Contents

rock

in geology

, naturally occurring and coherent aggregate of one or more minerals. Such aggregates constitute the basic unit of which the solid Earth is comprised and typically form recognizable and mappable volumes. Rocks are commonly divided into three major classes according to the processes that resulted in their formation. These classes are (1) igneous rocks, which have solidified from molten material called magma; (2) sedimentary rocks, those consisting of fragments derived from preexisting rocks or of materials precipitated from solutions; and (3) metamorphic rocks, which have been derived from either igneous or sedimentary rocks under conditions that caused changes in mineralogical composition, texture, and internal structure. These three classes, in turn, are subdivided into numerous groups and types on the basis of various factors, the most important of which are chemical, mineralogical, and textural attributes.

Overview

The three major classes of rock—igneous, sedimentary, and metamorphic—are based on the processes that formed them. These three classes are further subdivided on the basis of various factors, especially chemical, mineralogical, and textural attributes (*see* e.g., acid and basic rocks; crystalline rock; extrusive rock). *See also* felsic rock; intrusive rock; mafic rock.

Rock types

Igneous rocks are those that solidify from <u>magma</u>, a molten mixture of rock-forming minerals and usually volatiles such as gases and steam. Since their constituent minerals are crystallized from molten material, igneous rocks are formed at high temperatures. They originate from processes deep within the Earth—typically at depths of about 50 to 200 kilometres (30 to 120 miles)—in the mid- to lower-crust or in the <u>upper mantle</u>. Igneous rocks are subdivided into two categories: <u>intrusive</u> (emplaced in the crust), and extrusive (extruded onto the surface of the land or ocean bottom), in which case the cooling molten material is called <u>lava</u>.

Sedimentary rocks are those that are deposited and lithified (compacted and cemented together) at the Earth's surface, with the assistance of running water, wind, ice, or living organisms. Most are deposited from the land surface to the bottoms of lakes, rivers, and oceans. Sedimentary rocks are generally stratified—*i.e.,* they have layering. Layers may be distinguished by differences in colour, particle size, type of cement, or internal arrangement.

Metamorphic rocks are those formed by changes in preexisting rocks under the influence of high temperature, pressure, and chemically active solutions. The changes can be chemical (compositional) and physical (textural) in character. Metamorphic rocks are often formed by processes deep within the Earth that produce new minerals, textures, and crystal structures. The recrystallization that takes place does so essentially in the <u>solid state</u>, rather than by complete remelting, and can be aided by ductile deformation and the presence of <u>interstitial fluids</u> such as water. Metamorphism often produces apparent layering, or banding, because of the segregation of minerals into separate bands. <u>Metamorphic processes</u> can also occur at the Earth's surface due to meteorite impact events and pyrometamorphism taking place near burning <u>coal seams</u> ignited by lightning strikes.

Extrusive rock - any rock derived from <u>magma</u> (molten silicate material) that is poured out or ejected at the Earth's surface. By contrast, <u>intrusive rocks</u> are formed from magma that is forced into older rocks at depth within the Earth's crust; the molten material then slowly solidifies below the Earth's surface, where it may later be exposed owing to erosion. Extrusive rocks are usually distinguished from intrusive rocks on the basis of their texture and mineral composition.

Mafic rock in geology, <u>igneous rock</u> that is dominated by the silicates pyroxene, amphibole, olivine, and mica. These minerals are high in magnesium and ferric oxides, and their presence gives mafic rock its characteristic dark colour. Mafic rock is commonly contrasted with <u>felsic rock</u>, in which light-coloured minerals predominate. Common mafic rocks include basalt and its coarse-grained intrusive equivalent, gabbro.

Felsic rock - division of igneous rocks on the basis of their silica content. Chemical analyses of the most abundant components in rocks usually are presented as <u>oxides</u> of the elements; igneous rocks typically consist of approximately 12 major oxides totaling over 99 percent of the rock. Of the oxides, <u>silica</u> (SiO_2) is usually the most abundant. Because of this abundance and because most igneous minerals are silicates, silica content was used as a basis of early classifications; it remains widely accepted today. Within this scheme, rocks are described as felsic, intermediate, mafic, and ultramafic (in order of decreasing silica content).

Rock cycle

Geologic materials—mineral crystals and their <u>host rock</u> types—are cycled through various forms. The process depends on temperature, pressure, time, and changes in environmental conditions in the
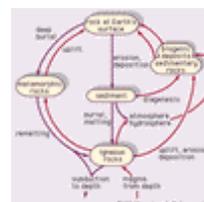
Earth's crust and at its surface. The rock cycle illustrated in Figure 1 reflects the basic relationships among igneous, metamorphic, and sedimentary rocks. Erosion includes weathering (the physical and chemical breakdown of minerals) and transportation to a site of deposition. Diagenesis is, as previously explained, the process of forming sedimentary rock by compaction and natural cementation of grains, or crystallization from water or solutions, or recrystallization. The conversion of sediment to rock is termed lithification.

Abundance of rock types

An estimate of the distribution of rock types in large structural units of the terrestrial crust is given in the Table. The relative abundance of main rock types and minerals in the crust is shown in the Table.

Table 30: Distribution of Rock Types in Large Structural Units of the Earth's Crust

crustal layers	average thickness (in kilometres)	volume (in cubic kilometres)	mass (10^{24} grams)	rock types and abundances (percent volume of layer)
Continental platform Sedimentary	1.8	135	0.35	sands, 23.6; clays, 49.5; carbonates, 21; evaporites, 2; basalts, 3.9
Continental geosynclinal folded belts Sedimentary	10.0	365	0.94	sands, 18.7; clays and shales, 39.4; carbonates, 16.3; evaporites, 0.3; basalts, 12.6; andesites, 10.2; rhyolites, 2.5
Subcontinental, shelf and slope Sedimentary	2.9	190	0.48	similar to above groups
Continental and subcontinental Granitic		3,590	9.81	granites, 18.1; granodiorites, 19.9; syenites, 0.3; gabbro, 3.7; peridotites, 0.1; gneisses, 37.6; schists, 9.0; marble, 1.5; amphibolite, 9.8
(continental)	20.1			
(subcontinental)	9.1			
Basaltic		3,760	10.91	acid igneous and metamorphic rocks, 50; basic igneous and metamorphic rocks, 50
(continental)	20.1			
(subcontinental)	11.7			
Oceanic Layer 1 (sedimentary)	0.4	120	0.19	(percent by area) terrigenous, 7.3; calcareous, 41.5; siliceous, 17; red clays, 31.2
Layer 2	0.6	175	0.44	(percent by volume) sediments, 50
	0.6	175	0.52	(percent by volume) basalts, 50
Layer 3	5.7	1,700	4.92	(percent by volume) oceanic tholeiitic basalts, 99; alkaline differentiates, 1

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Table 31: Abundances of Main Rock Types and Minerals in the Earth's Crust

rocks	percent volume of crust	minerals	percent volume of crust
Sedimentary		**Constituents of sedimentary rocks**	
Sands	1.7	Quartz	12
Clays and shales	4.2	Alkali feldspar	12
Carbonates (including salt)	2.0	Plagioclase feldspar	39
	7.9	Micas	5
Igneous		Amphiboles	5
Granite	10.4		73
Granodiorite, diorite	11.2	**Constituents of igneous rocks**	
Syenite	0.4	Pyroxene	11.0
Basalt, gabbro, amphibolite, eclogite	42.5	Olivine	3.0
Dunite, periodotite	0.2	Clay minerals (and chlorite)	4.6
	64.7	Calcite	1.5
		Dolomite	0.5
		(Titano) magnetite	1.5
Metamorphic		Others (garnet, kyanite, apatite, etc.)	4.9
Gneiss	21.4		27.0
Schist	5.1		
Marble	0.9	**Totals**	
	27.4	Quartz and feldspar	63
Totals		Pyroxene and olivine	14
Sedimentary	7.9	Hydrated silicates	14.6
Igneous	64.7	Carbonates	2.0
Metamorphic	27.4	Others	6.4
	100.0		100.0

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Texture

The texture of a rock is the size, shape, and arrangement of the grains (for sedimentary rocks) or crystals (for igneous and metamorphic rocks). Also of importance are the rock's extent of homogeneity (*i.e.,* uniformity of composition throughout) and the degree of isotropy. The latter is the extent to which the bulk structure and composition are the same in all directions in the rock.

Analysis of texture can yield information about the rock's source material, conditions and environment of deposition (for sedimentary rock) or crystallization and recrystallization (for igneous and metamorphic rock, respectively), and subsequent geologic history and change.

Classification by grain or crystal size

The common textural terms used for rock types with respect to the size of the grains or crystals, are given in the Table. The particle-size categories are derived from the Udden-Wentworth scale developed for sediment. For igneous and metamorphic rocks, the terms are generally used as modifiers—*e.g.,* medium-grained granite. Aphanitic is a descriptive term for small crystals, and phaneritic for larger ones. Very coarse crystals (those larger than 3 centimetres, or 1.2 inches) are termed pegmatitic.
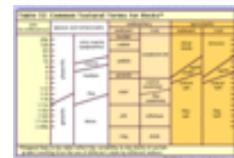
For sedimentary rocks, the broad categories of sediment size are coarse (greater than 2 millimetres, or 0.08 inch), medium (between 2 and $1/_{16}$ millimetres), and fine (under $1/_{16}$ millimetre). The latter includes silt and clay, which both have a size indistinguishable by the human eye and are also termed dust. Most shales (the lithified version of clay) contain some silt. Pyroclastic rocks are those formed from clastic

(from the Greek word for broken) material ejected from volcanoes. Blocks are fragments broken from solid rock, while bombs are molten when ejected.

Porosity

The term rock refers to the bulk volume of the material, including the grains or crystals as well as the contained void space. The volumetric portion of bulk rock that is not occupied by grains, crystals, or natural cementing material is termed porosity. That is to say, porosity is the ratio of void volume to the bulk volume (grains plus void space). This void space consists of pore space between grains or crystals, in addition to crack space. In sedimentary rocks, the amount of pore space depends on the degree of compaction of the sediment (with compaction generally increasing with depth of burial), on the packing arrangement and shape of grains, on the amount of cementation, and on the degree of sorting. Typical cements are siliceous, calcareous or carbonate, or iron-bearing minerals.

Sorting is the tendency of sedimentary rocks to have grains that are similarly sized—*i.e.,* to have a narrow range of sizes (see Figure 2). Poorly sorted sediment displays a wide range of grain sizes and hence has decreased porosity. Well-sorted indicates a grain size distribution that is fairly uniform. Depending on the type of close-packing of the grains, porosity can be substantial. It should be noted that in engineering usage—*e.g.,* geotechnical or civil engineering—the terminology is phrased oppositely and is referred to as grading. A well-graded sediment is a (geologically) poorly sorted one, and a poorly graded sediment is a well-sorted one.
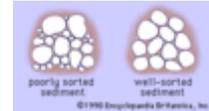
Total porosity encompasses all the void space, including those pores that are interconnected to the surface of the sample as well as those that are sealed off by natural cement or other obstructions. Thus the total porosity (ϕ_T) is

$$\varphi_T = 100 \left(1 - \frac{\mathrm{Vol}_G}{\mathrm{Vol}_B} \right) \%,$$

where Vol_G is the volume of grains (and cement, if any) and Vol_B is the total bulk volume. Alternatively, one can calculate ϕ_T from the measured densities of the bulk rock and of the (mono)mineralic constituent. Thus,

$$\varphi_T = 100 \left(1 - \frac{\rho_B}{\rho_G} \right) \%,$$

where ρ_B is the density of the bulk rock and ρ_G is the density of the grains (*i.e.,* the mineral, if the composition is monomineralogic and homogeneous). For example, if a sandstone has a ρ_B of 2.38 grams per cubic centimetre (g/cm^3) and is composed of quartz (SiO_2) grains having ρ_G of 2.65 g/cm^3, the total porosity is

$$\varphi_T = 100 \left(1 - \frac{2.38}{2.65} \right) = 10.2\%.$$

Apparent (effective, or net) porosity is the proportion of void space that excludes the sealed-off pores. It thus measures the pore volume that is effectively interconnected and accessible to the surface of the sample, which is important when considering the storage and movement of subsurface fluids such as petroleum, groundwater, or contaminated fluids.

Physical properties

Physical properties of rocks are of interest and utility in many fields of work, including geology, petrophysics, geophysics, materials science, geochemistry, and geotechnical engineering. The scale of investigation ranges from the molecular and crystalline up to terrestrial studies of the Earth and other planetary bodies. Geologists are interested in the radioactive age dating of rocks to reconstruct the origin of mineral deposits; seismologists formulate prospective earthquake predictions using premonitory physical or chemical changes; crystallographers study the synthesis of minerals with special optical or physical properties; exploration geophysicists investigate the variation of physical properties of subsurface rocks to make possible detection of natural resources such as oil and gas, geothermal energy, and ores of metals; geotechnical engineers examine the nature and behaviour of the materials on, in, or of which such structures as buildings, dams, tunnels, bridges, and underground storage vaults are to be constructed; solid-state physicists study the magnetic, electrical, and mechanical properties of materials for electronic devices, computer components, or high-performance ceramics; and petroleum reservoir engineers analyze the response measured on well logs or in the processes of deep drilling at elevated temperature and pressure.

Since rocks are aggregates of mineral grains or crystals, their properties are determined in large part by the properties of their various constituent minerals. In a rock these general properties are determined by averaging the relative properties and sometimes orientations of the various grains or crystals. As a result, some properties that are anisotropic (*i.e.,* differ with direction) on a submicroscopic or crystalline scale are fairly isotropic for a large bulk volume of the rock. Many properties are also dependent on grain or crystal size, shape, and packing arrangement, the amount and distribution of void space, the presence of natural cements in sedimentary rocks, the temperature and pressure, and the type and amount of contained fluids (*e.g.,* water, petroleum, gases). Because many rocks exhibit a considerable range in these factors, the assignment of representative values for a particular property is often done using a statistical variation.

Some properties can vary considerably, depending on whether measured in situ (in place in the subsurface) or in the laboratory under simulated conditions. Electrical resistivity, for example, is highly dependent on the fluid content of the rock in situ and the temperature condition at the particular depth.

Physical properties » Density

Density varies significantly among different rock types because of differences in mineralogy and porosity. Knowledge of the distribution of underground rock densities can assist in interpreting subsurface geologic structure and rock type.

In strict usage, density is defined as the mass of a substance per unit volume; however, in common usage, it is taken to be the weight in air of a unit volume of a sample at a specific temperature. Weight is the force that gravitation exerts on a body (and thus varies with location), whereas mass (a measure of the matter in a body) is a fundamental property and is constant regardless of location. In routine density measurements of rocks, the sample weights are considered to be equivalent to their masses, because the discrepancy between weight and mass would result in less error on the computed density than the experimental errors introduced in the measurement of volume. Thus, density is often determined using weight rather than mass. Density should properly be reported in kilograms per cubic metre (kg/m^3), but is still often given in grams per cubic centimetre (g/cm^3).

Another property closely related to density is specific gravity. It is defined, as noted above, as the ratio of the weight or mass in air of a unit volume of material at a stated temperature to the weight or mass in air of a unit volume of distilled water at the same temperature. Specific gravity is dimensionless (*i.e.,* has no units).

The bulk density of a rock is $\rho_B = W_G/V_B$, where W_G is the weight of grains (sedimentary rocks) or crystals (igneous and metamorphic rocks) and natural cements, if any, and V_B is the total volume of the grains or

crystals plus the void (pore) space. The density can be dry if the pore space is empty, or it can be saturated if the pores are filled with fluid (*e.g.,* water), which is more typical of the subsurface (in situ) situation. If there is pore fluid present,

$$\text{(saturated)} \quad \rho_B = \frac{W_G + W_{fl}}{V_B} \, ,$$

where W_{fl} is the weight of pore fluid. In terms of total porosity, saturated density is

$$\text{(saturated)} \quad \rho_B = \rho_G(1 - \varphi_T) + \rho_{fl}\varphi_T \, ,$$

$$\varphi_T = \frac{\rho_G - \rho_B}{\rho_G - \rho_{fl}} \, ,$$

and thus
where ρ_{fl} is the density of the pore fluid. Density measurements for a given specimen involve the determination of any two of the following quantities: pore volume, bulk volume, or grain volume, along with the weight.

A useful way to assess the density of rocks is to make a histogram plot of the statistical range of a set of data. The representative value and its variation can be expressed as follows: (1) mean, the average value, (2) mode, the most common value (*i.e.,* the peak of the distribution curve), (3) median, the value of the middle sample of the data set (*i.e.,* the value at which half of the samples are below and half are above), and (4) standard deviation, a statistical measure of the spread of the data (plus and minus one standard deviation from the mean value includes about two-thirds of the data).

A compilation of dry bulk densities for various rock types found in the upper crust of the Earth is listed in the Table. A histogram plot of these data, giving the percent of the samples as a function of density is shown in Figure 3. The parameters given include (1) sample division, the range of density in one data column—*e.g.,* 0.036 g/cm^3 for , (2) number of samples, and (3) standard deviation. The small inset plot is the percentage of samples (on the vertical axis) that lie within the interval of the "mode - x" to the "mode + x," where x is the horizontal axis.
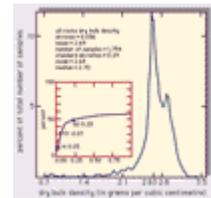
Dry bulk densities for various rock types

rock type	number of samples	mean (grams per cubic cm)	standard deviation	mode (grams per cubic cm)	median (grams per cubic cm)
all rocks	1,647	2.73	0.26	2.65	2.86
andesite	197	2.65	0.13	2.58	2.66
basalt	323	2.74	0.47	2.88	2.87
diorite	68	2.86	0.12	2.89	2.87
dolerite (diabase)	224	2.89	0.13	2.96	2.90
gabbro	98	2.95	0.14	2.99	2.97
granite	334	2.66	0.06	2.66	2.66
quartz porphyry	76	2.62	0.06	2.60	2.62
rhyolite	94	2.51	0.13	2.60	2.49
syenite	93	2.70	0.10	2.67	2.68

| trachyte | 71 | 2.57 | 0.10 | 2.62 | 2.57 |
| sandstone | 107 | 2.22 | 0.23 | 2.22 | 2.22 |

Source: After data from H.S. Washington (1917) and R.J. Piersol, L.E. Workman, and M.C. Watson (1940) as compiled by Gary R. Olhoeft and Gordon R. Johnson in Robert S. Carmichael, ed., *Handbook of Physical Properties of Rocks,* vol. III, CRC Press, Inc. (1984).

In , the most common (modal) value of the distribution falls at 2.63 g/cm^3, roughly the density of quartz, an abundant rock-forming mineral. Few density values for these upper crustal rocks lie above 3.3 g/cm^3. A few fall well below the mode, even occasionally under 1 g/cm^3. The reason for this is shown in Figure 4, which illustrates the density distributions for granite, basalt, and sandstone. Granite is an intrusive igneous rock with low porosity and a well-defined chemical (mineral) composition; its range of densities is narrow. Basalt is, in most cases, an extrusive igneous rock that can exhibit a large variation in porosity (because entrained gases leave voids called vesicles), and thus some highly porous samples can have low densities. Sandstone is a clastic sedimentary rock that can have a wide range of porosities depending on the degree of sorting, compaction, packing arrangement of grains, and cementation. The bulk density varies accordingly. Other distribution plots of dry bulk densities are given in Figures 5 and Figure 6B, with a sample division
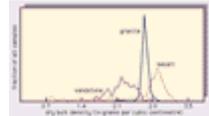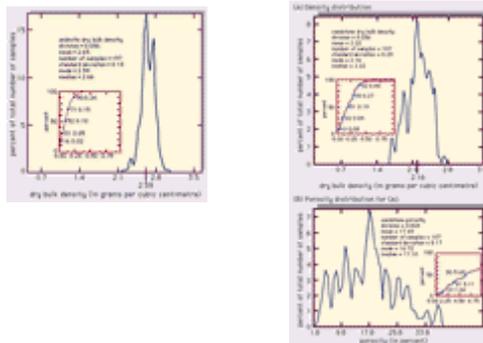
of 0.036 g/cm^3 for and and of 0.828 percent for . The Table lists typical ranges of dry bulk densities for a variety of other rock types as prepared by the American geologists Gordon R. Johnson and Gary R. Olhoeft.

Typical density ranges for some other rock types

rock type	density (grams per cubic cm)
amphibolite	2.79–3.14
andesite glass	2.40–2.57
anhydrite	2.82–2.93
anorthosite	2.64–2.92
basalt glass	2.70–2.85
chalk	2.23
dolomite	2.72–2.84
dunite	2.98–3.76
eclogite	3.32–3.45
gneiss	2.59–2.84
granodiorite	2.67–2.78
limestone	1.55–2.75

marble	2.67–2.75
norite	2.72–3.02
peridotite	3.15–3.28
quartzite	2.65
rock salt	2.10–2.20
schist	2.73–3.19
shale	2.06–2.67
slate	2.72–2.84

Source: After data from R.A. Daly, G.E. Manger, and S.P. Clark, Jr. (1966); A.F. Birch (1966); F. Press (1966); and R.N. Schock, B.P. Bonner, and H. Louis (1974) in Robert S. Carmichael, ed., *Handbook of Physical Properties of Rocks,* vol. III, CRC Press, Inc. (1984).

The density of clastic sedimentary rocks increases as the rocks are progressively buried. This is because of the increase of overburden pressure, which causes compaction, and the progressive cementation with age. Both compaction and cementation decrease the porosity.

Representative densities for common rock-forming minerals (*i.e.,* ρ_G) and rocks (*i.e.,* ρ_B) are listed in the Table. The bulk densities for sedimentary rocks, which typically have variable porosity, are given as ranges of both dry ρ_B and (water-) saturated ρ_B. The pore-filling fluid is usually briny water, often indicative of the presence of seawater when the rock was being deposited or lithified. It should be noted that the bulk density is less than the grain density of the constituent mineral (or mineral assemblage), depending on the porosity. For example, sandstone (characteristically quartzose) has a typical dry bulk density of 2.0–2.6 g/cm^3, with a porosity that can vary from low to more than 30 percent. The density of quartz itself is 2.65 g/cm^3. If porosity were zero, the bulk density would equal the grain density.

Table 35: Typical Densities

material		ρ (g/cm^3)	
minerals (approximately pure, no porosity)		ρ_G	
Halite (NaCl)		2.16	
Gypsum (CaSO$_4\cdot$2H$_2$O)		2.32	
Orthoclase feldspar (KAlSi$_3$O$_8$)		2.56	
Quartz (SiO$_2$)		2.65	
Calcite (CaCO$_3$)		2.71	
Anhydrite (CaSO$_4$)		2.9	
Olivine (Mg,Fe)$_2$SiO$_4$		3.3–3.6	
Hematite (Fe$_2$O$_3$)		5.26	

rocks (surface T and P)	typical porosity	dry ρ_B	saturated ρ_B
Loess		1.1	1.4–1.6
Silt		1.36	1.4–1.8
Rock salt		2.2	—
Shale	≧8%	1.9–2.4	2.1–2.5
Sandstone	~15%, variable	2.0–2.6	2.2–2.6
Limestone	~5%	1.9–2.6	2.1–2.7
Dolomite		2.5–2.8	2.6–2.8
Granite-diorite		2.6–2.9	
Gabbro-diabase		2.8–3.1	
Dunite-peridotite		3.1–3.4	
Granite; sial crust		2.67	
Basalt; sima crust		2.87	
Top of mantle		3.32	

other		ρ	
Water (0°C)		0.99987	
20°C		0.99823	
100°C		0.95841	
Seawater (0°C)		1.025	
Ice		0.92	
Oil		~0.8–0.9	
Methane gas (at 200 bars, $T\sim$80°C)		~0.12	

Saturated bulk density is higher than dry bulk density, owing to the added presence of pore-filling fluid.

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Calcite ($CaCO_3$)		2.71	
Anhydrite ($CaSO_4$)		2.9	
Olivine ($Mg,Fe)_2SiO_4$		3.3–3.6	
Hematite (Fe_2O_3)		5.26	
rocks (surface T and P)	typical porosity	dry ρ_B	saturated ρ_B
Loess		1.1	1.4–1.6
Silt		1.36	1.4–1.8
Rock salt		2.2	—
Shale	≥8%	1.9–2.4	2.1–2.5
Sandstone	~15%, variable	2.0–2.6	2.2–2.6
Limestone	~5%	1.9–2.6	2.1–2.7
Dolomite		2.5–2.8	2.6–2.8
Granite-diorite		2.6–2.9	
Gabbro-diabase		2.8–3.1	
Dunite-peridotite		3.1–3.4	
Granite; sial crust		2.67	
Basalt; sima crust		2.87	
Top of mantle		3.32	
other		ρ	
Water (0 °C)		0.99987	
20 °C		0.99823	
100 °C		0.95841	
Seawater (0° C)		1.025	
Ice		0.92	
Oil		~0.8–0.9	
Methane gas (at 200 bars, $T\sim$80° C)		~0.12	

Table also lists representative values for density of seawater, oil, and methane gas at a subsurface condition—pressure of 200 bars (one bar = 0.987 atmosphere, or 29.53 inches of mercury) and a temperature of about 80° C (176° F).

Mechanical properties » Stress and strain

When a stress σ (force per unit area) is applied to a material such as rock, the material experiences a change in dimension, volume, or shape. This change, or deformation, is called strain (ε). Stresses can be axial—*e.g.,* directional tension or simple compression—or shear (tangential), or all-sided (*e.g.,* hydrostatic compression). The terms stress and pressure are sometimes used interchangeably, but often stress refers to directional stress or shear stress and pressure (*P*) refers to hydrostatic compression. For

small stresses, the strain is elastic (recoverable when the stress is removed and linearly proportional to the applied stress). For larger stresses and other conditions, the strain can be inelastic, or permanent.

Physical properties » Mechanical properties » Elastic constants
In elastic deformation, there are various constants that relate the magnitude of the strain response to the applied stress. These elastic constants include the following:
(1) Young's modulus (E) is the ratio of the applied stress to the fractional extension (or shortening) of the sample length parallel to the tension (or compression). The strain is the linear change in dimension divided by the original length.
(2) Shear modulus (μ) is the ratio of the applied stress to the distortion (rotation) of a plane originally perpendicular to the applied shear stress; it is also termed the modulus of rigidity.
(3) Bulk modulus (k) is the ratio of the confining pressure to the fractional reduction of volume in response to the applied hydrostatic pressure. The volume strain is the change in volume of the sample divided by the original volume. Bulk modulus is also termed the modulus of incompressibility.
(4) Poisson's ratio (σ_p) is the ratio of lateral strain (perpendicular to an applied stress) to the longitudinal strain (parallel to applied stress).
For elastic and isotropic materials, the elastic constants are interrelated. For example,

$$\sigma_p = \frac{E}{2\mu} - 1;$$

$$E = 3k(1 - 2\sigma_p);$$

and

$$\mu = \frac{3}{2} k \left(\frac{1 - 2\sigma_p}{1 + \sigma_p} \right).$$

The following are the common units of stress:

$$1 \text{ bar} = 10^6 \text{ dynes per square centimetre}$$
$$= 10^5 \text{ newtons per square metre, or pascals (Pa)}$$
$$= 0.1 \text{ megapascal } (i.e., 0.1 \times 10^6 \text{ Pa}).$$

Thus 10 kilobars = 1 gigapascal (*i.e.,* 10^9 Pa).
Physical properties » Mechanical properties » Rock mechanics
The study of deformation resulting from the strain of rocks in response to stresses is called rock mechanics. When the scale of the deformation is extended to large geologic structures in the crust of the Earth, the field of study is known as geotectonics.
The mechanisms and character of the deformation of rocks and Earth materials can be investigated through laboratory experiments, development of theoretical models based on the properties of materials, and study of deformed rocks and structures in the field. In the laboratory, one can simulate—either directly or by appropriate scaling of experimental parameters—several conditions. Two types of pressure may be simulated: confining (hydrostatic), due to burial under rock overburden, and internal (pore), due to pressure exerted by pore fluids contained in void space in the rock. Directed applied stress, such as compression, tension, and shear, is studied, as are the effects of increased temperature introduced with depth in the Earth's crust. The effects of the duration of time and the rate of applying stress (*i.e.,* loading) as a function of time are examined. Also, the role of fluids, particularly if they are chemically active, is investigated.
Some simple apparatuses for deforming rocks are designed for biaxial stress application: a directed (uniaxial) compression is applied while a confining pressure is exerted (by pressurized fluid) around the

cylindrical specimen. This simulates deformation at depth within the Earth. An independent internal pore-fluid pressure also can be exerted. The rock specimen can be jacketed with a thin, impermeable sleeve (*e.g.,* rubber or copper) to separate the external pressure medium from the internal pore fluids (if any). The specimen is typically a few centimetres in dimension.

Another apparatus for exerting high pressure on a sample was designed in 1968 by Akira Sawaoka, Naoto Kawai, and Robert Carmichael to give hydrostatic confining pressures up to 12 kilobars (1.2 gigapascal), additional directed stress, and temperatures up to a few hundred degrees Celsius. The specimen is positioned on the baseplate; the pressure is applied by driving in pistons with a hydraulic press. The end caps can be locked down to hold the pressure for time experiments and to make the device portable.

Apparatuses have been developed, typically using multianvil designs, which extend the range of static experimental conditions—at least for small specimens and limited times—to pressures as high as about 1,700 kilobars and temperatures of about 2,000° C. Such work has been pioneered by researchers such as Peter M. Bell and Ho-Kwang Mao, who conducted studies at the Geophysical Laboratory of the Carnegie Institution in Washington, D.C. Using dynamic techniques (*i.e.,* shock from explosive impact generated by gun-type designs), even higher pressures up to 7,000 kilobars (700 gigapascal)—which is nearly twice the pressure at the centre of the Earth and seven million times greater than the atmospheric pressure at the Earth's surface—can be produced for very short times. A leading figure in such ultrapressure work is A. Sawaoka at the Tokyo Institute of Technology.

In the upper crust of the Earth, hydrostatic pressure increases at the rate of about 320 bars per kilometre, and temperature increases at a typical rate of 20°–40° C per kilometre, depending on recent crustal geologic history. Additional directed stress, as can be generated by large-scale crustal deformation (tectonism), can range up to 1 to 2 kilobars. This is approximately equal to the ultimate strength (before fracture) of solid crystalline rock at surface temperature and pressure (see below). The stress released in a single major earthquake—a shift on a fault plane—is about 50–150 bars.

In studying the deformation of rocks one can start with the assumption of ideal behaviour: elastic strain and homogeneous and isotropic stress and strain. In reality, on a microscopic scale there are grains and pores in sediments and a fabric of crystals in igneous and metamorphic rocks. On a large scale, rock bodies exhibit physical and chemical variations and structural features. Furthermore, conditions such as extended length of time, confining pressure, and subsurface fluids affect the rates of change of deformation.

Figure 7 shows the generalized transition from brittle fracture through faulting to plastic-flow deformation in response to applied compressional stress and the progressive increase of confining pressure.
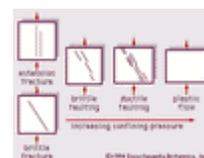
Mechanical properties » Stress-strain relationships

The deformation of materials is characterized by stress-strain relations. For elastic-behaviour materials, the strain is proportional to the load (*i.e.,* the applied stress). The strain is immediate with stress and is reversible (recoverable) up to the yield point stress, beyond which permanent strain results. For viscous material, there is laminar (slow, smooth, parallel) flow; one must exert a force to maintain motion because of internal frictional resistance to flow, called the viscosity. Viscosity varies with the applied stress, strain rate, and temperature. In plastic behaviour, the material strains continuously (but still has strength) after the yield point stress is reached; however, beyond this point there is some permanent deformation. In elasticoviscous deformation, there is combined elastic and viscous behaviour. The material yields continuously (viscously) for a constant applied load. An example of such behaviour is creep, a slow, permanent, and continuous deformation occurring under constant load over a long time in such materials as crystals, ice, soil and sediment, and rocks at depth. In firmoviscous behaviour, the material is essentially solid but the strain is not immediate with application of stress; rather, it is taken

up and released exponentially. A plasticoviscous material exhibits elastic behaviour for initial stress (as in plastic behaviour), but after the yield point stress is reached, it flows like a viscous fluid.

Some representative values of elastic constants and properties are listed in Table 36. The coefficient of viscosity (η) is the ratio of applied stress to the rate of straining (change of strain with time). It is measured in units of poise; one poise equals one dyne-second per square centimetre.

Some typical values of elastic constants and properties

|	elastic constants (at room temperature and pressure)	
material	Young's modulus (in 10^6 bars)	shear modulus (in 10^6 bars)
ice	0.1	0.03
shale	0.2–0.3	0.15
limestone	0.4–0.7	0.22–0.26
granite	0.3–0.6	0.2
basalt	0.7–0.9	0.3
steel	2.1	0.83
material	temperature (degrees Celsius)	coefficient of viscosity (poises)
lava (Mount Vesuvius)	1,100 1,400	28,300 250
lava (Oshima, Japan)	1,038 1,125	230,000 5,600
andesite lava	1,400	150–1,500
material	compressive strength (at room temperature and pressure, in kilobars)	
shale	0.8–1.8	
sandstone	0.5–2	
limestone	1–2	
granite	1.7–2.5	
basalt	1–3.4	

Rheology is the study of the flow deformation of materials. The concept of rheidity refers to the capacity of a material to flow, arbitrarily defined as the time required with a shear stress applied for the viscous strain to be 1,000 times greater than the elastic strain. It is thus a measure of the threshold of fluidlike behaviour. Although such behaviour depends on temperature, relative comparisons can be made. Some representative values of rheidity times are given in the Table.

Rheidity threshold of fluidlike deformation

material	approximate time
ice (e.g., glacier)	2 weeks
gypsum	1 year
rock salt (e.g., saltdome)	10–20 years
serpentine (a mafic silicate mineral)	10,000 years

Typical stress-strain (deformation) curves for rock materials are shown in

Figure 8. The stress σ, compression in the figure, is force per unit area. The strain ε is fractional shortening of the specimen parallel to the applied compression; it is given here in percent. The brittle material behaves elastically nearly until the point of fracture (denoted *X*), whereas the ductile (plastically deformable) material is elastic up to the yield point but then has a range of plastic deformation before fracturing. The ability to undergo large permanent deformation before fracture is called ductility. For plastic deformation, the flow mechanisms are intracrystalline (slip and twinning within crystal grains), intercrystalline motion by crushing and fracture (cataclasis), and recrystallization by solutioning or solid diffusion.
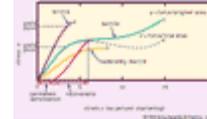
If the applied stress is removed while a ductile material is in the plastic range, part of the strain is recoverable (elastically), but there is permanent deformation. The ultimate strength is the highest point (stress) on a stress-strain curve, often occurring at fracture (which is the complete loss of cohesion). The strength of a material is its resistance to failure (destruction of structure) by flow or fracture; it is a measure of the stress required to deform a body. Typical compressive strengths (the stress required to cause failure under compression) are given in the Table.

Physical properties » Mechanical properties » Effect of environmental conditions

The behaviour and mechanical properties of rocks depend on a number of environmental conditions. (1) Confining pressure increases the elasticity, strength (*e.g.,* yield point and ultimate fracture stress), and ductility. (2) Internal pore-fluid pressure reduces the effective stress acting on the sample, thus reducing the strength and ductility. The effective, or net, confining pressure is the external hydrostatic pressure minus the internal pore-fluid pressure. (3) Temperature lowers the strength, enhances ductility, and may enhance recrystallization. (4) Fluid solutions can enhance deformation, creep, and recrystallization. (5) Time is an influential factor as well. (6) The rate of loading (*i.e.,* the rate at which stress is applied) influences mechanical properties. (7) Compaction, as would occur with burial to depth, reduces the volume of pore space for sedimentary rocks and the crack porosity for crystalline rocks.

Rocks, which are typically brittle at the Earth's surface, can undergo ductile deformation when buried and subjected to increased confining pressure and temperature for long periods of time. If stress exceeds their strength or if they are not sufficiently ductile, they will fail by fracture—as a crystal, within a bed or rock, on an earthquake fault zone, and so on—whereas with ductility they can flow and fold. Some strengths for various rock types under different temperatures and confining pressures are listed in the Table. The plastic yield strength here is the stress at a 2 percent strain; the ultimate strength, as stated above, is the highest point (stress) on the stress-strain curve.

Rock strengths, with varying temperature and pressure

rock type	temperature (°C)	confining pressure (kilobars)	plastic yield strength (kilobars)	ultimate strength (kilobars)
granite	500	5	10	11.5
	800	5	5	6
gabbro	500	5	4	8
peridotite	500	5	8	9
	800	5	5.5	8
basalt	500	5	8	10
	800	5	2	2.5
marble	24	2	2.5	5.5
	500	3	1	2
limestone	24	2	4.5	5.5

	500	3	2.5	3
dolomite	24	2	6	7
	500	5	4	6.5
shale	24	2	1.5	2.5
rock salt	24	1	0.5	1

An increase in confining pressure causes brittle fracture to become shear slippage and eventually causes flow (ductile) behaviour. This transition is also aided by higher temperature, decreased internal pore-fluid pressure, and slower strain rate.

The Table gives the values of some elastic constants—bulk modulus (k), Young's modulus (E), shear modulus (μ), and Poisson's ratio (σ_p)—at room pressure (1 bar) and high confining pressure (3,000 bars). The values for clastic sedimentary rocks would be particularly variable.

Variation of some elastic constants (in 10^6 bars) with rock type and confining pressure

at pressure = 1 bar

rock type	bulk modulus	Young's modulus	shear modulus	Poisson's ratio
granite	0.1	0.3	0.2	0.05
gabbro	0.3	0.9	0.6	0.1
dunite	1.1	1.5	0.5	0.3
obsidian	0.4	0.7	0.3	0.08
basalt	0.5	0.8	0.3	0.23
gneiss	0.1	0.2	0.1	0.05
marble	0.1	0.4	0.2	0.1
quartzite				
sandstone	0.07	0.2	0.08	0.1
shale	0.04	0.1	0.05	0.04
limestone	0.8	0.6	0.2	0.3

at pressure = 3,000 bars

rock type	bulk modulus	Young's modulus	shear modulus	Poisson's ratio
granite	0.5	0.6	0.4	0.25
gabbro	0.9	0.8	0.5	0.2
dunite	1.2	1.7	0.7	0.27
obsidian				
basalt	0.8	1.2	0.4	0.25
gneiss	0.5	0.7	0.3	
marble	0.8	0.7	0.3	0.3
quartzite	0.5	1.0	0.4	0.07
sandstone				
shale				
limestone				

Physical properties » Electrical properties

The electrical nature of a material is characterized by its conductivity (or, inversely, its resistivity) and its dielectric constant, and coefficients that indicate the rates of change of these with temperature, frequency at which measurement is made, and so on. For rocks with a range of chemical composition as well as variable physical properties of porosity and fluid content, the values of electrical properties can vary widely.

Resistance (R) is defined as being one ohm when a potential difference (voltage; V) across a specimen of one volt magnitude produces a current (i) of one ampere; that is, $V = Ri$. The electrical resistivity (ρ) is an intrinsic property of the material. In other words, it is inherent and not dependent on sample size or current path. It is related to resistance by $R = \rho L/A$ where L is the length of specimen, A is the cross-sectional area of specimen, and units of ρ are ohm-centimetre; 1 ohm-centimetre equals 0.01 ohm-metre. The conductivity (σ) is equal to $1/\rho$ ohm$^{-1}\cdot$ centimetre^{-1} (or termed mhos/cm). In SI units, it is given in mhos/metre, or siemens/metre.

Some representative values of electrical resistivity for rocks and other materials are listed in the Table. Materials that are generally considered as "good" conductors have a resistivity of 10^{-5}–10 ohm-centimetre (10^{-7}–10^{-1} ohm-metre) and a conductivity of 10–10^{7} mhos/metre. Those that are classified as intermediate conductors have a resistivity of 100–10^{9} ohm-centimetre (1–10^{7} ohm-metre) and a conductivity of 10^{-7}–1 mhos/metre. "Poor" conductors, also known as insulators, have a resistivity of 10^{10}–10^{17} ohm-centimetre (10^{8}–10^{15} ohm-metre) and a conductivity of 10^{-15}–10^{-8}. Seawater is a much better conductor (*i.e.,* it has lower resistivity) than fresh water owing to its higher content of dissolved salts; dry rock is very resistive. In the subsurface, pores are typically filled to some degree by fluids. The resistivity of materials has a wide range—copper is, for example, different from quartz by 22 orders of magnitude.

Typical resistivities

material	resistivity (ohm-centimetre)
seawater (18 °C)	21
uncontaminated surface water	$2(10^{4})$
distilled water	0.2–$1(10^{6})$
water (4 °C)	$9(10^{6})$
ice	$3(10^{8})$
Rocks (in situ)	
sedimentary	
clay, soft shale	100–$5(10^{3})$
hard shale	7–$50(10^{3})$
sand	5–$40(10^{3})$
sandstone	(10^{4})–(10^{5})
glacial moraine	1–$500(10^{3})$
porous limestone	1–$30(10^{4})$
dense limestone	$>(10^{6})$
rock salt	(10^{8})–(10^{9})
igneous	$5(10^{4})$–(10^{8})
metamorphic	$5(10^{4})$–$5(10^{9})$
Rocks (laboratory)	

dry granite	10^{12}
Minerals	
copper (18 °C)	$1.7(10^{-6})$
graphite	$5–500(10^{-4})$
pyrrhotite	0.1–0.6
magnetite crystals	0.6–0.8
pyrite ore	$1–(10^5)$
magnetite ore	$(10^2)–5(10^5)$
chromite ore	$>10^6$
quartz (18 °C)	$(10^{14})–(10^{16})$

For high-frequency alternating currents, the electrical response of a rock is governed in part by the dielectric constant, ε. This is the capacity of the rock to store electric charge; it is a measure of polarizability in an electric field. In cgs units, the dielectric constant is 1.0 in a vacuum. In SI units, it is given in farads per metre or in terms of the ratio of specific capacity of the material to specific capacity of vacuum (which is 8.85×10^{-12} farads per metre). The dielectric constant is a function of temperature, and of frequency, for those frequencies well above 100 hertz (cycles per second).

Electrical conduction occurs in rocks by (1) fluid conduction—*i.e.,* electrolytic conduction by ionic transfer in briny pore water—and (2) metallic and semiconductor (*e.g.,* some sulfide ores) electron conduction. If the rock has any porosity and contained fluid, the fluid typically dominates the conductivity response. The rock conductivity depends on the conductivity of the fluid (and its chemical composition), degree of fluid saturation, porosity and permeability, and temperature. If rocks lose water, as with compaction of clastic sedimentary rocks at depth, their resistivity typically increases.

Physical properties » Magnetic properties

The magnetic properties of rocks arise from the magnetic properties of the constituent mineral grains and crystals. Typically, only a small fraction of the rock consists of magnetic minerals. It is this small portion of grains that determines the magnetic properties and magnetization of the rock as a whole, with two results: (1) the magnetic properties of a given rock may vary widely within a given rock body or structure, depending on chemical inhomogeneities, depositional or crystallization conditions, and what happens to the rock after formation; and (2) rocks that share the same lithology (type and name) need not necessarily share the same magnetic characteristics. Lithologic classifications are usually based on the abundance of dominant silicate minerals, but the magnetization is determined by the minor fraction of such magnetic mineral grains as iron oxides. The major rock-forming magnetic minerals are iron oxides and sulfides.

Although the magnetic properties of rocks sharing the same classification may vary from rock to rock, general magnetic properties do nonetheless usually depend on rock type and overall composition. The magnetic properties of a particular rock can be quite well understood provided one has specific information about the magnetic properties of crystalline materials and minerals, as well as about how those properties are affected by such factors as temperature, pressure, chemical composition, and the size of the grains. Understanding is further enhanced by information about how the properties of typical rocks are dependent on the geologic environment and how they vary with different conditions.

Physical properties » Magnetic properties » Applications of the study of rock magnetization

An understanding of rock magnetization is important in at least three different areas: prospecting, geology, and materials science. In magnetic prospecting, one is interested in mapping the depth, size, type, and inferred composition of buried rocks. The prospecting, which may be done from ground

surface, ship, or aircraft, provides an important first step in exploring buried geologic structures and may, for example, help identify favourable locations for oil, natural gas, and economic mineral deposits. Rock magnetization has traditionally played an important role in geology. Paleomagnetic work seeks to determine the remanent magnetization (see below Types of remanent magnetization) and thereby ascertain the character of the Earth's field when certain rocks were formed. The results of such research have important ramifications in stratigraphic correlation, age dating, and reconstructing past movements of the Earth's crust. Indeed, magnetic surveys of the oceanic crust provided for the first time the quantitative evidence needed to cogently demonstrate that segments of the crust had undergone large-scale lateral displacements over geologic time, thereby corroborating the concepts of continental drift and seafloor spreading, both of which are fundamental to the theory of plate tectonics (see plate tectonics).

The understanding of magnetization is increasingly important in materials science as well. The design and manufacture of efficient memory cores, magnetic tapes, and permanent magnets increasingly rely on the ability to create materials having desired magnetic properties.

Physical properties » Magnetic properties » Basic types of magnetization

There are six basic types of magnetization: (1) diamagnetism, (2) paramagnetism, (3) ferromagnetism, (4) antiferromagnetism, (5) ferrimagnetism, and (6) superparamagnetism.

Diamagnetism arises from the orbiting electrons surrounding each atomic nucleus. When an external magnetic field is applied, the orbits are shifted in such a way that the atoms set up their own magnetic field in opposition to the applied field. In other words, the induced diamagnetic field opposes the external field. Diamagnetism is present in all materials, is weak, and exists only in the presence of an applied field. The propensity of a substance for being magnetized in an external field is called its susceptibility (k) and it is defined as J/H, where J is the magnetization (intensity) per unit volume and H is the strength of the applied field. Since the induced field always opposes the applied field, the sign of diamagnetic susceptibility is negative. The susceptibility of a diamagnetic substance is on the order of -10^{-6} electromagnetic units per cubic centimetre (emu/cm^3). It is sometimes denoted κ for susceptibility per unit mass of material.

Paramagnetism results from the electron spin of unpaired electrons. An electron has a magnetic dipole moment—which is to say that it behaves like a tiny bar magnet—and so when a group of electrons is placed in a magnetic field, the dipole moments tend to line up with the field. The effect augments the net magnetization in the direction of the applied field. Like diamagnetism, paramagnetism is weak and exists only in the presence of an applied field, but since the effect enhances the applied field, the sign of the paramagnetic susceptibility is always positive. The susceptibility of a paramagnetic substance is on the order of 10^{-4} to 10^{-6} emu/cm^3.

Ferromagnetism also exists because of the magnetic properties of the electron. Unlike paramagnetism, however, ferromagnetism can occur even if no external field is applied. The magnetic dipole moments of the atoms spontaneously line up with one another because it is energetically favourable for them to do so. A remanent magnetization can be retained. Complete alignment of the dipole moments would take place only at a temperature of absolute zero (0 kelvin [K], or -273.15° C). Above absolute zero, thermal motions begin to disorder the magnetic moments. At a temperature called the Curie temperature, which varies from material to material, the thermally induced disorder overcomes the alignment, and the ferromagnetic properties of the substance disappear. The susceptibility of ferromagnetic materials is large and positive. It is on the order of 10 to 10^4 emu/cm^3. Only a few materials—iron, cobalt, and nickel—are ferromagnetic in the strict sense of the word and have a strong residual magnetization. In general usage, particularly in engineering, the term ferromagnetic is frequently applied to any material that is appreciably magnetic.

Antiferromagnetism occurs when the dipole moments of the atoms in a material assume an antiparallel arrangement in the absence of an applied field. The result is that the sample has no net magnetization.

The strength of the susceptibility is comparable to that of paramagnetic materials. Above a temperature called the Néel temperature, thermal motions destroy the antiparallel arrangement, and the material then becomes paramagnetic. Spin-canted (anti)ferromagnetism is a special condition which occurs when antiparallel magnetic moments are deflected from the antiferromagnetic plane, resulting in a weak net magnetism. Hematite (α-Fe_2O_3) is such a material.

Ferrimagnetism is an antiparallel alignment of atomic dipole moments which does yield an appreciable net magnetization resulting from unequal moments of the magnetic sublattices. Remanent magnetization is detectable (see below). Above the Curie temperature the substance becomes paramagnetic. Magnetite (Fe_3O_4), which is the most magnetic common mineral, is a ferrimagnetic substance.

Superparamagnetism occurs in materials having grains so small (about 100 angstroms) that any cooperative alignment of dipole moments is overcome by thermal energy.

Physical properties » Magnetic properties » Types of remanent magnetization

Rocks and minerals may retain magnetization after the removal of an externally applied field, thereby becoming permanent weak magnets. This property is known as remanent magnetization and is manifested in different forms, depending on the magnetic properties of the rocks and minerals and their geologic origin and history. Delineated below are the kinds of remanent magnetization frequently observed.

CRM (chemical, or crystallization, remanent magnetization) can be induced after a crystal is formed and undergoes one of a number of physicochemical changes, such as oxidation or reduction, a phase change, dehydration, recrystallization, or precipitation of natural cements. The induction, which is particularly important in some (red) sediments and metamorphic rocks, typically takes place at constant temperature in the Earth's magnetic field.

DRM (depositional, or detrital, remanent magnetization) is formed in clastic sediments when fine particles are deposited on the floor of a body of water. Marine sediments, lake sediments, and some clays can acquire DRM. The Earth's magnetic field aligns the grains, yielding a preferred direction of magnetization.

IRM (isothermal remanent magnetization) results from the application of a magnetic field at a constant (isothermal) temperature, often room temperature.

NRM (natural remanent magnetization) is the magnetization detected in a geologic in situ condition. The NRM of a substance may, of course, be a combination of any of the other remanent magnetizations described here.

PRM (pressure remanent, or piezoremanent, magnetization) arises when a material undergoes mechanical deformation while in a magnetic field. The process of deformation may result from hydrostatic pressure, shock impact (as produced by a meteorite striking the Earth's surface), or directed tectonic stress. There are magnetization changes with stress in the elastic range, but the most pronounced effects occur with plastic deformation when the structure of the magnetic minerals is irreversibly changed.

TRM (thermoremanent magnetization) occurs when a substance is cooled, in the presence of a magnetic field, from above its Curie temperature to below that temperature. This form of magnetization is generally the most important, because it is stable and widespread, occurring in igneous and sedimentary rocks. TRM also can occur when dealing exclusively with temperatures below the Curie temperature. In PTRM (partial thermoremanent magnetization) a sample is cooled from a temperature below the Curie point to yet a lower temperature.

VRM (viscous remanent magnetization) results from thermal agitation. It is acquired slowly over time at low temperatures and in the Earth's magnetic field. The effect is weak and unstable but is present in most rocks.

Physical properties » Magnetic properties » Hysteresis and magnetic susceptibility

The concept of hysteresis is fundamental when describing and comparing the magnetic properties of rocks. Hysteresis is the variation of magnetization with applied field and illustrates the ability of a material to retain its magnetization, even after an applied field is removed.

Figure 9 illustrates this phenomenon in the form of a plot of magnetization (J) versus applied field (H_{ex}). J_s is the saturation (or "spontaneous") magnetization when all the magnetic moments are aligned in their configuration of maximum order. It is temperature-dependent, reaching zero at the Curie temperature. $J_{r,sat}$ is the remanent magnetization that remains when a saturating (large) applied field is removed, and J_r is the residual magnetization left by some process apart from IRM saturation, as, for example, TRM. H_c is the coercive field (or force) that is required to reduce $J_{r,sat}$ to zero, and $H_{c,r}$ is the field required to reduce J_r to zero.
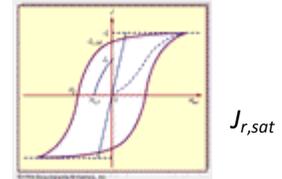
Magnetic susceptibility is a parameter of considerable diagnostic and interpretational use in the study of rocks. This is true whether an investigation is being conducted in the laboratory or magnetic fields over a terrain are being studied to deduce the structure and lithologic character of buried rock bodies. Susceptibility for a rock type can vary widely, depending on magnetic mineralogy, grain size and shape, and the relative magnitude of remanent magnetization present, in addition to the induced magnetization from the Earth's weak field. The latter is given as $J_{induced} = kH_{ex}$, where k is the (true) magnetic susceptibility and H_{ex} is the external (*i.e.,* the Earth's) magnetic field. If there is an additional remanent magnetization with its ratio (Q_n) to induced magnetization being given by

$$Q_n = \frac{J_{remanent}}{kH_{ex}},$$

then the total magnetization is

$$J = J_{induced} + J_{remanent}$$
$$= kH_{ex} + Q_n kH_{ex}$$
$$= k(1 + Q_n)H_{ex}$$
$$= k_{app}\, H_{ex},$$

where k_{app}, the "apparent" magnetic susceptibility, is $k(1 + Q_n)$.

Physical properties » Magnetic properties » Magnetic minerals and magnetic properties of rocks

The major rock-forming magnetic minerals are the following iron oxides: the titanomagnetite series, $x\mathrm{Fe_2TiO_4} \cdot (1 - x)\mathrm{Fe_3O_4}$, where $\mathrm{Fe_3O_4}$ is magnetite, the most magnetic mineral; the ilmenohematite series, $y\mathrm{FeTiO_3} \cdot (1 - y)\mathrm{Fe_2O_3}$, where $\alpha\text{-}\mathrm{Fe_2O_3}$ (in its rhombohedral structure) is hematite; maghemite, $\gamma\text{-}\mathrm{Fe_2O_3}$ (in which some iron atoms are missing in the hematite structure); and limonite (hydrous iron oxides). They also include sulfides—namely, the pyrrhotite series, $y\mathrm{FeS} \cdot (1 - y)\mathrm{Fe_{1-x}S}$.

The Table gives some typical values of the apparent susceptibility for various rock types, which usually include some remanent as well as induced magnetization. Values are higher for mafic igneous rocks, especially as the content of magnetite increases.

Approximate "apparent susceptibilities" for rock types

rock	apparent magnetic susceptibility (electromagnetic units per cubic centimetre)
iron ores	over 0.1
basalt	10^{-2}
andesite	10^{-3}
dacite	10^{-4}

metamorphic rocks 10^{-4}

sedimentary rocks 10^{-5}

Source: From T. Nagata, ed., *Rock Magnetism,* Maruzen Co., Tokyo (1961).

A distribution of measured (true) susceptibilities for various rock types is shown in the Table. Basic refers to those rocks high in iron and magnesium silicates and magnetite, extrusive means formed by cooling after extruding onto the land surface or seafloor. The data in each category are based on at least 45 samples.

Measured susceptibilities for rock types

rock type	% of samples with magnetic susceptibility (in 10^{-6} electromagnetic units per cubic centimetre)			
	less than 100	100–1,000	1,000–4,000	greater than 4,000
Basic extrusive (e.g., basalt)	5	29	47	19
Basic intrusive (e.g., gabbro)	24	27	28	21
Granite	60	23	16	1
Metamorphic (gneiss, schist, slate)	71	22	7	0
Sedimentary	73	19	4	4

Source: After D.H. Lindsley, G.E. Andreasen, and J.R. Balsley, "Magnetic Properties of Rocks and Minerals" in *Handbook of Physical Constants,* S.P. Clark, ed., Memoir 97, Geological Society of America, 1966, and L.B. Slichter, "Magnetic Properties of Rocks" in *Handbook of Physical Constants,* F. Birch, ed. (1942).

The Table lists representative values for the magnetic properties J_n (natural remanent magnetization), k (susceptibility), and ratio Q_n. Natural remanent magnetization is some combination of remanences; typically TRM in an [igneous rock](), perhaps DRM or CRM or both in a sedimentary rock, and all with an additional VRM. The ratio Q_n is typically higher for rocks with a strong, stable remanence—*e.g.,* magnetite-rich and fine-grained extrusive rocks such as seafloor basalts.

Typical magnetic properties of rocks

rock	J_n natural remanent magnetization (10^{-5} electromagnetic units per cubic centimetre)	k magnetic susceptibility (10^{-5} electromagnetic units per cubic centimetre)	ratio* $Q_n=J_n/k·H_{ex}$
Igneous			
granite	10–80	50–400	0.3–1
diabase	190–400	100–230	2–3.5
basalt	200–1,000	100–700	5–10
seafloor basalt (1–6 metres)	500–800	30–60	25–45
typical (average)	10–4,000	5–500	1–40
Sedimentary			
red sediments	0.2–2	0.04–6	2–4
sandstone		1–40	
shale		1–50	

limestone		0.5–20	
typical (average)	0.1–10	0.3–30	0.02–10
Ores			
magnetite ore	300,000–1,000,000	30,000–100,000	~10–50
hematite ore		10–70	

*For external magnetic field (H_{ex}) = 0.5 oersted, the cgs electromagnetic unit of magnetic field intensity.
Source: After Robert S. Carmichael, ed., *Handbook of Physical Properties of Rocks,* vol. II, CRC Press, Inc. (1982).

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Additional Reading

Standard mineralogical reference works include W.A. Deer, R.A. Howie, and J. Zussman, *Rock-forming Minerals*, 2nd ed. (1997); and Annibale Mottana, Rodolfo Crespi, and Giuseppe Liborio, *Simon and Schuster's Guide to Rocks and Minerals* (also published as *The Macdonald Encyclopedia of Rocks and Minerals*, 1978; originally published in Italian, 1977). Useful texts and monographs include Harvey Blatt, *Sedimentary Petrology* (1992); Richard V. Dietrich and Brian J. Skinner, *Rocks and Rock Minerals* (1979); Anthony Hall, *Igneous Petrology*, 2nd ed. (1996); Cornelis Klein, *Minerals and Rocks: Exercises in Crystallography, Mineralogy, and Hand Specimen Petrology* (1989); Cornelis Klein and Cornelius S. Hurlbut, Jr., *Manual of Mineral Science (After James D. Dana)*, 22nd ed. (2002); and Harvey Blatt and Robert J. Tracy, *Petrology: Igneous, Sedimentary and Metamorphic*, 2nd ed. (1996).
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D.H. Griffiths and R.F. King, *Applied Geophysics for Geologists and Engineers: The Elements of Geophysical Prospecting*, 2nd ed. (1981); Robert S. Carmichael (ed.), *Handbook of Physical Properties of Rocks*, 3 vol. (1982–84), also available in a 1-vol. abridged ed., *Practical Handbook of Physical Properties of Rocks and Minerals* (1989); Edgar W. Spencer, *Introduction to the Structure of the Earth*, 3rd ed. (1988); and D.R. Bowes (ed.), *The Encyclopedia of Igneous and Metamorphic Petrology* (1989), may also be consulted.