plate tectonics, the rock circulation has become a cycle: spreading center – subduction – volcanic eruption – sedimentation – subduction. Plate tectonics itself is a continuous rock cycle, as shown in Figure 1.6.

Figure 1.6 elucidates a typical plate tectonic scenario of the *rock cycle* concept. Considering the example of a magmatic belt above subduction zones, it is shown how the three formation processes for magmatic, sedimentary and metamorphic rocks are related to each other. The most important processes that move rocks from one to the next group are as follows: weathering, subduction and partial melting–remelting.

1.7 Basic Description of Petrophysical Parameters

Rocks are composed of mineral grains, porous space and a medium filling it. Thus, essentially, rocks are heterogeneous mineral mixtures. Nevertheless, one assumes macroscopic homogeneity of rocks and uses so-called mean values or "effective physical parameters" to characterize their properties. Physical properties in general are tensors having a certain rank, or a number of simultaneous directions, in which this property is defined and may be measured. For example, density is a scalar or a tensor of rank 0, velocity and force are *vectors* or *tensors* of rank 1, thermal conductivity, stress or strain are tensors of rank 2, and the elasticity coefficient tensor is a tensor of rank 4. So, in physics, a tensor variable characterizes the properties of a physical system depending on spatial directions,

Figure 1.5 (a) Fabric or texture of granite rocks as an example of plutonic rocks. Even-grained crystals without preferred orientation: feldspars are uniform yellow; mica shows close parallel black plates; white in the remaining space is quartz. Especially, feldspars form idiomorphic crystals, while quartz is xenomorphic. (b) Fluidal texture in volcanic rocks. Crystals follow in their arrangement the flow pattern of melt. (c) Porphyric structure of igneous rocks. Crystals float in a fine-grained crystalline matrix. Large crystals had already been formed within the magma chamber. After a rapid deflation of the chamber, they were transported upwards together with melt during a volcanic eruption; this itself causes fast cooling and, therefore, results in a finely crystalline or even glass-like solidified host matrix. (d) Structure of clastic sediments. Quartz is rounded and white, mica is of an almost parallel orientation and dark, while feldspar is grey. (e) Parallel texture of gneiss as an example of metamorphic rocks. (f) Texture in eye gneiss. (g) Puzzle-like arrangement of calcite crystals in metamorphic marble rock showing twinning planes. (h) Breccia texture consisting of different rock fragments randomly cemented or compressed together inside finer-grained host material.

and the rank of the tensor is the measure of its spatial directions. For example, vectors and complex numbers are tensors of rank 1, piezo-electric coefficients are tensors of rank 3, etc. In physics, the definition of tensor rank comes from the number of directions that are involved in the parameter definition. For example, mass has only a magnitude and zero directions (rank 0), force and velocity possess magnitude and direction (rank 1), stress and strain are characterized by a magnitude and two directions (rank 2), magnetostriction coefficients or strains per magnetic