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Manual of
Rapid Mineral Identification

Volume II

Principles & Practices

Foundational background in practical mineralogy
and the science of mineral identification methods

by Dr. Uwe R. Kackstaetter

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Preface

The Manual of Rapid Mineral Identification is a unique two set volume for use in practical mineralogy and mineral identification. It was written with a broad spectrum of users in mind, from the novice rockhound to the trained professional, including students, prospectors, geoscientists and mineral laboratories. While many books have been published on the identification of minerals, an easy to follow yet accurate approach for identifying unknown samples is commonly lacking. This publication is literally a step-by-step cookbook approach to mineral identification and includes easily administered "how-to methods" some of which are proprietary. In addition, an exhaustive collection of identification tables and charts integrated into the mineral identification process are included.

For the first time it is now possible for a novice to derive at a reasonable positive identification of an unknown sample. Trained professionals and laboratories may find the publication useful as a ready reference for certain steps or methodologies which can be executed with a high degree of accuracy. It might be surprising to the reader that expensive research equipment is not necessary to achieve positive results.

This Volume II of the Manual of Rapid Mineral Identification series, Principles & Practices, is a primer dealing with the foundational background in practical mineralogy and the science of mineral identification methods. It was written for those needing to learn the fundamentals of practical mineralogy and includes common and extended mineralogical topics such as crystallography, optics, unit cells, mineral chemistry, mineral groups and other elementary mineralogical topics. But it also explains hardness, color, chemical impurities, optical properties and variations and much more and describes how to assess them for mineral identification.

Volume I titled Mineral ID Tests and Determinations is an easy to follow step-by-step and hands-on field and laboratory approach to mineral identification. Systematic steps will guide the user to accurate results when followed explicitly. It assumes a basic background in mineralogical testing without necessarily explaining the foundational science behind the methods.

These manuals, no matter how well written, will not be able to replace sound expertise from trained geologists and mineralogists in interpreting the multitude of analytical measurements made during mineral identification procedures. Those who need additional help in providing answers to their geological / mineralogical questions can take advantage of a FREE mineral analytical service as explained below:

FREE MINERAL ANALYTICAL SERVICES: The Department of Earth and Atmospheric Sciences Mineral Laboratory at the Metropolitan State University of Denver offers FREE non-destructive and destructive mineral identification services as part of their geoscientist training program and university community outreach. For details please visit " http://college.earthscienceeducation.net/MIN/MINID.pdf " or use the pictured QR code. Certain restrictions apply. Samples are usually only processed during Spring semester and extended time is needed for sample processing as explained in the link.

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- I am indebted to the former publication of the *Mineral-Rock Handbook, by Paul Dean Proctor, P. Robert Peterson, Uwe Kackstaetter, November 1989, Paulmar Inc.* for sections of text, photos of most rocks & minerals, many tables and some graphics. I am especially grateful for the association with the late Paul Proctor, an outstanding geologist, who has always been an inspiration and geologic father figure for my work.
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1 MINERALS

Minerals are the building blocks of geologic materials. They do not need to be beautiful crystals, but can occur as aggregates or masses. Over 5000 mineral species have been described with more new minerals added every year. However, most of these 5000 minerals are rare. Only about 200 are common enough to make up macroscopic rocks and of these, the most abundant 50 make up 99.9% of the Earth's crust.

In order for a geologic material to qualify as a mineral, it must meet the following 4 definitions:

1. A mineral forms and occurs naturally
2. A mineral is a homogenous solid
    a. Mercury (Hg) and water (H_2O) can also occur in liquid form
3. A mineral has a definite chemical composition
4. A mineral has an ordered atomic arrangement or crystal structure
    a. Debatable are opal and amber, which have no crystal structure and are sometimes NOT included in mineral listings

1.1 MINERAL CRYSTAL STRUCTURES

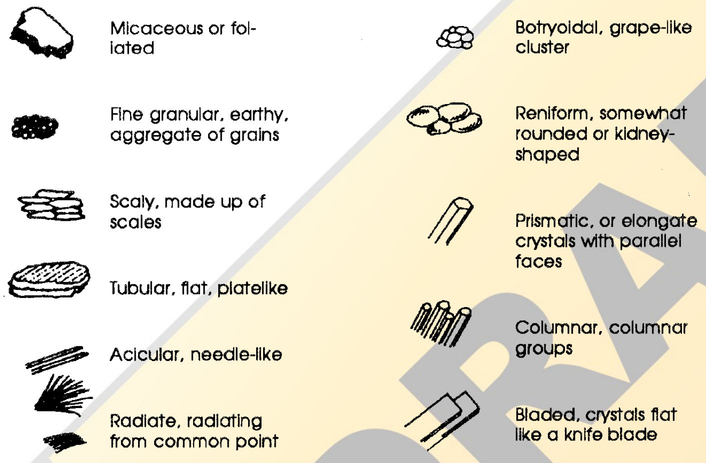


Figure 1.1i - Common mineral crystal structures and crystal aggregates

Only a very few minerals develop into well-formed crystals which come near to resembling the idealized crystal or crystals of the crystallographer. However, many minerals do occur in forms or have a habit of crystallization quite unique to that mineral. Some of the characteristic forms of crystals are sketched in figure 1.1i. The scale of the crystals is usually small, and is measured in millimeters or fractions of an inch. On rare occasions large scale forms may appear in veins and other openings in rocks and in pegmatites, where conditions for growth are favorable. Mica, for example, has a typical "micaceous" or "foliated" appearance when small such as mica grains in an igneous or metamorphic rock, yet it can mass a foot or more in diameter when developing in pegmatites - an igneous rock characterized by large grain size. Other specific minerals almost always crystallize in a radiating form such as wavellite, while others form fibrous, fiber-like aggregates such as chrysotile (asbestos) and satinspar (gypsum). These habits or forms are an expression of the crystal system of the mineral and are very useful aids in identifying the mineral substance.

Many minerals occur in nature with definite and characteristic external geometric forms. These we call crystals. The outward shapes are an expression of an orderly internal atomic arrangement of the elements which compose the crystals. The possible number of crystal patterns is limited by geometric consideration. As a result, all crystals can be classified within one of six possible crystal systems.

Mineral crystals may occur in single forms, or combinations of more than one form within specific crystal systems. In some cases twin crystals may develop of the same mineral and these are intergrown according to definite laws of crystallography.

1.1.1 CRYSTAL GROWTH & FORMATION

In order to form crystals, atoms and molecules that combine must be mobile. This may happen in melts, aqueous solutions or even gases. Crystals will form when the solvent or melt is oversaturated in respect to the solute. Oversaturation can be achieved by:

a. reducing the amount of solvent (e.g.; evaporation)
b. changing solubility of the solute (e.g.; temperature)
c. adding more solute (e.g.; chemical weathering)
d. chemical exchange reactions (e.g.; mixing & deposition)

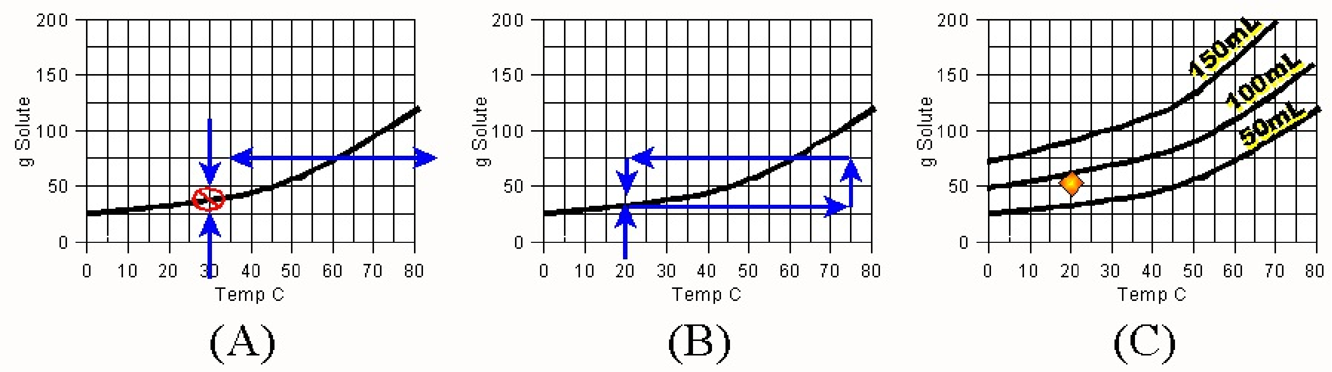
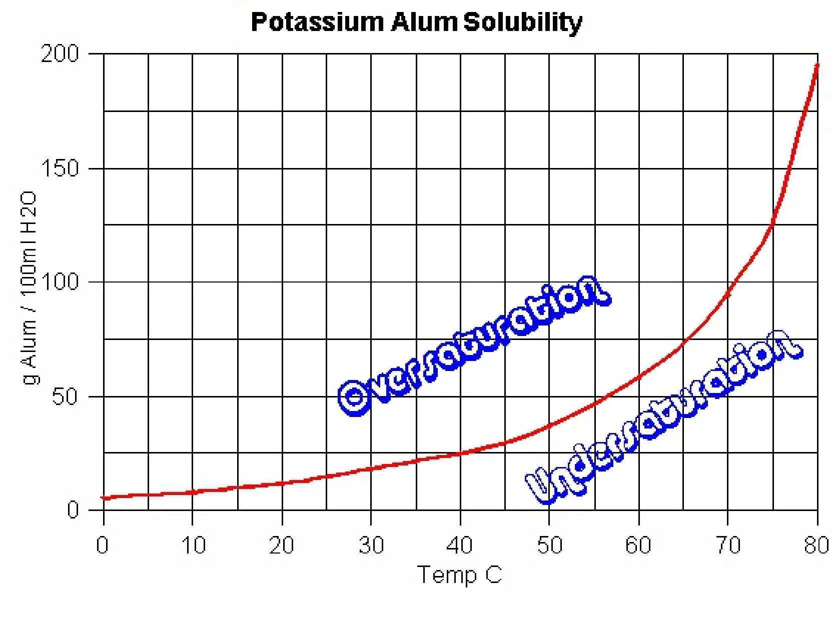


Figure 1.1.1.i - Temperature dependent solubility of an assumed solute. Black curve shows saturated condition with area above curve indicating "oversaturated" and area below the curve "undersaturated" conditions.

Most substances exhibit a temperature dependent solubility. The higher the temperature of the solvent, the greater amount of solute can be dissolved as indicated by the black curves in figure 1.1.1.i. The area above the curves shows oversaturated conditions, while the area below is indicative of undersaturation. The curve itself displays a completely saturated environment. Changes from undersaturated to oversaturated conditions can be achieved by temperature changes as shown in figure 1.1.1.iA by the horizontal blue double arrow. However, changing conditions vertically across the saturation curve is impossible. Any attempt to increase the concentration by adding more solute stops abruptly once the saturation is reached. Excess material will stay undissolved.

In order to reach oversaturation and desired crystal growth, one must increase the temperature as indicated in figure 1.1.1.iB. Here at 20°C (Room Temperature), solute can be added until saturation is reached. Following the outline of the blue arrows, temperature must now be raised (to 75°C in the example) before more solute can be added successfully. Letting the solution now cool back to room temperature allows the crystal growth medium to cross into the oversaturated area and crystals will develop. Crystal growth will remove the oversaturated amount from the solution until saturation is reached at which point all growth stops.



Growing crystals through the process of solvent evaporation is displayed in figure 1.1.1.iC. Creating an undersaturated starter solution and letting the solvent evaporate changes the saturation curves. Eventually the solvent is reduced to a point where oversaturation is reached and crystal growth occurs. Simple crystal growth experiments can illustrate the conditions of crystal formation. Potassium Aluminum Sulfate Dodecahydrate ($KAl(SO_4)_2 \cdot 12H_2O$) is well suited to show crystal growth within a short amount of time. Figure 1.1.1.ii depicts the extreme change in solubility of Potassium Alum depending on temperature. An undersaturated solution reaches quickly oversaturated conditions when the temperature is dropped.

Figure 1.1.1.ii - Potassium Alum Solubility according to temperature

1.1.2 CRYSTAL GROWTH & STENO'S LAW

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Nicolas Steno (1638 - 1686) was a Danish scientist who contributed much to the field of geology. Next to three stratigraphic principles (Law of Superposition, Principle of Original Horizontality, Principle of Lateral Continuity), Steno also introduced a crystallographic principle now known as Steno's Law.

Steno's Law, better known as Steno's Law of Constant Angles, was discovered while observing euhedral quartz crystals in a great variety of shapes and forms. He came to the conclusion that even while crystal shapes may vary greatly, the angles between crystal faces of the same species of mineral show no variation and are indeed constant. This break through discovery lead to the realization of ordered atomic arrangements in any mineral crystal, a basic principle of modern mineralogy.

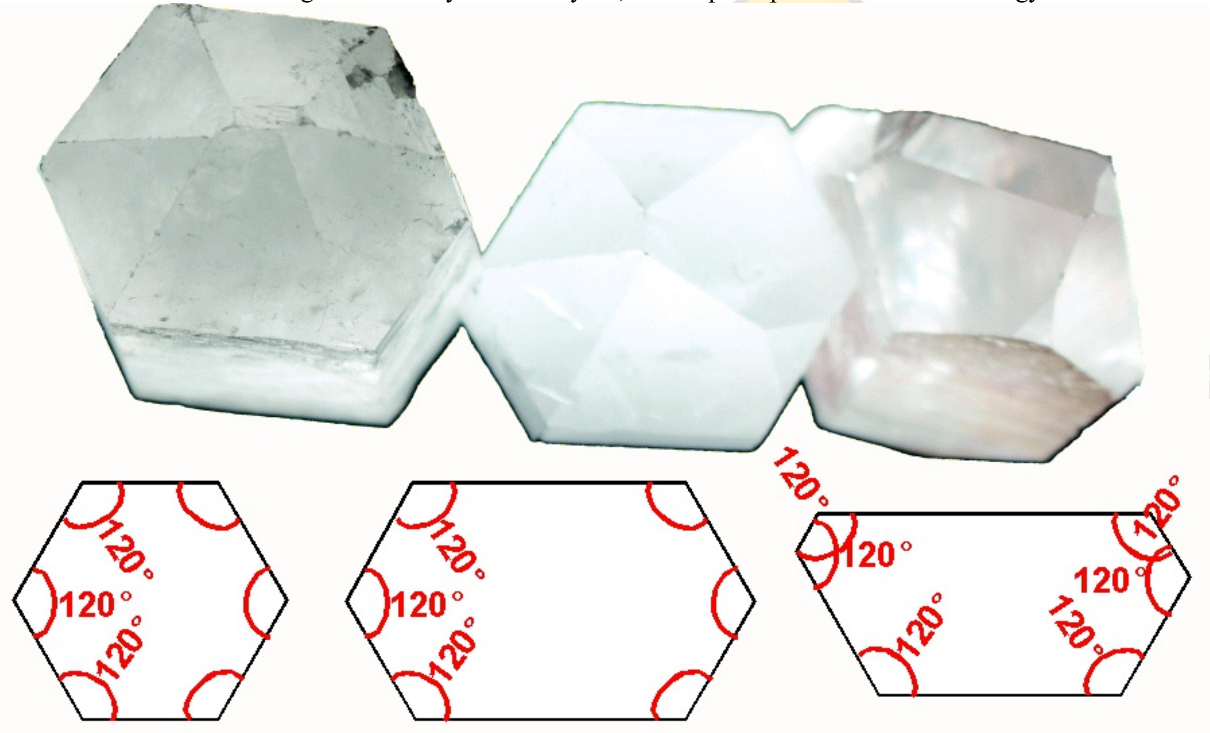

Figure 1.1.2.i - Steno's Law of Constant Angles

Measuring accurate crystal face angles can be accomplished with an optical petrographic microscope using a rotating stage goniometer and associated vernier scale. Crystals grown on a glass slide or very small mineral crystals can be aligned along their faces with the microscope cross hairs. By rotating the stage to the next crystal face, very accurate interfacial angles can be obtained via the goniometer.

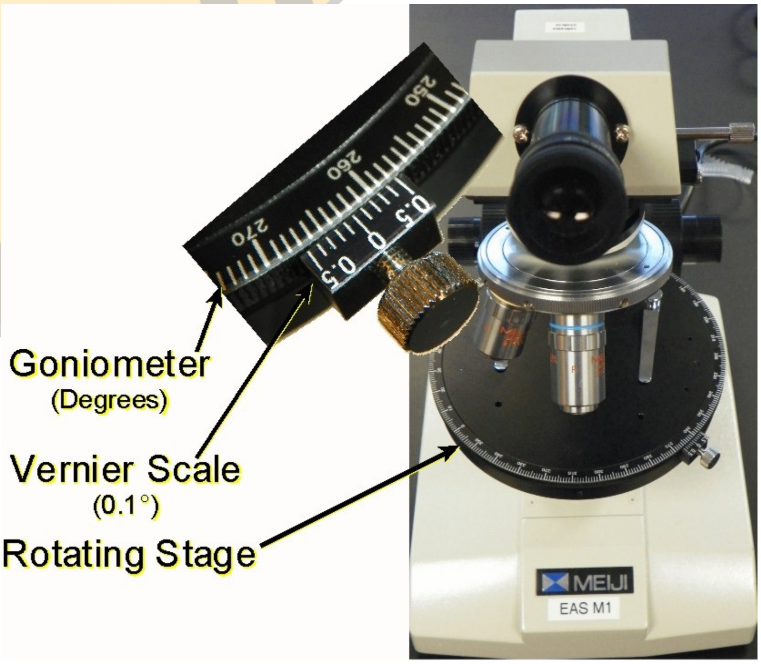

Figure 1.1.2.ii - Optical Petrographic Microscope for measuring accurate angles using a rotating stage goniometer and vernier scale.

1.1.3 CRYSTAL GROWTH & UNIT CELLS

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While minerals have a definite chemical composition, the arrangements of atoms and molecules within a crystal structure is 3 dimensional and repeats in a regular pattern, thus creating a crystalline shape in an ordered and repeatable atomic arrangement. A Unit Cell is a small group of atoms that have a fixed geometry relative to one another. This geometry of atoms is arranged in a 3D box with parallel sides and its own symmetry. These unit cell boxes must be infinitely stackable using only translation. Thus, repetitions of the unit cell make up a crystal with an ordered internal arrangement. According to Klein (2002); “the unit cell of a mineral is the smallest divisible unit of a mineral that possesses the symmetry and properties of the mineral.”

Atoms may be at the corners, on the edges, on the faces, or wholly enclosed in each unit cell box, and each cell in the crystal is identical in geometry and size and atomic arrangement and composition. Unit cell chemical compositions, however, do not equal chemical formulas of the mineral.

In each Unit Cell the crystallographic order of symmetry and crystal class and the chemical formula of the internal atomic arrangements are combined. Both concepts are discussed a little later in this lab manual. To understand crystals and unit cells it is therefore imperative to have a sound understanding of crystal chemistry. A review of basic chemistry (elements, valence, bonding, and the periodic table) is strongly advised.

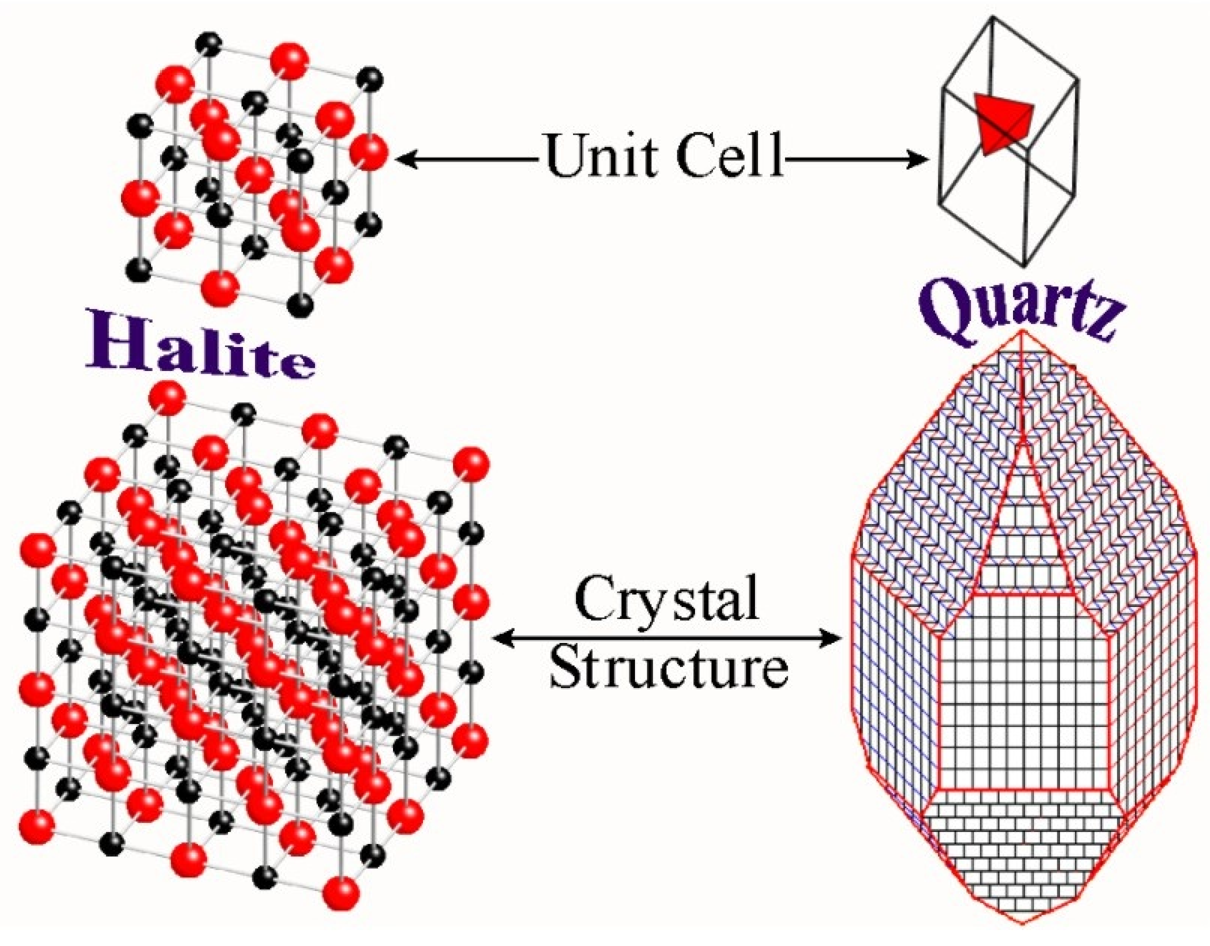

Figure 1.1.3.i - Unit Cell representation and crystal growth of halite and quartz

All unit cells have actual dimensions and side angles. As discussed later, unit cell dimensions can be established using XRD (X-ray diffraction) data. Thus Volume and even theoretical mineral densities can be calculated. Unit cell dimensions and geometry also translate into the various crystal systems discussed in the next section. Below is a quick representation of all possible unit cell systems, their dimensions, angles and other useful parameters.

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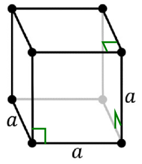
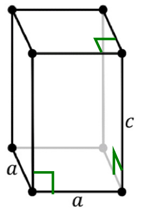
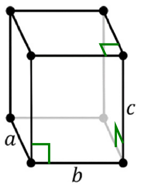
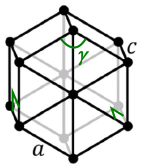
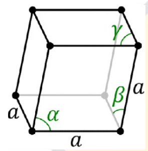
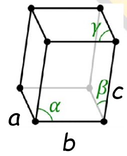
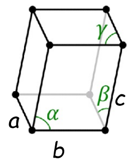
	Three equal axis at right angles a=b=c α=β=γ=90°	Unit cell volume in cubic angstroms: $V = a^3$ Theoretical mineral density: ρ_{calc} (g/cm^3)=[Molecular Weight]×[Z])/([Volume]×0.60225) Molecular Weight (g/mole), Z (formula units/cell), Volume (Å^3) and 0.60225 (Avogadro constant / 1.0 x 10^{24})
	Two equal axis; All three axis at right angles a=b≠c α=β=γ=90°	Unit cell volume in cubic angstroms: $V = a^2c$ Theoretical mineral density: ρ_{calc} (g/cm^3)=[Molecular Weight]×[Z])/([Volume]×0.60225) Molecular Weight (g/mole), Z (formula units/cell), Volume (Å^3) and 0.60225 (Avogadro constant / 1.0 x 10^{24})
	Three unequal axis; all three axis at right angles a≠b≠c α=β=γ=90°	Unit cell volume in cubic angstroms: $V = abc$ Theoretical mineral density: ρ_{calc} (g/cm^3)=[Molecular Weight]×[Z])/([Volume]×0.60225) Molecular Weight (g/mole), Z (formula units/cell), Volume (Å^3) and 0.60225 (Avogadro constant / 1.0 x 10^{24})
	Two equal coplaner axis at 120°, third axis at right angles a=b≠c α=β=90°γ=120°	Unit cell volume in cubic angstroms: $V = a^2c \sin(60°)$ Theoretical mineral density: ρ_{calc} (g/cm^3)=[Molecular Weight]×[Z])/([Volume]×0.60225) Molecular Weight (g/mole), Z (formula units/cell), Volume (Å^3) and 0.60225 (Avogadro constant / 1.0 x 10^{24})
	Three equal axis, equally inclined a=b=c α=β=γ≠90°	Unit cell volume in cubic angstroms: $V = a^3\sin(\alpha)$ Theoretical mineral density: ρ_{calc} (g/cm^3)=[Molecular Weight]×[Z])/([Volume]×0.60225) Molecular Weight (g/mole), Z (formula units/cell), Volume (Å^3) and 0.60225 (Avogadro constant / 1.0 x 10^{24})
	Three unequal axis; one pair not at right angles a≠b≠c α=γ=90°≠β	Unit cell volume in cubic angstroms: $V = abc \sin(\beta)$ Theoretical mineral density: ρ_{calc} (g/cm^3)=[Molecular Weight]×[Z])/([Volume]×0.60225) Molecular Weight (g/mole), Z (formula units/cell), Volume (Å^3) and 0.60225 (Avogadro constant / 1.0 x 10^{24})
	Three unequal axis, unequally inclined and none at right angles a≠b≠c α≠β≠γ≠90°	Unit cell volume in cubic angstroms: $V = abc\ (1- \cos^2 \alpha - \cos^2 \beta - \cos^2 \gamma) + 2(\cos(\alpha) \cos(\beta) \cos(\gamma))^{½}$ Theoretical mineral density: ρ_{calc} (g/cm^3)=[Molecular Weight]×[Z])/([Volume]×0.60225) Molecular Weight (g/mole), Z (formula units/cell), Volume (Å^3) and 0.60225 (Avogadro constant / 1.0 x 10^{24})

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1.2 CRYSTAL SYSTEMS

The six crystal systems are isometric, tetragonal, orthorhombic, monoclinic, triclinic, and hexagonal and correspond nicely with the unit cell systems previously discussed. The system names are indicative of the characteristics of imaginary crystal axes which lie within the crystals. We measure three axes for descriptive purposes of man made objects such as length, width, and depth. In crystal nomenclature three imaginary crystal axes are also established. The position and orientation of these imaginary axes inside the crystal shape classifies the respective crystal systems. The central, length or vertical axis is labeled the "c" axis, while the other two dimensional axis are called "a" and "b". (see figure 1.2.i). Crystal faces can also be classified using Miller indices, named after British mineralogist William Hallowes Miller (1801 – 1880). Miller's three-number code describes the crystal faces according to their position in regards to the crystal axes, using the integer sequence referred to as "h k l", where "h" is the cut with the "a" axis, "k" indicates the cut with the "b" axis and "l" refers to the intersection of the crystal face with the "c" axis . If a crystal face cuts one axis at right angles and parallels the other two (in a cube for example as shown in figure 1.4), it is numbered 100 (1 = cuts; 0 = parallel), thus indicating that the crystal face cuts the "a" axis but parallels axis "b" and "c". The last of the three digits refers always to the vertical imaginary or "c" axis. Hence a 001 face in a cube would parallel the two horizontal axes and cut the vertical one.

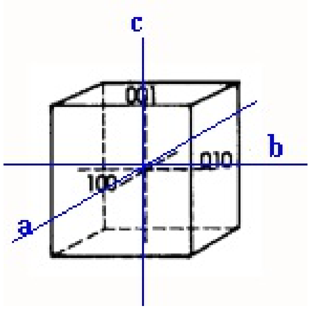


Figure 1.2.i - three number code of crystal faces in a cube

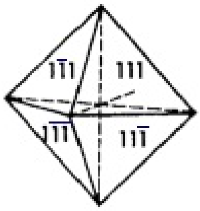
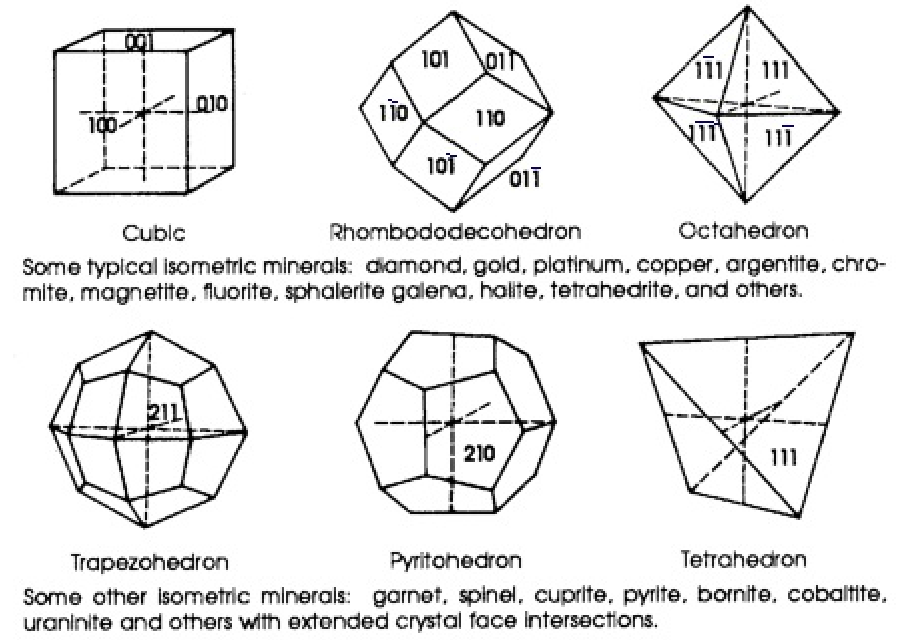


Figure 1.2.ii - three number code of crystal faces in an octahedron

But how do we distinguish between the front crystal face labeled 100 and the crystal face exactly opposite in a cube. It is indeed a different crystal face and as such should have its very own Miller indices. This can be accomplished by picturing the crystal axis as an x,y,z grid, where we have positive values in one direction and negative ones in the opposite direction. Thus the crystal faces cutting the positive side of the axis are labeled as "1", the opposite side which is cutting the negative end would be labeled "-1" or "$\bar{1}$" as shown in figure 1.2.ii.

What about an octahedron (see figure 1.2.ii)? A 111 octahedral face would have eight possible repeats and cut the edge of intersecting faces equally and axes at the same distance from their common point in space. Some crystal faces may not visibly cut an axis. The trapezohedron in figure 1.6 cuts the horizontal axis if extended right, but it is twice as far away as for regular intersecting axes and crystal faces. The farthest intersection is labeled 1 (1 = cut), the other number reflects the fraction of the distance needed to make a regular cut (2 = cut at ½), and is numbered 211. Proportional distances 3, 4 or 5 times greater have equivalent fractions to reflect the distances. The well-known mineral galena, for example, commonly develops cubic, octahedral, and even dodecahedral (12 faces) in it's crystallization habit (figure 1.2.iii).



Some typical isometric minerals: diamond, gold, platinum, copper, argentite, chromite, magnetite, fluorite, sphalerite galena, halite, tetrahedrite, and others.

Some other isometric minerals: garnet, spinel, cuprite, pyrite, bornite, cobaltite, uraninite and others with extended crystal face intersections.

Figure 1.2.iii - Examples of the Isometric crystal system. Mineral forms and common mineral axes in the system shown by dashed lines.

Miller Indicies for irregular faces can be calculated, thus making the assignment of crystal faces with numbers 2, 3, 4 or more easier. The following table explains the calculation procedure:

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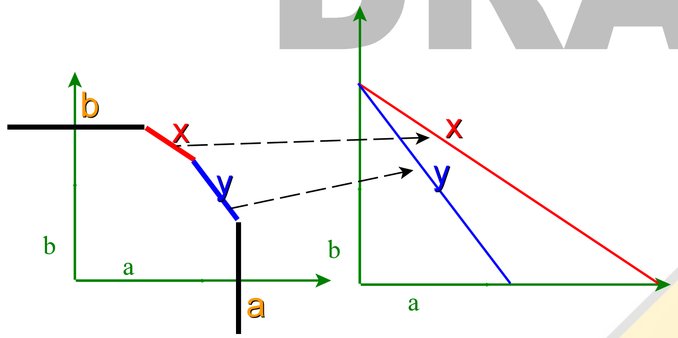
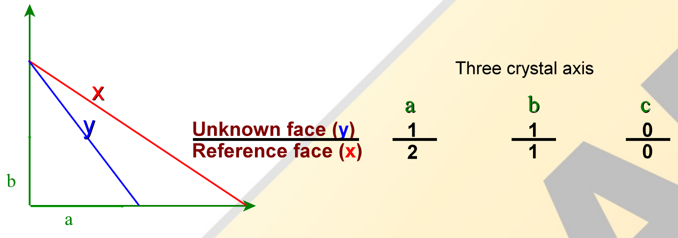


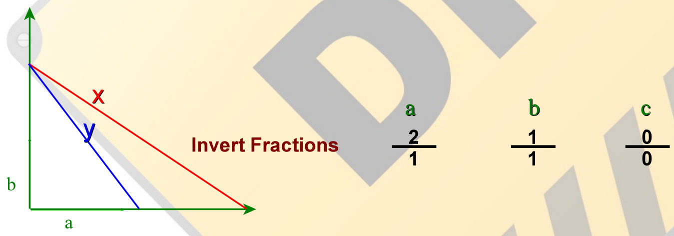
Step 1: Draw the crystal axis and faces and transpose to a 2 dimensional grid system.



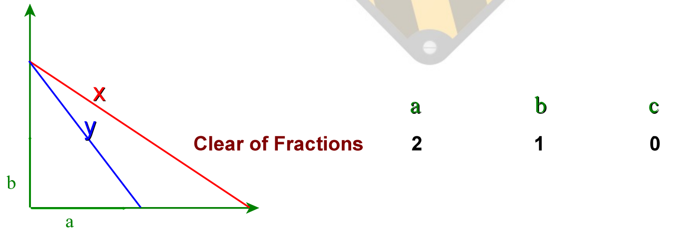
Step 2: Pick a reference crystal face and set up as fraction with the unknown face (Always "unknown" over "reference"). Example: Reference x!

Determine if the crystal faces cut at the same point (1) or are separated by significant distance differences (2 or more times apart).

Mark 0 for crystal faces that do not intersect any axis.



Step 3: Invert the fractions.



Step 4: Clear of fractions. The result is the Miller Index of your unknown crystal face.

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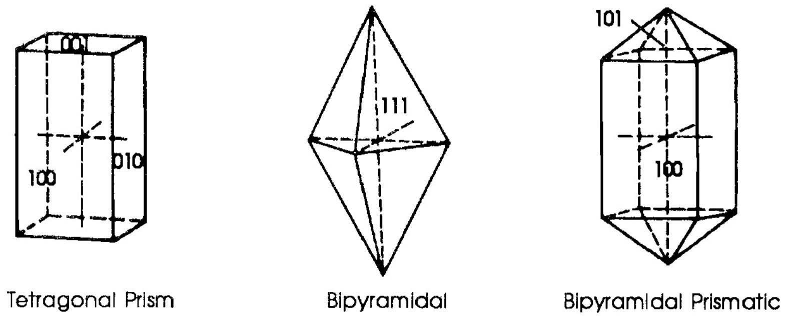
1.2.1 ISOMETRIC CRYSTAL SYSTEM

In the **Isometric system** three axes are at right angles to each other and are of equal length (Iso = equal, metron = measure). The Isometric crystal system has many forms and shapes and high order symmetry. It must always have, however, three equal length imaginary axes at right angles to each other. Some crystal forms of the Isometric system are shown In figure 1.6. Three imaginary crystal axes are sketched in all but one of the crystals. Visualize where they would be in this rhombododecohedron model.

1.2.2 TETRAGONAL CRYSTAL SYSTEM

The **Tetragonal System** (Tetra = four; gonon = angle) has also three imaginary crystal axes at right angles to each other. However, while the two horizontal axes are of equal length, the vertical axis is longer or shorter than the other two. Zircon, for example, crystallizes in the tetragonal system. A common crystal form is a 4-sided pyramid which may have vertical sides or prismatic faces, and an upside down pyramid. Figure 1.7 shows three crystal forms in the tetragonal system and gives a short list of some common tetragonal minerals. This crystal system contains the fewest of all minerals.

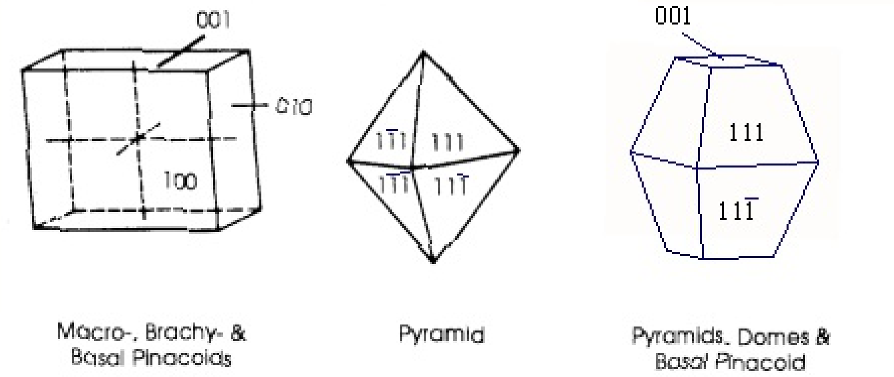


Common tetragonal minerals include: cassiterite, chalcopyrite, rutile, scheelite, thorite, idocrase, wulfenite, zircon and others.

Figure 1.7 - Some crystal forms in the Tetragonal crystal system.

1.2.3 ORTHORHOMBIC CRYSTAL SYSTEM

The **Orthorhombic System** (orthos = straight) has again three imaginary crystal axes at right angles to each other. However, each one has a different length. In the realm of man-made objects, a brick or shoe box would belong to the orthorhombic system for example. In the world of minerals, the common mineral barite crystallizes in the system. Figure 1.8 shows some typical crystal forms in the system, inclusive a brief list of some common orthorhombic minerals.



Common minerals in the system include: aragonite, azurite, barite, brochantite, cerussite, columbite, enargite, staurolite, stibnite, strontianite, marcasite, olivine, topaz and others.

Figure 1.8 - Some crystal forms In the orthorhombic crystal system.

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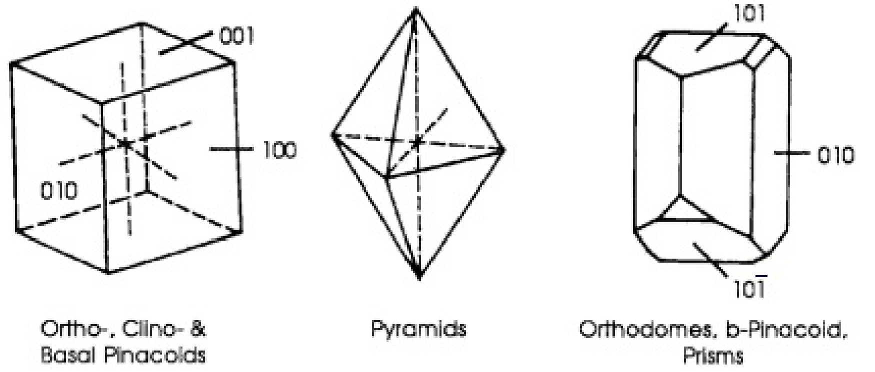
1.2.4 MONOCLINIC CRYSTAL SYSTEM

The **Monoclinic System** (mono = single, cline = inclined) has three crystal axes of unequal length, as is the case in the orthorhombic system as well. While two of the axis in the monoclinic system intersect at right angles to each other, the third axis does not, but intersects the other two obliquely. The gypsum crystal exhibits these characteristics as well as crystal forms of biotite and muscovite micas. It should be noted that the oblique angle of the third axis in monoclinic minerals can be very small, as low as 1° in muscovite mica. Common crystal forms and minerals in the system are shown in figure 1.9.

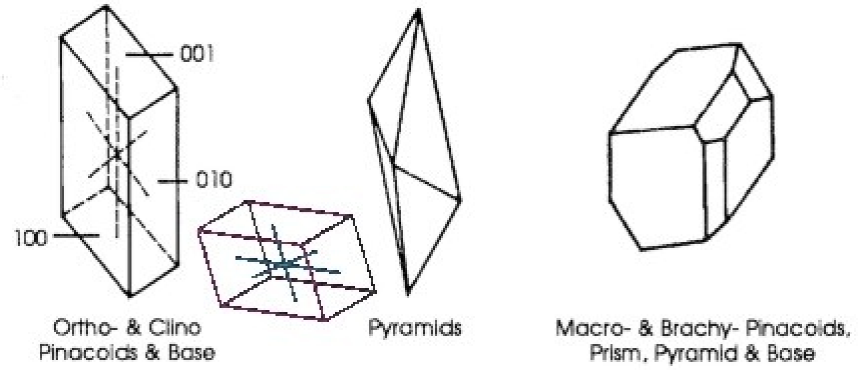


Augite, biotite, crocoite, gypsum, hornblende, malachite, monazite, muscovite, orthoclase, tenorite, titanite, wolframite and numerous others are examples.

Figure 1.9 - Common mineral forms in the monoclinic system. Drawings are shown with very small oblique angle of the third axis.

1.2.5 TRICLINIC CRYSTAL SYSTEM

The **Triclinic System** (tri = three, cline = inclined) has three crystal axis of varying length and all three intersect not at right angles but obliquely to each other. Graphically, this makes this crystal system a challenge to represent in form of a drawing. Kyanite, plagioclase and rhodonite crystallize within this system. Some crystal forms in the triclinic system are shown in figure 1.10 together with a list of some common minerals.



Plagioclase, kyanite, rhodonite, turquoise, ulexite and axinite are common minerals in system.

Figure 1.10 - Some crystal forms in the triclinic system.

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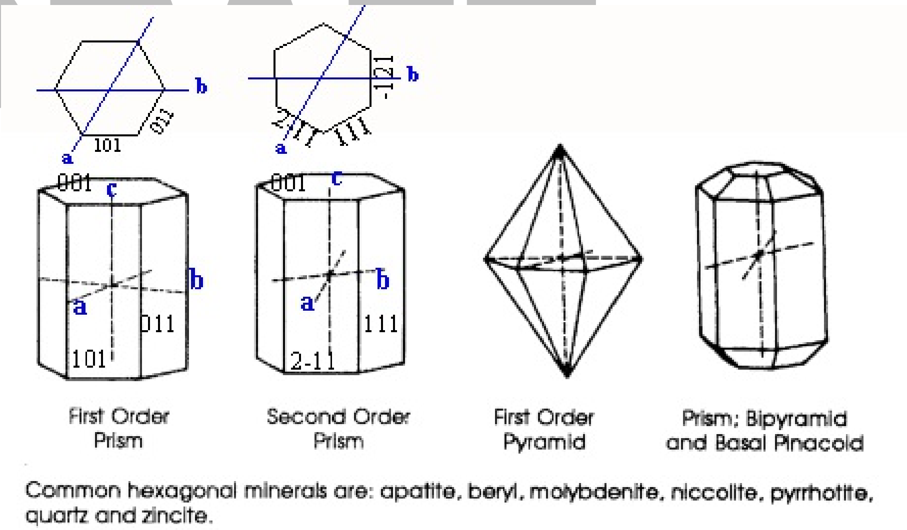
1.2.6 HEXAGONAL CRYSTAL SYSTEM

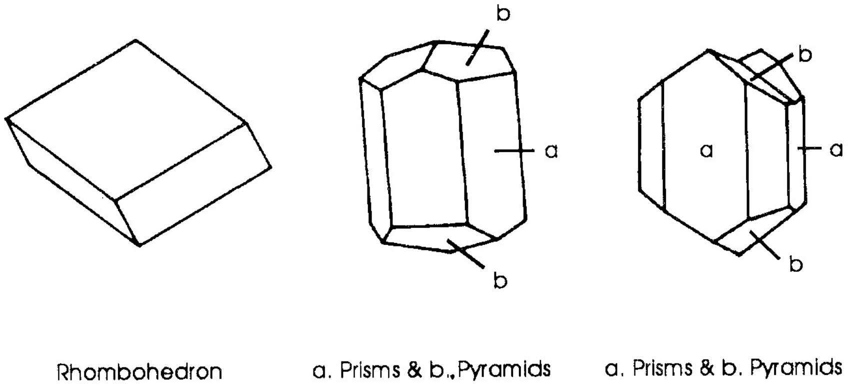
In the old **Hexagonal System** (hexa = six; gonon = angle) four crystal axes were employed instead of the usual three, thus inserting another integer into Miller's *hkl* classification, namely *hkil*. In the new system, the regular *hkl* designation is used. Here, the *a* and *b* axes are equal in length and in a horizontal plane, intersecting in angles of 120 degrees to each other (see figure 1.11.) The *c* axis is of different length and at right angles to the others and in the vertical plane of the crystal. The common mineral quartz shows the distinctive characteristics of the hexagonal system (figure 1.11).



Common hexagonal minerals are: apatite, beryl, molybdenite, niccolite, pyrrhotite, quartz and zincite.

Figure 1.11 - Hexagonal crystal forms with crystal axes.

Note that hexagonal first order and second order prisms differ. In first order prisms the *a* and *b* axes intersect the crystal faces at their edges. In second order prisms intersection of crystal axes occurs at the central part of four of the crystal faces.



Some common minerals in the class: chabazite, brucite, beryl, alunite, calcite, dolomite, tourmaline, rhodochrosite, siderite, corundum, hematite, cinnabar

Figure 1.12 - Examples of rhombohedral class crystals in the hexagonal system.

An odd branch of the hexagonal system is the rhombohedral class (figure 1.12). At first glance a simple rhombohedron locks like a triclinic crystal. Calcite, which displays rhombohedral hexagonal characteristics is indeed often mistaken for such a triclinic geometrical form. However, when extending the main or c-axis through the most opposite corners of a rhombohedron, rotation around this c-axis will reveal perfect 6-folded rotation symmetry. Some famous gemstones such as emeralds (beryl), rubies and sapphires (corundum), and tourmaline belong to the rhombohedral hexagonal crystal class.

The crystal forms illustrated are only a few of the very large number possible and known in nature. Yet, these can be very useful guides as you study and try to identify various natural crystals. The minerals pyrite and garnet and their various crystal forms exemplify some of the numerous possibilities. Pyrite may develop cubic, pyramidal or pyritohedral forms. In some cases the different crystal forms may be intermixed In the same mineral deposits.

Of considerable significance in the development of crystal faces is the *law of the constancy of interfacial angles*, also known as Steno's Law (Nicholas Steno, 1638-1686). Wherever a mineral species occurs, whether natural or laboratory created, the angle of intersection between like faces on the crystals is essentially constant. This is a direct expression of the internal atomic arrangement, and is an Important diagnostic feature of the species. This law applies irrespective of the size of the crystal formed. A mechanical goniometer, a semicircular protractor with a movable arm, is used to measure these angles. The reflecting goniometer, a more sophisticated instrument, measures the angles through the use of reflected light from the crystal faces.

1.3 CRYSTAL TWINNING

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Two or more crystals may be intergrown in various ways. Certain parts of the different crystals may be parallel and other parts of the individuals are in reverse position in relationship to each other. These relationships relate to a deducible law of growth and the final crystal product is a twin crystal. Twinning may occur over a plane common to both and be called a twin plane. A point may have two individual faces symmetrical to it and the twin is then called a twin center. Twin laws indicate whether the twinning is by center, axis or plane and give the crystallographic orientation for either the plane or axis.

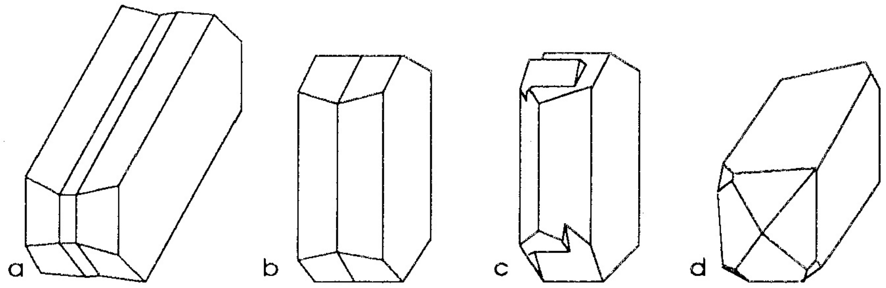

Figure 1.13 - Some crystal twinning types: a) albite type twinning. b) contact twin. c) inter-penetration twin. d) saveno type twin

Twin crystals may take the form of contact twins or penetration twins. Contact twins have a definite composition surface which separates the two individuals. The twin law is defined by the twin plane. Albite feldspar (plagioclase) beautifully shows this type of twinning, and the name albite twinning is used (figure 1.13a). Inter-penetrating crystals have an irregular composition surface (figure 1.13c). The twin axis (line) defines the twin law. Multiple twins and repeated twins, of three or more parts, are all twinned according to the same law. Parallel successive composition surfaces form polysynthetic twins. If the successive composition planes are not parallel, cyclic twins result. Sometimes the Individuals In polysynthetic twinning are so closely spaced that when a crystal face or cleavage plane crosses the composition planes, distinct striations show because of the reversed positions of adjacent crystals.

Table 1.1 - Most common twins applicable to the different crystal systems.

Crystal System	Twin Planes	Penetration Twins	Twin Axis
Isometric	X	X	X
Tetragonal	X		
Hexagonal (Rhombohedral)	X	X	X
Orthorhombic	X		
Monoclinic	X	X	X
Triclinic	X		X

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1.4 MINERAL CHEMISTRY

All minerals are composed of atoms from those consisting of single atomic composition such as gold to those of complex composition involving several different types of atoms. The latter, based on electrical charge, are divided into ions. Atoms and molecules with a positive charge are called cations, while those with a negative charge are referred to as anions. In the Berzellan classification of minerals, chemical composition is the main basis. Minerals are placed in chemically related classes depending on the dominant anion (negative charged ion) or anionic group. Such an anionic group may consist of a combined C (carbon) plus 3 O (oxygen), creating the polyatomic anion $CO_3{}^{2-}$, which is the basis of the carbonate group.

A person desiring to know more about minerals should understand the chemical components, or elements, of which they are composed. They should also know something about atoms, of which elements are made, and their general structure, since an atom is the smallest subdivision of matter which retains elemental characteristics. Yet, even the atom is subdivided into smaller components such as a very small nucleus made up of protons and neutrons surrounded by electrons spread through a much larger space. Let us examine these various components of matter and see how they determine the kind and character of the minerals we study.

Almost 200 years ago Dalton, an English chemist, after intensive research, proposed that all elements consist of very small particles called atoms. He further stated that for a given elemental substance the atoms are alike and exhibit identical chemical properties. Different elements have different chemical properties, and in an ordinary chemical reaction, no atom disappears or changes into another element. All 109 elements currently known to man are listed in a unique table called the "Periodic Table of Elements". Of those, only 92 are naturally occurring and only 77 of them are present in detectable average concentrations in the earth's crust and thus of importance in geology (see table 1.2).

Table 1.2 - Abundance of elements in the earth's crust in decreasing ppm (parts per million) or grams / metric ton, their valence or ionic charge and their ionic radius (IR) or size

Element	Abundance (ppm)	Valence	IR	Element	Abundance (ppm)	Valence	IR	Element	Abundance (ppm)	Valence	IR
O	445 000	-2	1.40	Nd	40	+3	1.00	Ge	1.5	+4	0.53
Si	272 000	+4	0.41	La	35	+3	1.06	Ho	1.3	+3	0.90
Al	83 000	+3	0.54	Y	31	+3	0.90	Mo	1.2	+4	0.65
Fe	62 000	+2; +3	0.86; 0.65	Co	29	+2	0.75	W	1.2	+4	0.62
Ca	46 600	+2	0.99	Sc	25	+3	0.75	Tb	1.2	+3	0.92
Mg	27 640	+2	0.72	Nb	20	+5	0.69	Tl	0.7	+1	1.50
Na	22 700	+1	1.02	N	19	+5	0.13	Tm	0.5	+3	0.87
K	18 400	+1	1.38	Ga	19	+3	0.62	I	0.46	-1	2.2
Ti	6 320	+4	0.61	Li	18	+1	0.76	In	0.24	+3	0.80
H	1 520	+1	0.012	Pb	13	+2	1.32	Sb	0.2	+5	0.76
P	1 120	+5	0.38	Pr	9.1	+3	1.01	Cd	0.16	+2	0.97
Mn	1 060	+2	0.46	B	9.0	+3	0.23	Ag	0.08	+1	1.26
F	544	-1	1.33	Th	8.1	+4	0.97	Hg	0.08	+2	1.02
Ba	390	+2	1.35	Sm	7.0	+3	0.96	Se	0.05	+2	0.50
Sr	384	+2	1.12	Gd	6.1	+3	0.94	Pd	0.015	+2	0.86
S	340	-2	1.84	Er	3.5	+3	0.88	Pt	0.01	+2	0.63
C	180	+4	0.15	Yb	3.1	+2	0.86	Bi	0.008	+3	1.03
Zr	162	+4	0.72	Hf	2.8	+4	0.71	Os	0.005	+4	0.63
V	136	+3	0.59	Cs	2.6	+1	1.67	Au	0.004	+1	0.85
Cl	127	-1	1.81	Br	2.5	-1	1.96	Ir	0.001	+4	0.63
Cr	122	+3	0.52	U	2.3	+4; +6	1.08; 0.52	Te	0.001	+2	0.97
Ni	99	+2	0.69	Sn	2.1	+4	0.69	Re	0.0007	+4	0.56
Rb	78	+1	1.52	Eu	2.1	+2	0.95	Ru	0.0001	+3	0.68
Zn	76	+2	0.74	Be	2.0	+2	0.35	Rh	0.0001	+2	0.68
Cu	68	+1; +2	0.9; 0.73	As	1.8	+5	0.58				
Ce	66	+3	1.67	Ta	1.7	+5	0.72				
Common Polyatomic Ions:				SO_4		-2	2.30	CO_3		-2	1.50

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The different elements can react with each other and combine to form compounds according to chemical laws. A compound is a substance made of two or more elements combined in definite and constant amounts expressed as whole numbers or simple fractions. In water, for example, there are always two atoms of hydrogen (H) to one of oxygen (0), represented in the famous chemical formula H_2O, supporting the law of constant composition. Dalton also concluded that when two elements combine to form two different compounds, the weights of one which combine with a fixed weight of another is always in a simple ratio of whole numbers, e.g., 1 :2, 3:2, etc. This is the law of multiple proportions.

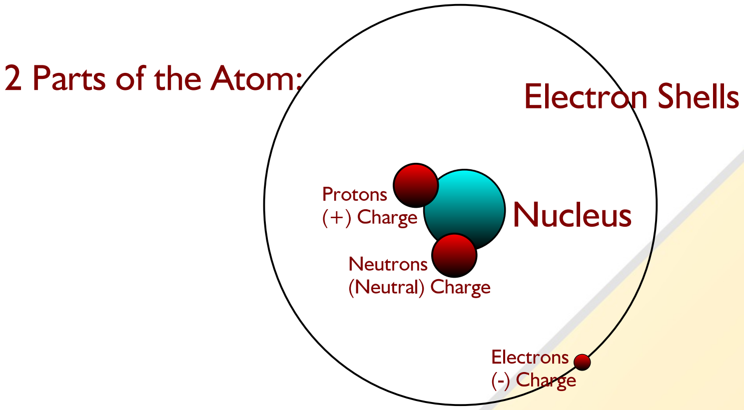

Figure 1.14 - Basic atom model showing various subatomic components

Minerals can be chemically classified as either compounds or pure elements (native minerals). Compounds (combined elements) are kept together by chemical bonds. In order to understand how these bonds are formed, we must take a quick look at the structure of an atom. An atom consist of an atomic nucleus and various electron or energy shells. The nucleus contains protons (positively charged) and neutrons (neutral charge) which in turn a made of up-quarks and down-quarks. Electrons (negative charge) reside on the outer edge of the atom in electron or energy shells.

THE NUMBER OF PROTONS IN AN ATOM DEFINES THE CHEMICAL ELEMENT

An atom with exactly 82 protons is a lead (Pb) atom. If we could selectively remove 3 protons from each lead atom to bring the total proton number down to 79, we would literally turn lead into gold, since gold (Au) has 79 protons. Unfortunately this is impossible. However, gold is gold because it has 79 protons in it's atoms, lead is defined by 82 protons in it's atomic nucleus and iron (Fe) contains 26 protons in each of it's atoms. While the number of protons does NOT vary for each given element, the number of neutrons and electrons may change without changing the element.

ISOTOPES ARE SAME ELEMENTS WITH DIFFERENT NUMBERS OF NEUTRONS

An **isotope** is a specific element with different neutron counts in it's nucleus. Hydrogen (H) for example usually does not have neutrons. Nevertheless, certain Hydrogen atoms with 1 or 2 neutrons do exist, being called deuterium and tritium respectively. Both deuterium and tritium are isotopes of hydrogen. Hydrogen, deuterium and tritium would be chemically indistinguishable. While isotopes have therefore no influence in the reacting or combining of elements, their atomic masses differ, thus making the element heavier or lighter than usual. Most isotopes of common elements are also radioactive.

Isotopes are usually named by their unique atomic mass. The average atomic masses are routinely printed in the periodic table of elements. Isotopes deviate from these averages. For example, carbon (C) has an average atomic mass of 12. The heavier carbon isotope C-14 is labeled as such. While isotopes are less interesting in the identification of minerals, they play an important role in the radiometric dating (age determination) of geologic materials.

ELECTRON CONFIGURATION DETERMINES THE REACTION AND BONDING BETWEEN ELEMENTS

The **electrons** and their configuration (position on the energy level) are important in the bonding of various chemical compounds. The total number of electrons in a chemical element commonly corresponds with the numbers of protons. While protons have a positive charge, electrons are negative particle. Thus the charges equalize and the resulting total charge of an atom is therefore zero or neutral. However, unlike protons, electrons can be removed from one atom and added to another, resulting in an unbalanced charged either positive or negative. An atom that loses electrons has a surplus of positive charges in the protons and therefore becomes positive. Gaining of electrons will yield an overabundance of negative charges and the whole atom becomes negatively charged. Charged atoms are called ions (cations = positive; anions = negative). We may apply some generalizations to cations and anions. For example, the size of the ion depends on the state of ionization. Cations, generally, are smaller, more rigid, and less expansible than anions. An example is the marked contrast between sulfur (S^{-2}) with an atomic size of 1.85 Å and (S^{+6}) at size 0.30 Å. Anions are larger, more easily polarized, and less likely to have variation in size with change in valence, or charge, than cations.

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The charges mentioned are referred to as oxidation numbers or valences and are summarized in table 1.2 for elements of interest in mineralogy. Since opposite charges attract, anions usually combine with cations and visa versa to create a variety of differing chemical compounds and in our case, a variety of minerals. In order to derive at the correct combination of such compounds, the anionic and cationic charges must balance out. For example, the chemical element sodium (Na) has a charge of +1 (Na^{+1}; see table 1.2). In order to make a compound it must find a negative counterpart, such as chlorine (Cl) with a charge of -1 (Cl^{-1}; see table 1.2). Both charges are 1 and therefore cancel out. The resulting compound is NaCl, known to the chemist as *sodium chloride* and to the geologist as the natural occurring mineral *halite*, also known as rocksalt.

But what happens when charges are unequal yet opposite? Can a compound still be created? Lets take a look at silicon (Si) with a charge of +4 and oxygen (O) with a -2 valence. At first glance +4 and -2 do not neutralize each other, leaving as with a surplus +2 charge. But keep in mind that in chemical reactions we are not dealing with single atoms, but with billions and trillions of atoms all reacting together. It is therefore easy to imagine that more than two atoms might be involved in forming a compound. If two oxygen atoms (each -2) react with one silicon atom (+4), the charges do indeed cancel out (-2 + -2 = -4 oxygen; +4 silicon), thus creating the compound SiO_2, *silicon oxide* to the chemist or the mineral *quartz* to the geologist. The subscript denotes the number of atoms involved in the reaction, two oxygens and one silicon.

The brief discussion above is a prologue of how elements are bound together to form minerals. From the forgoing we recognize the importance of electrical charges among the elements. So it is in mineral formation. The intensity and type of these electrical charges determine many of the physical and chemical properties of minerals. If the electrical bond is strong, the crystal is hard, the boiling point is higher, and thermal expansivity is less because of the strong electrical bonds. Such factors help to explain the differences in properties between graphite (C; very soft) and diamond (C; very hard).

The electrical forces, or chemical bonds, belong to one of four principal bond types: ionic, covalent. metallic and van der Waals', with transitions between all types. In mineral genesis nature acts to form the best geometry of structural units in space combined with electrical neutrality and minimum lattice energy. These arrangements determine the physical and chemical properties of the mineral.

THE IONIC BOND - All atoms strive to achieve a stable configuration, which is to have all possible electron sites filled. As noted, helium, neon, and argon have achieved this state, as well as krypton and xenon. When elements have room to gain one or more additional electrons and form a stable configuration, the resulting total charge will be negative as explained above. Chlorine (Cl$^-$), bromine (Br$^-$), and others are examples. Some elements form divalent anions with a double negative charge, such as oxygen (O^{-2}). sulphur (S^{-2}) and others. A positive charge results when electrons are lost in order to have a stable electron configuration. Examples would be potassium (K^{+1}) losing one electrons or calcium (Ca $^{+2}$) losing two. When minerals form in molten volcanic rock, continuous collision keeps them separate until the temperature is lowered enough that the attraction of opposite electrical charges of the ions exceeds the collision disruptive forces. When a large enough number of such cations and anions attach themselves to a growing nucleus, crystal growth occurs. This nucleating mass is in essence the embryonic mineral with all of its characteristic features. The type of bonding where opposite electric charges keep the atoms in a mineral together is defined as the ionic bond. Most materials that are ionically bonded are also soluble in water.

The bond strength is the amount of energy required to break the bond and is dependent on center-to-center spacing of the ions and their total charge. If this spacing or the interionic distance is small, e.g., NaCl with spacing 2.81 Å versus a NaI (Sodium iodide) with 3.23 Å, the melting point decreases from 8010 to 6510° C, respectively, indicating an inverse relationship. Likewise, hardness of a mineral is dependent on interionic distance (ID). BeO (beryllium oxide) with an ID of 1.75 Å has a hardness of 9.0 on the Mohs scale, yet BaO (barium oxide) with 2.75 Å has a hardness of only 3.3 Mohs hardness.

IONICALLY BONDED MINERAL PROPERTIES - Some properties of ionic-bonded crystals are moderate hardness, moderate specific gravity, fairly high melting and boiling points, and poor conductors of heat and electricity. Because of the uniformity of the electrostatic charge, the resultant crystal has high symmetry.

THE COVALENT BOND - Electron sharing between elements develops the strongest of the chemical bonds. Chlorine, as an example, is a very chemically reactive anion and seizes the nearest ion, usually another chlorine ion. The two ions interact and the same resultant stable configuration develops as in an inert gas. Some elements, especially those toward the middle of the periodic table such as carbon (C), silicon (Si), aluminum (Al) and sulfur (S), have two, three and four vacant electron spaces. Not all of the bonds are used in linking just to a neighboring atom. These elements tend to link together in covalent bonding with adjacent atoms and form fixed forms of set dimension. These in turn link together and form aggregates or groups.

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We earlier noted the contrast between the polymorphs diamond and graphite. In diamond the carbon atom has four electon vacancies. These are shared with four other carbon cations and form a very stable, firmly bonded tetrahedral form with carbon atoms at the four apices, or pointed ends. The bond is very strong in the vicinity of the shared electrons and results in a very rigid and polarized structure, forming the hardest known natural substance. Yet, in the mineral graphite, with exactly the same chemical composition, cohesion between carbon cations through strong covalent bonding is limited within a sheetlike structure. Here, the van der Waals' bond, a much weaker force, keeps the sheets together. Hence, there is an ease of excellent cleavage. or parting, along this weakly bonded sheet interface and the mineral is relatively soft and of greasy feel. Such examples indicate that minerals have more than one bonding force. As a general rule the alkali halides and alkali oxides tend to have ionic bonding, while like-atoms or atoms close together in the periodic table will be covalent.

COVALENTLY BONDED MINERAL PROPERTIES - Electron sharing in atoms within minerals results in general insolubility, a greater stability and therefore hardness, very high melting and boiling points and non-conducting of electricity. The local and sharply directional bonds formed result in a lower symmetry for formed crystals than when ionic bonding occurs.

THE VAN DER WAALS' BOND - The very weak bond which ties neutral molecules (one or more combined atoms) and neutral structural units together into a lattice through small residual charges on their surfaces is the van der Waals' bond.

VAN DER WAALS' BONDED MINERAL PROPERTIES - The van der Waals' bond is most un- common in minerals and if present, such as in graphite and mica, defines a zone of ready cleavage and low hardness.

THE METALLIC BOND - When atomic nuclei are bound together by the aggregate electrical charges of the electrons that surround them, then a metallic bond occurs. The electrons are so weakly tied to the metal structure that they may drift through it, or even out of it without disrupting the structure, thus making metallically bonded compounds excellent conductors of electricity. As suspected, all true metals therefore have metallic bonds. Introduction of outside energy, even sunlight, may actually free a large number of electron from certain metals, causing the so-called photoelectric effect prevalent in photo voltaic sheets of solar collector panels.

METALLICALLY BONDED MINERAL PROPERTIES - The close metallic bonding results in high plasticity, ductility, tenacity, electrical conductivity, and low hardness. Melting and boiling points also relate directly to this bonding. Gold, silver, copper and a few other native metallic minerals have metallic bonding.

The individual type(s) of interatomic linkages or chemical bonds in minerals, metallic, ionic or polar, covalent and van der Waals, imparts characteristic properties the matter in which it occurs. More than one type of chemical bond may be present in a single compound: 1. mixed and 2. like bonds. Several physical properties are the result of the mixed type such as hardness. melting point. and strength. Each physical characteristic is determined by the weakest bonds. These are direct1y affected by increased mechanical or thermal stress. Still, the classification of bonding is arbitrary and many compounds may have intermediate types. For example, silicate minerals are neither purely ionic nor covalent but intermediate. As earlier noted. the geometry assumed by any solid is such that the entire crystal lattice tends to form with minimum potential energy.

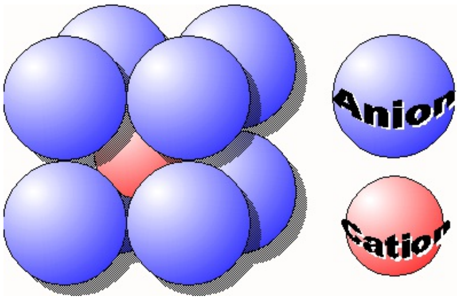

Figure 1.15 - 8-fold (cubic) coordination of anions around a cation

1.4.1 COORDINATION PRINCIPLE and RADIUS RATIO

The ions of the different elements vary in size of their ionic radii (Table 1.2). The positively charged cations are usually much smaller than the negatively charged anions. Given opposite charged ions which unite to form a crystal structure, an ion tends to gather to itself as many ions of the opposite sign as the ion size permits. The gathered anions cluster about the central coordinating cation so that their centers are at the apices, or pointed ends, of a regular polyhedron. The coordination number (CN) is the number of anions in the polyhedron. For the mineral fluorite (CaF_2) each calcium (Ca^{2+}) Ion is at the center of eight fluorine ions and the CN is 8 in respect to fluorine (see figure 1.15). Yet, in the same mineral each fluorine ion has only four close calcium neighbors and is in 4-fold coordination (CN-4) in respect to calcium. The ratio between the cation and anion ionic radius is called the maximum radius stability ratio and can be used to estimate the CN, which in turn can be used to estimate the resulting unit cell geometry for each crystal class of the formed compound (see table 1.3).

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Strongest forces exist between ions which are nearest to each other. The geometric form and coordination number depend on the relative sizes of the coordinated ions. The relative sizes are expressed as radius ratios. As an example for CaF_2, the Ionic radius for Ca is 0.99 Å and for F (Fluorine) 1.33 Å, hence $IR_{Ca}/IR_F = 0.99/1.33 = 0.74$. Ratio determinations may be calculated for other cation anion pairs or combinations. When these are compared we find that there are limiting coordination conditions for various ratios as shown in Table 1.3.

Table 1.3 - Coordination Numbers (CN) and radius ratios

Coordination Crystal Packing **(unit cell geometry)**	Radius Ratio = IR_{cation} / IR_{anion}	CN	Example
Hexagonal or Cubic	> 1.000	12 (closest packing)	Cesium & Potassium Ions
Cubic	0.732 - 1.000	8	Fluorite
Octahedral	0.414 - 0.732	6	Na, Cl in Halite Ca, CO_3 in Calcite
Tetrahedral	0.255 - 0.414	4	SiO_4 groups in silicates ZnS (sphalerite) structure
Triangular / Hexagonal	0.155 - 0.255	3	Carbonate (CO_3), NO_3 and BO_3 ionic groups
Linear	<0.155	2-fold	Rare in ionic bonded crystals: Uranyl and cuprite (Cu_2O) groups

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You will note that the oxygen anions are so large compared to most cations, that the structure of a mineral is mainly the packing of oxygen ions with cations in the interstices. Because the radius ratio of some cations lies near the theoretical border between two types of coordination, they may occur in both. Aluminum is a case in point. Finally, electrical neutrality and geometrical stability are prime factors in determining the structure of ionic compounds.

The table above predicts the unit cell geometry denoted in the left hand column. The actual crystal structure of covalent compounds using CN and radius ratios is more difficult to predict, since combinations generate building blocks (unit cells) first and these unit cells combine into crystalline structures. Quartz is case in point. With an ionic radius for Si - cation of 0.40Å and the O - anion of 1.40Å, radius ratio is equal to 0.286, which puts the quartz unit cell into the tetrahedral group. Indeed, the building blocks of silicates like quartz are silica tetrahedra, which have the predicted shape. Individual tetrahedra are now combining to form quartz, resulting in a hexagonal shaped final crystal.

1.4.2 IONIC SUBSTITUTION

Substitution of an ion or ionic group of a different element may occur in a given structure. This is called ionic substitution and the extent to which it occurs depends on several factors. Most important is ionic size. For example, ions of two elements can substitute for each other if the ionic radii are smaller or differ by less than 15%. If more than this, substitution is rare. Temperature of mineral formation may also affect substitution within the above limits. The higher the temperature, the greater the thermal activity and the less rigorous the space requirements within the lattice structure of the mineral. Crystals formed at high temperatures may show extensive ionic substitution, which is not possible at lower temperatures. A similar valence or electrical charge favors substitution, but is not a requirement as long as electrical neutrality can be maintained by other ions.

Complete ionic substitution may take place in mineral groups of exactly equal structure (Isostructural). For example siderite, $FeCO_3$, can have Mg^{2+} enter the lattice structure of the mineral in any proportion, or Fe^{2+} may enter the lattice of magnesite, $MgCO_3$, in all proportions. The valence charge is the same. The most abundant mineral group in the world, the plagioclase feldspars, have a type of solid solution or substitution of Na and Ca from the calcite rich anorthite to the sodium rich albite, i.e., $NaAlSi_3O_8 \rightarrow CaAl_2Si_2O_8$.

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1.5 MINERAL GROUPS

According to chemistry, minerals can be classified or subdivided into various mineral groups. The silicate minerals comprise the largest of the groups and are discussed under 1.6 below. Figure 1.16 summarizes the most common mineral groups according to major chemical composition.

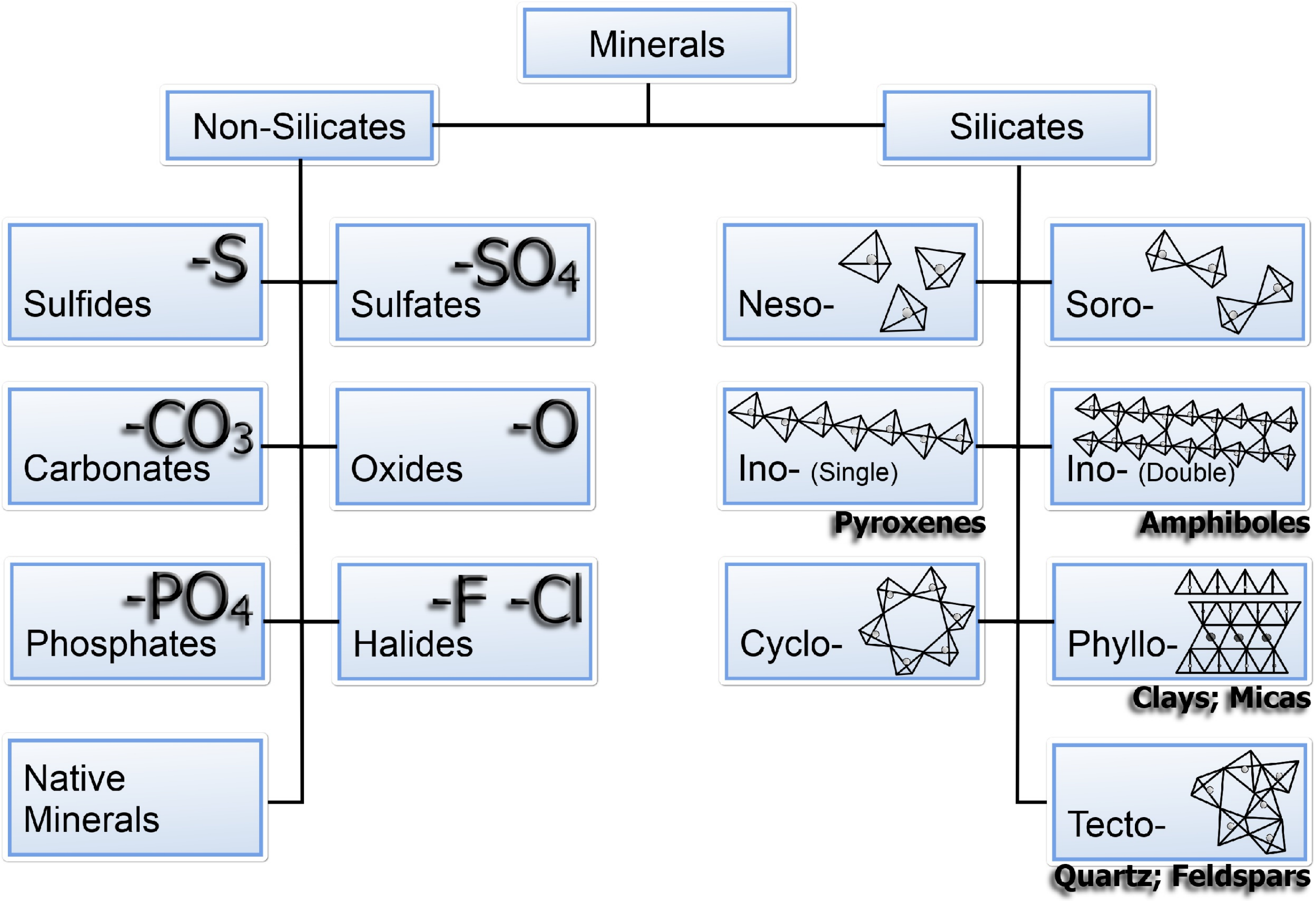

Figure 1.16 - Common mineral groups divided into silicates and nonsilicates according to chemical composition. Line with elemental symbol denotes major anion in group.

1.5.1 SULFIDE GROUP

Sulfide minerals contain anionic sulfur, but NO oxygen (do NOT confuse with sulfates). Minerals in this group are predominantly metallic and contain many economically important metal ores such as galena (PbS; Lead ore), chalcopyrite ($CuFeS_2$; Copper ore), molybdenite (MoS_2; Molybdenum ore) and sphalerite (ZnS; Zinc ore). Multiple metallic elements may combine with sulfur to create exotic ore minerals like tetrahedrite-tennanite ($((Cu,Fe)_{12}(SbAs)_4S_{13}$ with Ag, Pb, Hg, and Zn as ionic substitutions, such making it a multiple metal ore mineral. The most common sulfide however is the economically worthless yet beautiful pyrite or fools' gold (FeS_2). The affinity of sulfur to combine with a second metallic or semi-metallic anion expands the sulfide group into arsenides (Arsenopyrite FeAsS) and antimonides (Stephanite Ag_5SbS_4).

1.5.2 SULFATE GROUP

Sulfates contain the negatively charged sulfate ion SO_4, and form most commonly from an aqueous solution, such as evaporite deposits of a arid lacustrine sedimentary environment or within low temperature hydrothermal or hot-spring deposits. Chemical weathering of sulfide ores may also produce sulfate mineralization. The most common sulfate minerals are gypsum ($CaSO_4*2H_2O$), anhydrite ($CaSO_4$) and barite ($BaSO_4$). Economically valuable sulfate ore minerals include chalcanthite ($CuSO_4*5H_2O$; Copper ore) and anglesite ($PbSO_4$; Lead ore).

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1.5.3 CARBONATE GROUP

Carbonate minerals are wide spread and contain the CO_3 or carbonate anion. Most carbonate minerals will effervesce vigorously when exposed to acid, thus making this group easy to identify. Minerals in this group often form in an marine environment or in the interaction of saline and fresh waters. The most common carbonate minerals are dolomite ($CaMg(CO_3)_2$), as well as calcite with hexagonal rhombohedral crystals and its orthorhombic counterpart aragonite both with the identical chemical formula $CaCO_3$. Copper ore carbonates are represented by green malachite ($Cu_2CO_3(OH)_2$) and the deep blue azurite ($Cu_3(CO_3)_2(OH)_2$), often a chemical weathering byproduct in the oxidizing zone of copper ore deposits.

1.5.4 OXIDE GROUP

Oxides contain oxygen (well, duh!) but NO silicon. Do NOT confuse with silicate minerals, all of which contain oxygen but also anionic silicon. Many minerals in this group are extraordinarily hard (corundum Al_2O_3 - Mohs hardness: 9) and make excellent gemstones. For example the highly priced blue sapphires and red rubies are both corundum minerals. The oxide group also contains many important ore minerals such as magnetite (Fe_3O_4; Iron ore), ilmenite ($FeTiO_3$; Iron-Titanium ore), chromite ($Fe_2Cr_2O_4$; Iron-Chromium ore) and cuprite (Cu_2O; Copper ore). This group also contains the hydrated forms of within the oxidized zone of chemical weathering or hydroxides, namely bauxite ($Al_2O_3*2H_2O$; Aluminum ore), Limonite ($2FeO*3H_2O$) or Goethite ($FeO(OH)$).

1.5.5 PHOSPHATE GROUP

The phosphate mineral group comprises minerals which not only contain the obvious PO_4 phosphate ion, but also AsO_4, VO_4, WO_4 or any mineral with an AO_4 polyatomic ion format such as turquoise ($CuAl_6(PO_4)_4(OH)_8*5H_2O$; Gemstone), carnotite ($K_2(UO_2)_2(VO_4)_2*3H_2O$; Uranium-Vanadium ore) or Scheelite ($CaWO_4$; Tungsten ore). The most common phosphate mineral is probably apatite ($Ca_5(F_3Cl)(PO_4)_3$) which is often found in biological environments such as teeth and bones.

1.5.6 HALIDE GROUP

Halides are the anionic species of either fluorine, chlorine, iodine, or bromine ion. Minerals in this group are natural salts and many of them are water soluble such as halite (NaCl; table salt) or sylvite (KCl; fertilizer and deicer) and form under similar circumstances as the sulfate minerals. Hence sulfates and halides are often found together. Hydrothermal halide group minerals include fluorite (CaF_2; Fluorine ore) and even precious metal ores such as the rare cerargyrite (AgCl; Silver ore).

1.5.7 NATIVE ELEMENT GROUP

Minerals in this group are chemical elements in their native state without being bonded to other elements. Many of the most valuable minerals occur as native elements such as gold (Au), platinum (Pt) and diamond (C). Other common occurrences are copper (Cu), graphite (C), and sulfur (S).

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1.6 SILICATE MINERALS

In the silicates, the most abundant of all mineral types, silicon lies between four oxygen atoms. This structural block is universal in the silicates and is called the silica tetrahedron. The four oxygens always occur at the corners of a tetrahedron of almost constant size and regular shape (figure 1.17).

The different types of silicates result from the various ways in which the silicon-oxygen tetrahedra are linked to each other. These various linkages, or lack of, form several silicate structural types, namely *nesosilicates, sorosilicates, inosilicates, cyclosilicates, phyllosilicates*, and *tectosilicates*. These individual silicate groups and their respective minerals are discussed below.

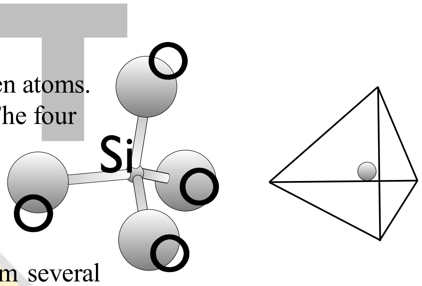

Figure 1.17- The silicate tetrahedron represented in "ball and stick" and "stylized tetrahedra format"

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1.6.1 NESOSILICATES

Independent tetrahedral groups, or **nesosilicates**, have the composition SiO_4. Here the individual Si-tetrahedra do not touch each other, but are connected through another, usually metallic, ion. The mineral olivine, variety forsterite [Mg_2SiO_4], is typical of this group. Other common nesosilicates are the Al_2SiO_5 group of andalusite, sillimanite and kyanite, as well as staurolite, topaz, the garnet group, zircon and sphene. Figure 1.18 shows individual silica tetrahedra linked by other large atoms or ions.

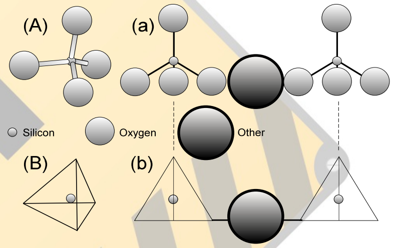

Figure 1.18 - Nesosilicates

1.6.2 SOROSILICATES

The **sorosilicates** or disilicates are formed by two tetrahedra sharing a common oxygen. The composition is Si_2O_7 for this subclass. Common minerals include vesuvianite (*former* idocrase) [$Ca_{10}Mg_2Al_4(SiO_4)_5(Si_2O_7)_2(OH)_4$], epidote [$Ca_2(Fe^{3+},Al)_3O(SiO_4)(Si_2O_7)OH$], and hemimorphite[$Zn_4Si_2O_7(OH)_2 \bullet H_2O$]. The drawing (figure 1.19) displays two linked tetrahedra separated by large other atoms or ions.

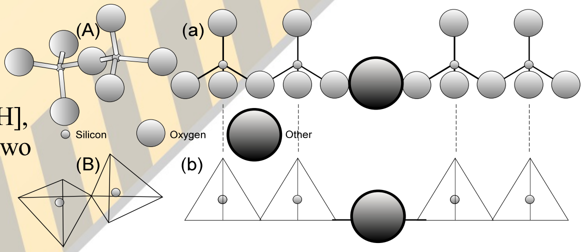

Figure 1.19 - Sorosilicates

1.6.3 INOSILICATES

You will recognize the possibility of the development of chain-like structures of two types and of indefinite extent: a) single chains, and b) double chains. These are called **inosilicates**. In the single chain each silica tetrahedra shares two oxygens with adjacent Si-tetrahedra and has an Si:O ratio = 1:3 (figure 1.20 (C)). The pyroxene mineral group with its common mineral member augite consists of such single silica chains. If alternate tetrahedra in two parallel single chains are linked, the Si:O ratio is 4:11. Here, every second tetrahedron in the chain shares three oxygens with adjoining neighbors (figure 1.20 (D)). This structure is characteristic of the well-known, and quite abundant amphibole group, such as hornblende. (figure 1.20).

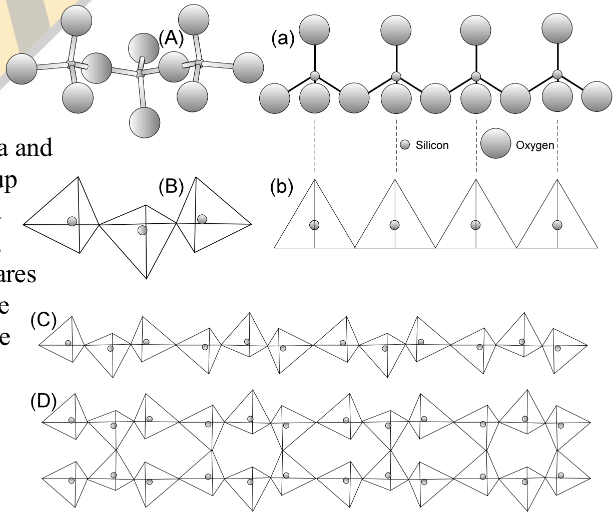

Figure 1.20 - Inosilicates

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1.6.4 CYCLOSILICATES

When silica tetrahedra form closed rings, **cyclosilicates** are formed. These rings may consist of either 3, 4, 6 or 9 tetrahedra, have a 1:3 Si:O ratio and the general composition of Si_nO_{3n}. Ring structures are connected vertically and horizontally within the crystal lattice by other elements, and rarely are linked directly one to another. However, double ring silicate minerals exist, such as milarite $[(K,Na)Ca_2AlBe_2Si_{12}O_{30}\bullet H_2O]$ or osumilite $[(K,Na)(Fe^{2+},Mg)_2(Al,Fe^{3+})_3(Si,Al)_{12}O_{30}\bullet nH_2O]$, where two 6 member rings are stacked on top of each other. Cyclosilicates composed of individual six-linked tetrahedra rings include beryl $[BeAl_2Si_6O_{18}]$, dioptase $[Cu_6Si_6O_{18}\bullet 6H_2O]$, and the tourmaline group. Other minerals in this group are benitoite $[BaTiSi_3O_9]$ composed of three-linked tetrahedral rings, and the blue-colored, four-member ring mineral papagoite $[CaCuAlSi_2O_6(OH)_3]$. Figure 1.21D shows various ring structure permutations.

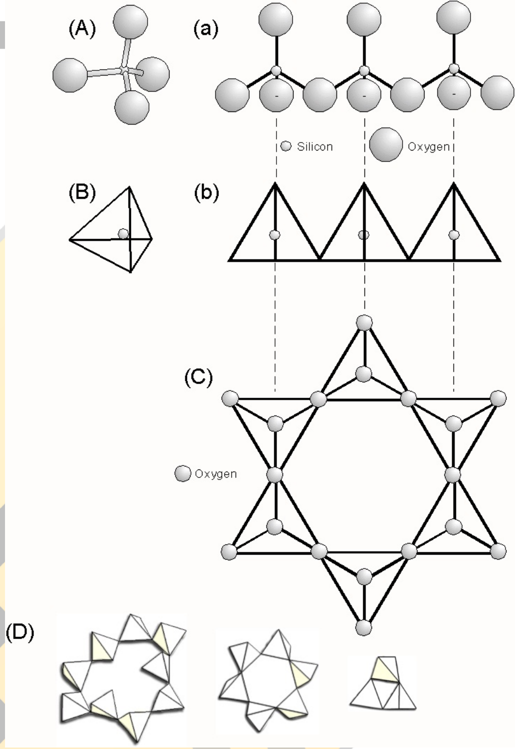

Figure 1.21 - Cyclosilicates

Most cyclosilicates fall within the hexagonal crystal class with the notable exception of orthorhombic cordierite $[Mg_2Al_4Si_5O_{18}]$. Many minerals in this group are rather hard ($H > 7$) and make good candidates for gemstones, such as beryl (emerald, aquamarine), cordierite (iolite), and the tourmaline group, which includes the recently discovered, exquisitely rare, neon-blue Paraiba tourmaline, able to command 5 digit prices per carat (200mg).

<u>Side Note:</u> The Paraiba tourmaline is actually a copper bearing variety of the tourmaline species elbaite. These cuprian blue tourmalines have been found in the Pairaba region of Eastern Brazil, as well as in Nigeria (Western Africa), thus giving possible additional external evidence for the theory of continental drift and plate tectonics.

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1.6.5 PHYLLOSILICATES

A sheet structure, known as the **phyllosilicates**, develops when three oxygen ions of each tetrahedra are shared with adjacent tetrahedra. This is basically the double-chain inosilicate structure extended indefinitely in two directions forming a sheetlike structure. In more detail these sheets tend to have a hexagonal planar framework which results in a false hexagonal habit. Perfect basal cleavage occurs parallel to the plane of the sheet, where the sheets are interconnected by weak van der Waals bonds. Such linkage yields a ratio of Si:O of 2:5. All micas and clay structures are built on this basic unit, as well as talc. Figure 1.22 depicts the development of such a sheet from individual Si-tetrahedra. See also table 1.4a.

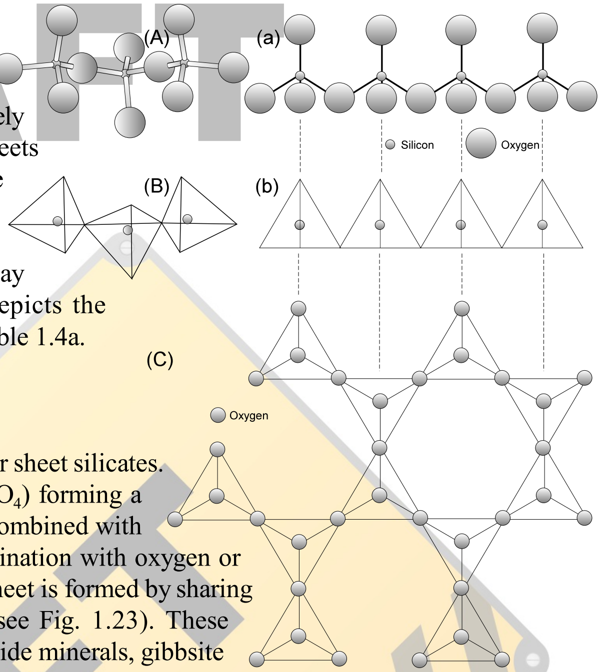

Figure 1.22 - Phyllosilicates

General Mineralogy of Clays

Clays are silicate minerals and belong to the family of phyllosilicates or sheet silicates. Their basic structural building blocks are linked silica tetrahedra (SiO_4) forming a sheet by sharing oxygen atoms (see Fig. 1.22). This layer or sheet is combined with another layer grouping of cations, usually Al, Mg, or Fe in six-coordination with oxygen or hydroxyl anions, forming an octahedral pattern around the cation. A sheet is formed by sharing of the oxygen or the hydroxyl anions with neighboring octahedra (see Fig. 1.23). These octahedral sheets are usually named after two common, layered hydroxide minerals, gibbsite ($Al(OH)_3$) and brucite ($Mg(OH)_2$), forming dioctahedral and trioctahedral layers, respectively (Lambe and Whitman, 1979). The silica tetrahedra, gibbsite, and brucite layers are stacked on top of each other and combined by sharing of oxygen and hydroxyl ions. A two-layer mineral or 1:1 layer results when only one surface of an octahedral sheet is shared with a tetrahedral layer (e.g. kaolinite). In this case, the unshared octahedral surface consists of hydroxyl ions (OH). A 2:1 or three-layered sheet is formed by sandwiching an octahedral sheet between two tetrahedral layers, sharing both octahedral surfaces with adjacent tetrahedral layers (see table1.4a).

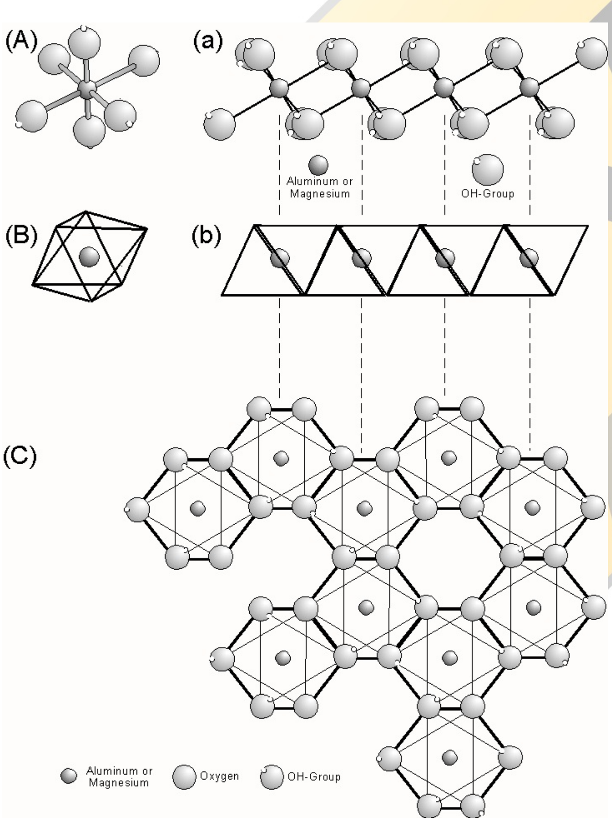

Figure 1.23 - Schematics of octahedra and associated sheet structure.3D representation of single octahedra: (A) ball-and-stick model, (B) diagrammatic with center cation. 2D linked sheet structure: (a) ball-and-stick, (b) diagrammatic. (C) Projected ab-plane diagrammatic view of sheet.

Those individual 1:1 or 2:1 layers may combine to form a specific mineral. If the tetrahedral cations consist exclusively of Si and all octahedral cations are either Al (dioctahedral) or Mg (trioctahedral) without any substitution, then the resulting layers are electrostatically neutral. Individual 2:1 sheets are held together by van-der-Waals bonds (e.g. talc, pyrophyllite). However, isomorphous substitution of lower charged for higher charge cations results in an overall negative layer charge. These unequal charge is balanced by positively charged interlayer material, combining the separate 1:1 or 2:1 layers. The material between the layers can be individual cations (micas), hydrated cations (expanded clays), hydroxide groups (chlorite-like clays) and whole additional octahedral hydroxide sheets (chlorites) (Weaver, 1989). Thus, chlorite consists of an additional sheet as interlayer material and is therefore referred to as 2:2 or 2:1:1 layer unit.

Stacking of composite octahedral-tetrahedral layers occurs always in the crystallographic c-axis direction. Most phyllosilicates exhibit a monoclinic or triclinic crystal structure and display a pseudohexagonal nature in the a-b plane of the crystal. All sheet-silicates show perfect basal cleavage between the individual layers (Berry and Mason, 1959). The various phyllosilicate types can be generally classified according to their structural unit (e.g. 1:1, 2:1, 2:2), their dioctahedral (gibbsite) and trioctahedral (brucite)

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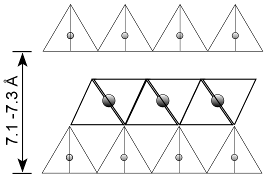
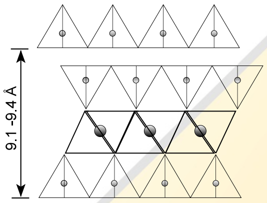
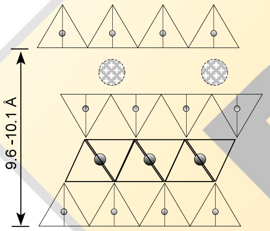
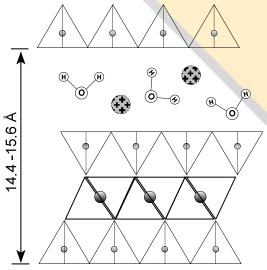
characteristics, and their unit spacing (see table 1.4a). Five major groupings of sheet silicate minerals are thus established: (1) the kaolinite-serpentine group, (2) the talc-pyrophyllite group, (3) the micas and brittle micas, (4) the smectite-vermiculite group, and (5) the chlorites (Bailey, 1980). Chamley (1989) and Weaver (1989) identify a sixth type, the 2:1 layer palygorskite-sepiolite group of fibrous clays and inverted ribbons. They do not represent sheet silicates in a strict sense but rather exhibit a needle-like morphology. Their growth along the c axis is limited, however, the tetrahedral sheets extend for considerable distances in the a and b directions. These sheets invert at periodic intervals along the b axis, forming a checkerboard pattern with water molecules filling the empty spaces (Weaver, 1989).

Table 1.4a - Classification of major phyllosilicate groups after WEAVER (1989), LAGALY & KÖSTER (1993), CHAMLEY (1989), and BERRY & MASON (1959).

Layer type	Dioctahedral (Gibbsite-Type Layers)	Trioctahedral (Brucite-Type Layers)
1:1 Layer 	Kaolintes Basic Chemistry $Al_4Si_4O_{10}(OH)_8$ Examples Kaolinite, Nacrite, Dickite	Serpentines Basic Chemistry $Mg_6Si_4O_{10}(OH)_8$ Examples Chrysolite, Antigorite
2:1 Layers 	Pyrophyllites (fire-clays) Basic Chemistry $Al_2Si_4O_{10}(OH)_2$ Examples Pyrophyllite	Talcs Basic Chemistry $Mg_3Si_4O_{10}(OH)_2$ Examples Talc
	① Micas ② Brittle Micas Basic Chemistry ① $KAl_2(AlSi_3O_{10})(OH)_2$ ② $CaAl_2(AlSi_3O_{10})(OH)_2$ Examples ① Illite, Muscovite, Glauconite ② Margarite	① Micas ② Brittle Micas Basic Chemistry ① $KMg_3(AlSi_3O_{10})(OH)_2$ ② $CaMg_3(AlSi_3O_{10})(OH)_2$ Examples ① Phlogopite, Biotite (Fe for Mg) ② Anandite, Clintonite
	① Smectites (② Vermiculites) Basic Chemistry $R_xAlSi_4O_{10}(OH)_2 * nH_2O$ Examples ① Montmorillonite, Nontronite (② Vermiculite (di))	① Smectites ② Vermiculites Basic Chemistry $R_xMgSi_4O_{10}(OH)_2 * nH_2O$ Examples ① Saponite, Hectorite, Stevensite ② Vermiculite (tri)

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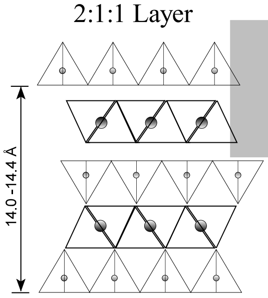
2:1:1 Layer



Chlorites

Basic Chemistry
$Al_2(AlSi_3O_{10})(OH)_8$

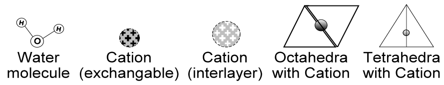
Examples
Donbassite (di)
Cookeite, Sudoite (di,tri)

Chlorites

Basic Chemistry
$Mg_3Al(AlSi_3O_{10})(OH)_8$

Examples
Chlorite, Clinochlore (tri)

Legend:



Basic Chemsitry

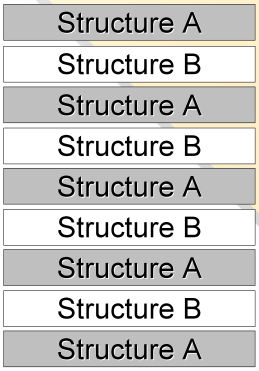
Substitutions in formula possible
e.g. Fe

R = exchangeable cation

Another major group of phyllosilicates is known as mixed layered minerals. Here, a single clay crystal is made up of a composite of different basic structures in the layer planes (Velde, 1995). Smectite, for example, might have some of the layers exchanged by a layer of mica, resulting in a new mineral with characteristic properties. Major classifications of mixed layer minerals are shown in Table 1.4b. Phyllosilicates with regular mixed layering are relatively limited and usually associated with high temperature environments, such as hydrothermal alterations. Their mineral layers exhibit a regular, repeated pattern with equal proportions of the two component structures stacked in sequence, as illustrated in Table 1.4b. Most common regularly stacked clays are alternating layers of illite/smectite and chlorite/smectite. Randomly layered minerals can be described according to type and proportion of the two or more types of layers (see Tab. 1.4b). These phyllosilicates are by far the most common and are probably the second most abundant clay minerals next to illite. The most frequently encountered representatives of this group consist of randomly alternating illite/smectite stacks and are generally referred to as I/S physils (Weaver, 1989). However, certain regularities may exist in the layer sequences of I/S physils, leading to a classification by Reichwite numbers, where R0 = random I/S layering, R1 = I:S, R2 = I:I:S, and R3 = I:S:I:I (Jagodzinski, 1949).

Table 1.4b - Classification of mixed layered phyllosilicates with examples of most common minerals (modified after VELDE (1995) and REYNOLDS (1980))

Regular Mixed Layering



Mica : Smectite

Dioctahedral
illite/smectite
rectorite (Na mica)
allevardite (K mica)

Trioctahedral
biotite/smectite
hydrobiotite

Chlorite : Smectite

Dioctahedral
sudoite

Trioctahedral
correnite

Dioctahedral - Trioctahedral
tosudite

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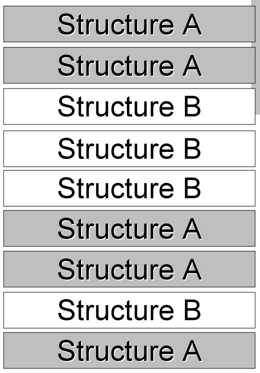
Random Mixed Layering



Mica : Smectite

illite/smectite
biotite/smectite
celadonite/smectite
glauconite/smectite

Chlorite : Smectite

Kaolinite : Smectite

Mica : Chlorite

1.6.6 TECTOSILICATES

The **tectosilicates** are a three dimensional or polymerized network with a SiO_4 tetrahedron sharing all its corners with other tetrahedra (figure 1.24). The Si:O ratio is 1 :2. Quartz and other silica forms such as tridymite and cristobalite have this structure. Because Al^{3+} may substitute for Si^{4+}, additional positive ions are required to maintain electrical neutrality. The very abundant feldspars, nepheline, and less abundant zeolites, are examples.

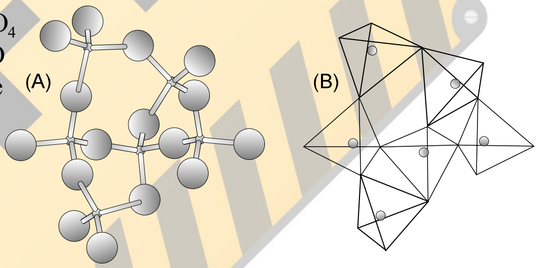


Figure 1.24 - Tectosilicate structures: quartz and sodic feldspar

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1.7 BOWENS REACTION SERIES

Because of the extreme importance of the SiO_4 tetrahedral linkages. or its type of polymerization during magma crystallization, a reaction series diagram may help to clarify the relationships of chemistry and geometry in the development of the very important silicate minerals from igneous melts. Norman Levi Bowen (1887 - 1956) and his co-workers reported on productive research results on the characteristics of magmatic crystallization. They were able to show that the common igneous rock minerals, the most abundant in the mineral kingdom, can be arranged into two series:

1: a discontinuous reaction series made up of the ferromagnesian minerals, or minerals containing iron and magnesium
2: a continuous reaction series of the feldspars.

Figure 1.25 illustrates this reaction series. For the discontinuous series the crystallization begins at the highest temperature with olivine, one of the isolated SiO_4 tetrahedral groups, or nesosilicates. The temperature decreases and if still molten magma remains, the olivine completely reacts with the melt and gradually changes to pyroxene, the single chain tetrahedra of SiO_4. At the same time calcium rich feldspar continues to react with the melt and sodium becomes part of the feldspar composition. If some of the olivine and bytownite are removed from the crystallizing system, e.g. by settling of the olivine in the magma, or the floating of the less dense feldspar phenocrysts to the top of the magma chamber, then the reaction series continues to take place in the molten magma. Amphiboles and labradorite feldspar may crystallize. If complete crystallization takes place at this stage, a rock with the composition of amphibole-labradorite forms. The latter is an andesite or diorite depending on cooling rate and texture.

The longer the magma remains fluid, the more the reaction series proceeds. Granites and pegmatites, zeolitic rich rocks, and finally hydrothermal minerals may form, including the metallic ore minerals. Each of these is derived from the still molten mass or the residual fluids of the original magma. Pegmatites form as a result of residual high water content and other mineralizer fluids in the late stage magma. These permit lower temperatures of crystallization, high mobility of the various ions or elements, and endow a greater fluidity to the end-stage magma to form rather large crystals. Sizes of several feet are not uncommon.

From olivine to quartz, all structural forms develop and may disappear. You will note the geometric progression from the isolated tetrahedra of SiO_4 first, then the single chains, double chains, sheets, and finally the three-dimensional networks or tectosilicates. This reaction series is a marvel of orderly development and geometric stability for the conditions extant at the time of the crystallization of the magma. Herein lies the secrets of not only structure of silicates minerals, but composition as well.

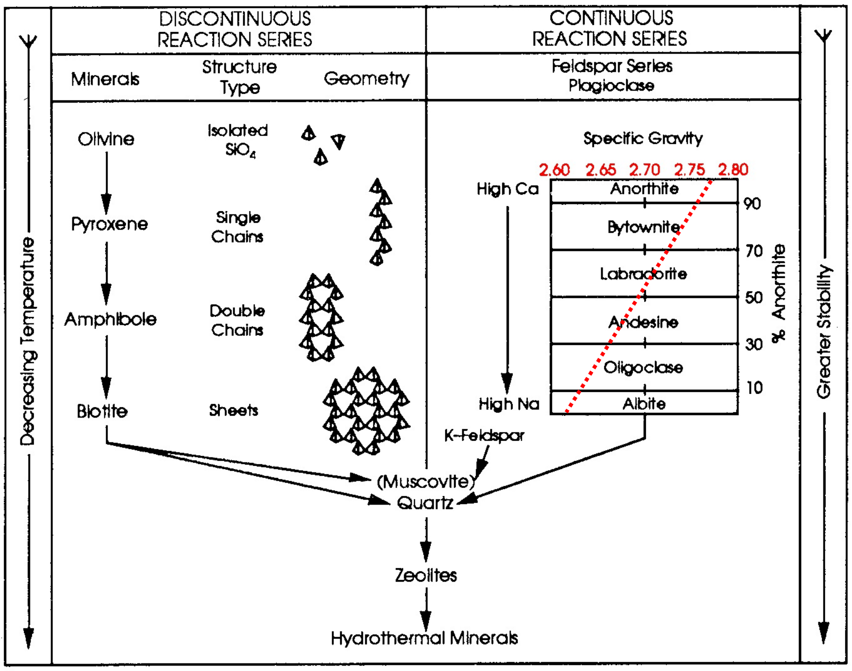

Figure 1.25 - Bowen's reaction series (modified)

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1.8 SOLID SOLUTION PHASE DIAGRAMS AND MINERAL IDENTIFICATION

As indicated in Bowen's reaction series, the plagioclase group forms a solid solution series as cooling progresses. Plagioclase mineral species are continuously added to the growing crystals like growth rings found in trees. It is not uncommon that the center of the plagioclase is a different mineral species than the outside rim.

Solid solution crystals form in a specific manner. As the melt cools, a phase of "crystal mush" is encountered, which is liquid dispersed with already formed crystals. Progress in cooling increases the size of the crystals while the melt is continually depleted until all magma has solidified. Fields of all liquid melt, crystal mush, and all solid crystal is depicted in the plagioclase solid solution diagram (Fig. 1.26). Different plagioclase species have different melting and solidification points and various sized crystal mush fields depending on the CaO - Na_2O ratios present in the melt and as shown on the x-axis.

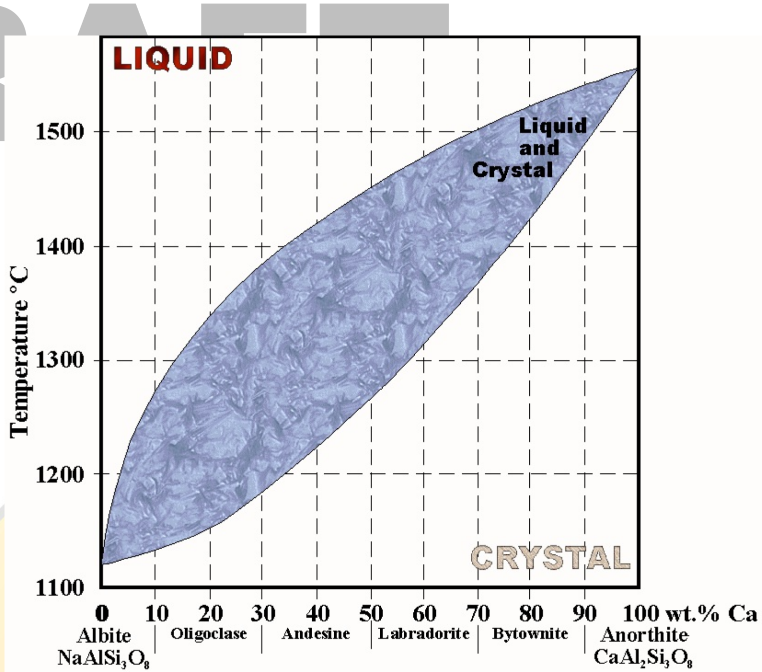

Figure 1.26 - Plagioclase Solid Solution Phase Diagram

The point where crystals start forming, but are being in a mix with the melt is represented by the upper curve of the "liquid and crystal" or "crystal mush" field in figure 1.26. This curve is called **liquidus**. The opposite or lower curve between the "liquid and crystal" field and the "solid" crystal field depicts the transformation of the mush into all solid, called the **solidus** curve.

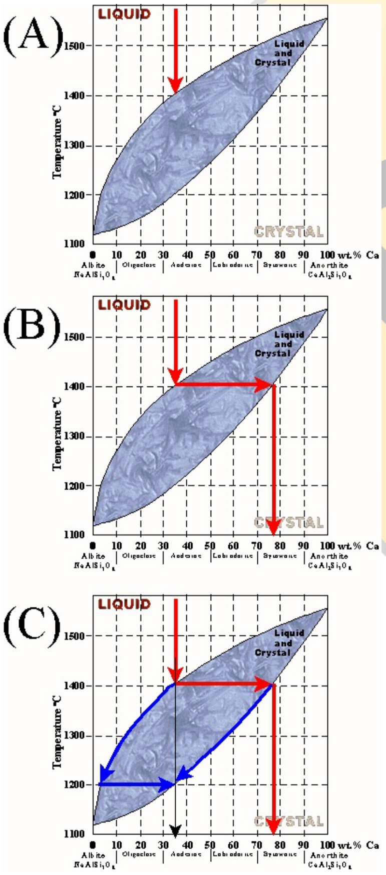

Figure 1.27 - Plagioclase Solid Solution Phase Diagram Reading

How to read such phase diagrams? Using the example given in figure 1.27 we start with a CaO - Na_2O percentage of 35% CaO (inferred 65%Na_2O) in the melt. A crystal with this Ca/Na ratio would be andesine as seen on the x-axis. At this point however, no crystals have formed. As the melt is cooled, the liquidus curve is encountered at ~1410°C (see fig. 1.27A) and the first small crystals will start to form at this point.

The composition of these first crystals forming at 1410°C needs to be read in the solid crystal field, because these are indeed solid crystals and no longer melt. Hence one must move horizontally from the liquidus 1410°C point to the solidus 1410°C point. Dropping down vertically as shown in figure 1.27B, the composition of these first crystals is indicated as approximately 77%CaO, which is the plagioclase mineral bytownite.

As cooling continues, see black arrow in figure 1.27C, the melt and crystal chemistry changes continuously but synchronized along the solidus and liquidus curves (blue arrows). The crystals that form will have less and less calcium, but increase in sodium. The melt also decreases in Ca content, keeping lock-step with the chemistry of crystals being formed indicated by the blue arrows. Approaching roughly 1200°C, the black arrow in the figure crosses the solidus line. At this point ALL the melt has solidified.

However, we can use the graph to determine the composition of the final crystal and the chemistry of the last drop of melt. At the 1200°C mark, the liquidus line has dropped to roughly 2%CaO which is the composition of the last melt. At the same time and temperature, the solidus line is now at 35%CaO which equates to the chemistry of the last crystal that forms, an andesine plagioclase. Thus the growth of the plagioclase minerals has progressed from bytownite through labradorite to finally andesine during the cooling process. Growth rings within the formed plagioclase should attest to this fact.

2 PHYSICAL PROPERTIES OF MINERALS

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Minerals can be identified by using diagnostic physical properties. Such properties should be easy to determine and are therefore limited to the following: color, cleavage, hardness, luster, streak, specific gravity and tenacity. Tenacity includes sectility, malleability, brittleness, flexibility, and elasticity. Elasticity is a property common to all minerals but is optically visible in only a very few. A single physical property may not permit specific identification, but several such properties in combination will.

2.1 COLOR

Color: The coloration of minerals depends in general upon the selective light absorption which it exerts upon the light reflected by it. If a mineral is green, for example, it reflects all the light waves except those which combined yield the sensation of green. The causes for color and color variation even within the same mineral is highly complex. Small impurities of certain chemical elements within the crystal structure yield coloration. For example, the red variety of the mineral corundum (Al_2O_3), namely rubies are colored by small amounts of chromium (<1%) in their crystal structure, while the same impurity in a beryl ($Be_3Al_2Si_6O_{18}$) will produce the deep green gemstone variety emerald. On the other hand the blue coloration in sapphires, also a gem corundum variety, results from a cation to cation charge transfer between Fe^{+2} and Ti^{+4} within the crystal lattice. Light photons dislocate atomic electrons and transfer them to another cation or anion, causing coloration in the reflected wavelengths. Another cause for color in minerals are so called color centers or defects in the atomic crystal lattice which trap electrons, causing coloration in the reflected light. A good example for this type of coloration is the blue variety of barite ($BaSO_4$). Since color centers are low energy crystal structure flaws, they can be repaired by adding small amounts of additional energy such as gentle heating, radiation treatments or simply exposing the sample to sunlight. Thus a fine specimen of sky-blue barite will turn completely clear after exposure for a few days in the full sun. Artificially induced energy can be used in creating more valuable gems from baser varieties if coloration is due to color centers. For example, undesired brown zircon ($ZrSiO_4$) can be turned into a desired blue gemstone (starlite) by heating it to a few 100°C. Despite the many causes of coloration in minerals, color is an important and easily observed diagnostic property of each mineral. In studying the identification charts you will recognize that some individual minerals appear on several color charts. Such color variations may be present in the same outcrop, or in rare occasions in the same mineral specimen.

Mineral color can basically be grouped into three categories, which are Idiochromatic, Allochromatic and Pseudochromatic.

Idiochromatic Minerals show a fixed color that is due to the specific chemical composition of the mineral. The color in these minerals is predictable and does not change. Good examples for idiochromatic minerals are sulfur (always yellow), malachite (always green) and azurite (always blue).

Allochromatic Minerals on the other hand show color variations which are due to trace impurities or crystal structure defects. Common coloring agents are impurities consisting mainly of the chemical elements chromium (red - green), cobalt (blue), copper (green - blue), iron (red - green - yellow), manganese (pink), nickel (green), titanium (blue), and vanadium (green - color change). Examples of allochromatic minerals are corundum (red = ruby, blue = sapphire) or beryl (green = emerald, blue = aquamarine)

Pseudochromatic Minerals attribute their color changes to physical optical properties that affect light reflection or refraction. In transparent or translucent minerals, light scattering can produce the multicolored properties of an opal, or asterism in star corundum, garnet and quartz. Interferences through thin mineral surface tarnishes are responsible for the peacock feather colors in bornite. Other such rainbow color effects can be attributed to diffraction like surface grating as in labradorite or iris agate for example.

The observation of color in minerals is straight forward. It is a great tool for minerals of idiochromatic or constant color, thus meeting the "Common Color" description in the determinative table at all times.

2.2 CLEAVAGE & FRACTURE

Cleavage: Cleavage is the second major primary Identifier on the mineral identification charts. This physical property is the tendency of a mineral to break or cleave along crystallographic planes. Thus, cleavage is a more or less smooth surface of breaking in the mineral. The cleavage planes are reflections of possible crystal faces and, in turn, planes of the internal atomic structure of the mineral. These cleavage planes show minimum values of cohesion parallel to the cleavage surface.

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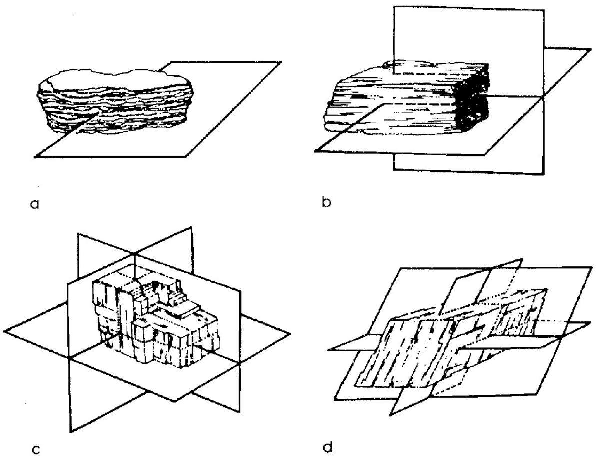

Figure 2.1 - Examples of mineral cleavage: a) one-direction. b) two direction. c) 3- direction at right angles. d) 3-direction not at right angles.

Cleavage is readily recognized by the mirror-like reflection of light from the mineral surface when it is rotated in strong light. Keep in mind that cleavage is a broken surface of planar form in contrast to a primary crystal face. The mineral identification charts consider only the presence of distinct or indistinct cleavage in the mineral. Remember that several cleavage types are possible in various minerals. Some show one-direction cleavage, others two-direction at right angles, two direction not at right angles, and even three direction cubic, three direction rhombohedral, and others. Examples of some of the more common cleavages are shown in figure 2.1.

Each individual mineral description includes a comment on the character of, or lack of cleavage, for that specific mineral and the kind of cleavage to expect. For your practical use, however, the presence or absence of cleavage is generally sufficient for most mineral identifications.

2.3 LUSTER

Luster: In the identification tables METALLIC and NON-METALLIC properties of the mineral are used as a major identifier. The metallic luster is a reflectance characteristic of metals such as copper, iron, tin, and others. The non-metallic luster includes several varieties such as adamantine, or that of the highly refractive index such as for diamond; vitreous luster, or that of broken glass or of the freshly fractured surface of the mineral quartz; resinous, greasy luster, like that of oiled glass; pearly luster and silky luster, or a reflectance similar to that of silk. The latter is generally related to a very fibrous mineral.

Descriptive Luster	Appearance	Mineral Optical Properties
Adamantine	Very shiny brilliant surfaces on non-opaque minerals	translucent / transparent, never opaque; RI 1.9 - 2.6
Subadamantine	< Adamantine	
Earthy / Dull	Matt surface, diffusive reflection, dull, absence of luster	usually opaque
Greasy / Oily	Grease coated looking surfaces	translucent to semi-opaque
Metallic	Reflective surfaces like polished metal	always opaque, even splinters
Submetallic	Almost metallic reflection	opaque, splinters are translucent
Pearly	Reminiscent of pearl or mollusk shell surfaces; Produced by lamellar microcleavage [***Note:*** *Pearly lustered minerals have perfect cleavage*]	translucent
Resinous	Resin like surfaces, commonly honey colored yellow to dark-brown minerals	translucent / transparent; moderate RI
Silky	Reminiscent of silk; produced by parallel mineral microfibers	translucent to semi-opaque
Vitreous / Glassy	Glassy looking surface; most common luster (70% of minerals)	translucent / transparent; RI 1.3 - 1.9
Waxy	Candle wax surface; semi-dull (< greasy luster). Commonly in cryptocrystalline or amorphous minerals	translucent to semi-opaque

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2.4 HARDNESS

Hardness: The resistance a mineral offers to abrasion, or it's hardness, can be determined by scratching it with a material or a mineral of known hardness. For example, a mineral of known hardness will scratch one of lesser hardness. In making such a hardness test remember that a scratch cannot be rubbed off. The powder formed in the hardness test may resemble a scratch, but if it can be rubbed away, and no scratch impression remains, the mineral has not been scratched. Simply check with your fingernail if a grove or scratch has occurred. If no scratch can be determined, the material used for the hardness test is softer than the tested or unknown material.

The hardness scale based on the relative hardness of certain minerals is known as the Mohs Scale of Hardness. Carl Friedrich Christian Mohs (1773 -1839) was a German geologist and mineralogist. Seeing the inadequacy of descriptive mineralogy in the early 19th century, Mohs developed this Hardness scale in 1813 while professor of mineralogy in Graz, Austria. The Mohs Hardness Scale did not become popular until 1820 when Mohs publication “Die Charaktere der Klassen, Ordnungen, Geschlechter und Arten des naturhistorischen Mineral-Systems” featuring his scale was widely circulated in the English language. It is still the present standard for classifying the hardness of various minerals.

While the scale is not quantitative in the absolute sense, it is most useful in identifying this special physical property of individual minerals. Ten minerals with different and progressively increasing hardness make up the scale. These, in order of increasing hardness, include: 1. Talc, 2. Gypsum, 3. Calcite, 4. Fluorite, 5. Apatite, 6. Feldspar, 7. Quartz, 8. Topaz, 9. Corundum, 10. Diamond. As a supplement to the Mohs Scale of Hardness, the fingernail has a hardness of 2.5, a copper penny 3.0, and a glass or pocket knife blade about 5.5. A good steel file is 6.0 on the scale. (See figure 2.2)

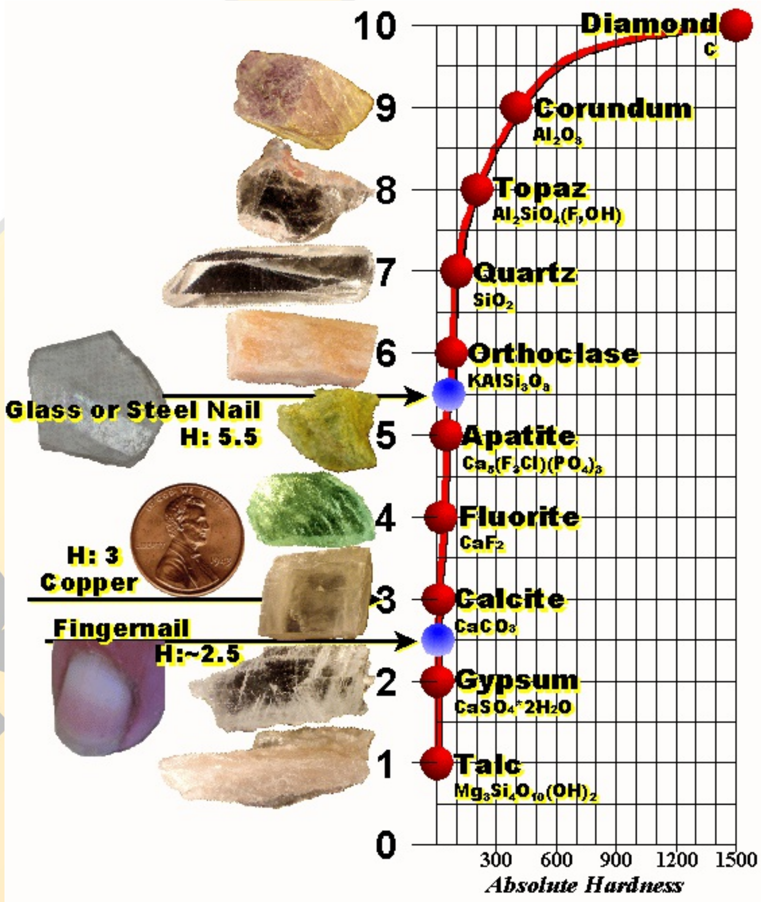

Figure 2.2 - Mohs Hardness Scale in relation to absolute hardness

The description of the individual minerals always includes the Mohs hardness number. Some minerals show different hardness numbers for the various cleavage faces. These variations relate to the internal atomic arrangement within the mineral. An example is kyanite. It has a hardness of 5 parallel to its length and 7 across the width of the crystal. Calcite is another example. It has a hardness of 3 on all surfaces except the base. The latter can be scratched by a fingernail of 2.5 hardness.

Figure 2.2 shows the Mohs Scale of Hardness and the minerals making up the scale in their position of relative hardness. In addition, a quantitative measurable hardness derived from precision instruments is also shown as the Absolute hardness for comparison.

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2.5 SPECIFIC GRAVITY

Specific Gravity (S.G.): Specific Gravity or density is defined as the ratio of the weight of an object compared to its volume. While determining the weight of an object is a simple and straight forward procedure, establishing volume with a high degree of accuracy might be somewhat daunting task. The process of establishing volume is based on Archimedes' Principle, where the volume of an irregular object equals the volume of a liquid displaced by the object, most commonly water. Specific gravity (*SG*) is obtained directly by determining the weight of the mineral in air (W_{air}), then obtaining the weight when the mineral is submerged in water (W_{inH2O}). The difference between the two weights (W_{air} - W_{inH2O}) is, of course, the weight of the volume of water equal to that displaced by the solid, which in turn is equal to the volume of the solid. D_{liquid} is a correction factor employed when using testing liquids other than water. The specific gravity is then determined by the equation:

$$SG(\tfrac{g}{cm^3}) = \frac{W_{air}(g)}{V(cm^3)} \times D_{liquid}\,\tfrac{g}{cm^3} = \frac{W_{air}(g)}{(W_{air}(g) - W_{inH_2O}(g))} \times D_{liquid}(\tfrac{g}{cm^3})$$

In research and mineralogical laboratories pycnometers and specialized balances are used for routine density determination. A pycnometer is a calibrated device that measures volumes of air, water or other liquids or gases displaced by the sample. A quick and very accurate method for routine SG measurements is the Single Pan Hydrostatic Method as explained below.

Other available liquids for SG determination and workable average D_{liquid} values are summarized in the following table:

70% Isopropyl Alcohol *	~ 0.80 - 0.92 g/cm^3	Ethanol	0.78 g/cm^3
Acetone	0.78 g/cm^3	Methanol	0.79 g/cm^3
Toluene	0.87 g/cm^3	Water	1.00 g/cm^3

70% Isopropyl Alcohol evaporates, hence densities increases with age: fresh 70% Isopropyl Alcohol ~ 0.85g/cm^3, dated 70% Isopropyl Alcohol ~ 0.88g/cm^3.

2.5.1 Accurate Single Pan Hydrostatic Method (see fig. 2.4)

This is by far the most accurate method and rivals those determinations made with expensive analytical laboratory balances. Developed by the author, this method was published as *Kackstaetter, U.R., 2014, A Rapid, Inexpensive and Portable Field and Laboratory Method to Accurately Determine the Specific Gravity of Rocks and Minerals; Peer Reviewed Article, The Professional Geologist, Vol.51, No. 2, Apr/May/Jun 2014, p.56 - 60).*

Figure 2.4 - Mineral Density Quick Test using the single pan hydrostatic method and a load cell balance

A load cell consumer balance with a minimum resolution of 0.01g is used. It is important that these load cell balances are very frequently calibrated using a precise calibration weight, especially when doing field measurement. This takes only a few seconds since the indicated consumer load cell balances often have a rapid or auto calibrate function. After calibration, an unknown mineral sample within the capacity of the scale is weighed directly on the balance and the result is recorded as weight in air (W_{air}). Then the sample is tied to a piece of string or thread with a slip knot. A plastic cup (~100mL) is filled with water, placed on the scale, and tarred to zero. Holding the sample by the string, the specimen is completely submerged without touching the sides of the vessel as indicated in figure 2.4. The readout on the scale is now equal to the volume of the sample, because the buoyant force of the water displaced equals the volume of the object. This force is measured directly by the balance. .

The density of water (D_{H2O})is considered to be 1.0 g/cm^3 for practical purposes and the above equation is simplified to:

$$SG(\tfrac{g}{cm^3}) = \frac{W_{air}(g)}{V(cm^3)}$$

Since samples measured by this method usually range several grams in weight, correction for water density according to temperature has little bearing on the calculated outcome and can therefore be ignored. For ease of calculation in the field and in the lab, a specifically designed Nomograph like the one developed in this manual can be used, eliminating the necessity for electronic calculation devices.

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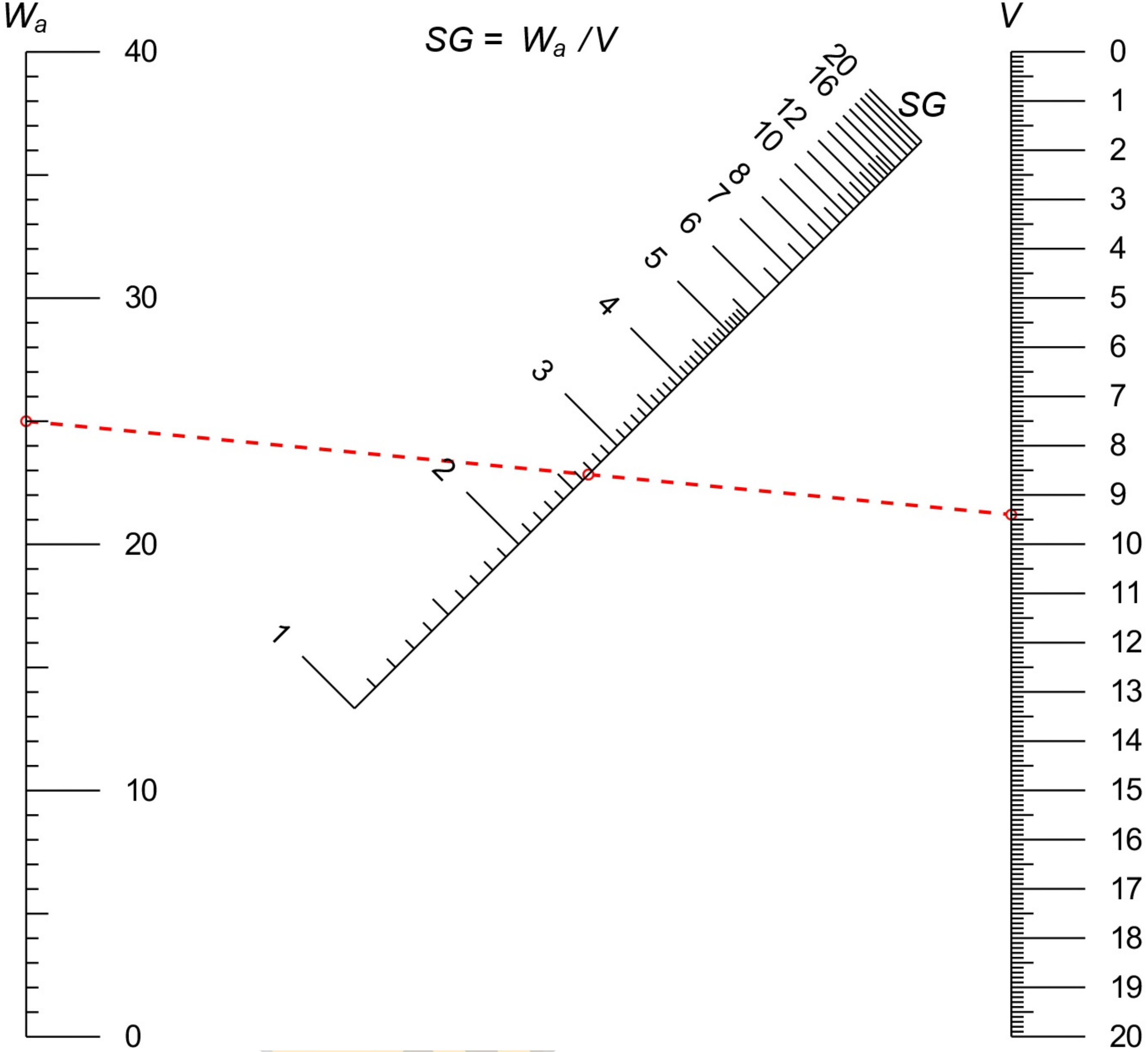
How to use a Nomograph: A nomograph is a graphical calculation device consisting usually of three scales or axis. When the measured values are marked on the outside scales and these marks are connected with a straightedge (or piece of string), the intersection of this straight line with the central axis or scale gives the answer.



Specific Gravity Nomograph for rapid determination of rock / mineral density using the Hydrostatic One Pan Method and a consumer load balanced weighing scale with a resolution of 0.01g. Nomograph generated using Python programming language scripting in conjunction with PyNomo Version 1.1 Release 0.2.2 software (Doerfler 2009).

Example: Plot W_{air} *of a specimen on the left vertical scale of the nomograph, here 25.00g. Mark V of specimen obtained through buoyancy measurements on the right vertical scale of the nomograph, example 9.40g. Connect both plots with a straight line. The intersection of this line with the diagonal scale give the specific gravity, here 2.65 g/cm*^{3}.

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2.5.2 PYCNOMETER MEASUREMENTS

Pycnometer measurements can be highly accurate, especially for small samples, if precise lab procedures and strict laboratory protocols are followed. A pycnometer for single mineral specimen testing consists of a vessel of known weight and volume. A special glass stopper assures that the volume in the bottle is constant when refilled and filled again. Excess liquid will just drain off. While distilled water is often used, other liquids can be employed, especially if the samples are small and the formation of bubbles interferes with accurate measurements. Before you start, make sure your mineral specimen fits through the opening of the bottle. Also measure the temperature of the distilled or DI water you will be using. Weigh the dry mineral first (W_{sample}) and then weigh the water filled pycnometer bottle (Pyc_{liquid}). Make sure there are no air bubbles inside. Any air is usually removed by placing the pycnometer into a vacuum desiccator for a few minutes. You may have to add more liquid to the pycnometer bottle to replace the volume lost through air bubbles. Make sure the outside of your vessel is completely dry and not a drop of liquid is clinging to it. Place the mineral sample in the bottle, excess liquid will spill, and replace the stopper. Again, avoid air bubbles using the desiccator. Wipe the outside of the vessel completely dry and weigh the pycnometer vessel with the mineral specimen and liquid ($Pyc_{sample+liquid}$). Now you can calculate the weight of the water displaced by specimen or specimen volume and conclude the specific gravity (SG) with the equation:

$$SG = \frac{W_{sample}}{\left(Pyc_{liquid} + W_{sample} - Pyc_{sample+liquid}\right)} \times D_{liquid}$$

The weight of the displaced water is directly related to the volume of the mineral specimen. Since 1g of distilled water is exactly equal to 1ml or 1cm^3 of volume only at 4°C, the volume changes slightly at elevated temperatures. As seen in the above equation, this change in specific gravity becomes a correction factor and your result must be corrected. Some values for distilled water are given in table 2.1.

Table 2.1 - Water densities at certain temperature for distilled water

Temp °C $_{H2O}$	D_{H2O}	Temp °C $_{H2O}$	D_{H2O}	Temp °C $_{H2O}$	D_{H2O}
15	0.999099	19	0.998405	23	0.997538
16	0.998943	20	0.998203	24	0.997296
17	0.998774	21	0.997992	25	0.997044
18	0.998595	22	0.997770	26	0.996783

For deionized water with an average content of 5mg/L TDS, densities at specific temperatures can be read from the DI Water Density Graph (fig. 2.5a). These densities can be used as correction factors when calculating mineral specific gravities using a pycnometer and the above equation.

Example (see fig. 2.5a): You measure a DI water temperature of 20.0°C. Locate the 20.0°C mark on the x-axis of the graph and draw a straight vertical line upward to intersect the graph (see red line). From the intersection point, draw a straight horizontal line to the y-axis (see red line). Read the DI water density of 0.99823 g/cm^3 on the y-scale.

Air bubbles clinging to pycnometer vessel walls and the mineral sample can be very cumbersome and yield erroneous results. Using liquids with lower surface tensions can effectively circumvent the bubble problem. While great variety of liquids may be used, 70% Isopropyl or "Rubbing" alcohol is readily available on an international basis. The procedure is exactly the same as with water explained above, except the density of the liquid much lower and a different density correction factor has to be used. The correction factor for 70% Isopropyl alcohol according to temperature is summarized in figure 2.5b.

Example (see fig. 2.5b): You measure a 70% isopropyl alcohol temperature of 20.0°C. Locate the 20.0°C mark on the x-axis of the graph and draw a straight vertical line upward to intersect the graph (see red line). From the intersection point, draw a straight horizontal line to the y-axis (see red line). Read the 70% isopropyl alcohol density of 0.8690 g/cm^3 on the y-scale.

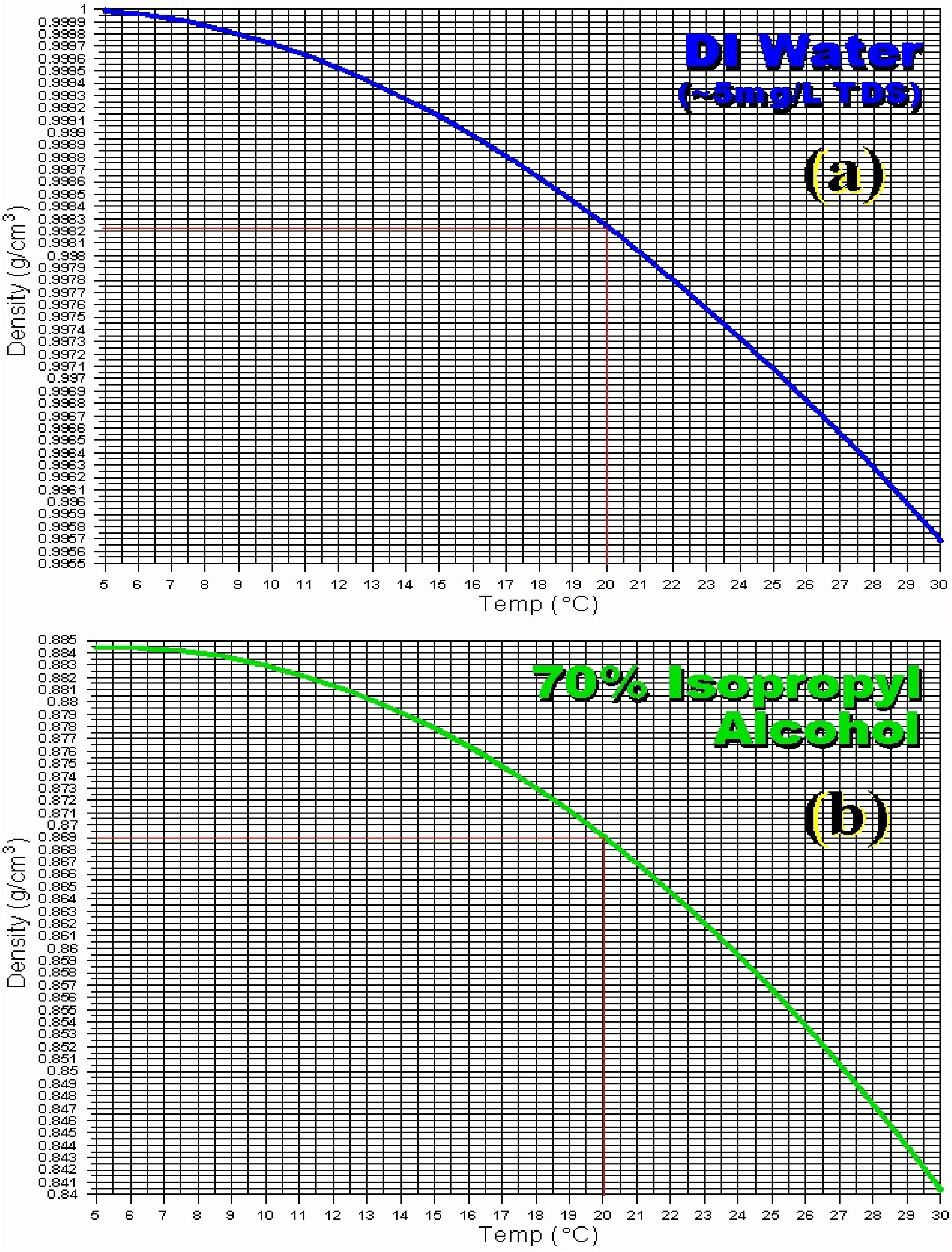

Figure 2.5 - (a) Densities of DI water (~5mg/L TDS) (b) Densities of 70% isopropyl alcohol for certain temperatures.

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2.5.3 JOLLY BALANCE

A popular instrument for determining mineral densities was the Jolly Balance (figure 2.6), invented by German physicist Philipp von Jolly in 1864. The instrument has been replaced by precise analytical balances in modern laboratories. However, on occasion one of these old instruments becomes available on Ebay, thrift stores or similar venues and its operation is included in this lab manual. The Jolly Balance consists of a spiral spring balance with two weighing pans, an upper one for measurements in air, and a lower one immersed in water. Weights of the mineral are established with a relative reading for the spring alone with the upper pan in air and the lower pan submerged in water (n), then placing the mineral on the pan in the air, leaving the lower pan immersed at the same depth in water and determining the scale reading (N,). The pan in water is raised out of the water, the sample placed on it and then it is again immersed in water with the sample on it to the same level. A scale reading is then obtained (N_2). The specific gravity (G) is thus:

$$G = \frac{N_1 - n}{N_1 - N_2}$$

Example:		
	$n = 35$	No mineral on pans, spring scale reading only
	$N_1 = 65$	Spring reading for mineral weight in air
	$N_2 = 55$	Spring reading for mineral weight in water

$$G = \frac{65 - 35}{65 - 55} = \frac{30}{10} = 3.0$$

2.5.4 BERMAN MICRO BALANCE

The Berman Micro Balance is another antique but sophisticate instrument for the determination of mineral densities. It measures the buoyancy of a unknown mineral in a liquid of known density with very low surface tension, such as toluene. Densities of pinhead sized samples could be accurately determined to the second, and even third decimal place for specific gravity with this instrument. While no longer commercially available, Huestis and Gormley from Pacific Lutheran University in Tacoma, WA have published an article in 1991 in the Journal of Chemical Education titled "Construction and Use of a Berman-Type Balance". All that is needed is a decommissioned single pan mechanical balance and their simple modification instruction to build your own. For those interested, their publication can be found at http://pubs.acs.org/doi/pdfplus/10.1021/ed068p882 .

2.5.5 HEAVY LIQUIDS

A multiple of much smaller mineral fragments occur in stream sediments and in rocks. A "sinkfloat" system can be used for those minerals of 2.8 or greater S.G. through the use of heavy liquids. While the dense (4.0 - 5.0 and very dense > 5.9) cannot be determined singly with this method, they can be grouped because of their relatively small number. For those minerals with S.G. between 2.89 - 4. 15 the judicious use of the heavy liquids permits fairly precise determinations of this mineral property, even with a sample size as small as a spec of dust. Table 2.2 identifies commonly used heavy liquids their S.G. and other important properties.

Table 2.2 - Commonly used heavy liquids for mineral density determination

Name	Chemistry	S.G.	Health Hazard
Bromoform	$CHBr_3$	2.889 g/cm^3	Lachrymator (eye irritant). Toxic. Carcinogenic.
Tetrabromoethane	$C_2H_2Br_4$	2.966 g/cm^3	Irritant.
Methylene Iodide	CH_2I_2	3.325 g/cm^3	Harmful. Slight irritant.
Clerici solution	Thallium(I) malonate/ formate solution	4.36 g/cm^3	Harmful by inhalation, in contact with skin and if swallowed.
Polytungstates	most commonly sodium polytungstate (SPT) and lithium metatungstate (LMT)	up to 2.9 g/cm^3 (room temp.) up to 3.5 g/cm^3 (elev. temp.)	None

If a mineral fragment is immersed in a heavy liquid of known density and the mineral floats. the mineral is less dense than the liquid. If it sinks, it is, of course more dense than the known liquid. In the latter case clean the mineral fragment and try immersion in the next denser liquid. If it falls between the two, the mineral identification charts can be used since specific gravity is the third identifier of five on these charts. If a precise determination of S.G. is required, then intermixing of the heavy liquids or diluting with acetone in known proportions gives a new liquid of known density in which the mineral can be tested. Practice can bring good results on the determination of this important physical property.

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While some of these chemicals pose serious health hazards when improperly used, the new development in polytungstates appears to solve this particular drawback. However, the later have the misfortune of high viscosities, thus increasing settling times in mineral density testing.

2.6 OTHER PHYSICAL PROPERTIES OF MINERALS

If it is not possible to make a specific mineral identification after careful check of the physical properties and reference to the mineral identification charts, then other tests may be necessary. Some of these are excellent for confirming a presumed identified mineral. These additional tests range from relatively simple physical to more complex chemical tests.

MAGNETIC RESPONSE: Simple magnets and strong neodymium magnets are very useful in identifying the four or five minerals which show sufficient magnetic response to be readily recognized. However, modern cell phones are made with a variety of build in electronic sensors. Powered by appropriate software apps, such as http://www.rotoview.com/magnetometer.htm , it will turn your phone into working magnetometer capable of measuring even small magnetic field strength emitted by a mineral. If such electronics are not available, suspending a small magnet on a thread or string and bringing the mineral close to it is a common alternate magnetic test. The most magnetic mineral is magnetite (Fe_3O_4). Lodestone, the very magnetic variety of magnetite, was once used as a crude compass in ancient times. It is recognized in the field by the broken grains which line up parallel to the magnetic field of the mineral when struck by a hammer.

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Pyrrhotite (FeS), an iron sulfide mineral, is slightly magnetic and affects a strong hand magnet. Likewise, chromite ($(Fe, Mg)Cr_2O_4$), a chromium oxide with a molecule of ferrous oxide, also slightly attracts the suspended hand magnet. Small particles of specular hematite (Fe_2O_3), are commonly affected by a strong magnet, as also franklinite ($(Fe,Mn,Zn)(Fe,Mn)_2O_4$) and even iron-bearing platinum. The magnetic character of the former two may result from minute magnetite inclusions.

The use of a very strong neodymium magnets permits a distinction of two classes of minerals: diamagnetic and paramagnetic. The latter are attracted by the magnetic field and the former repelled by it. The properties are commonly used in the laboratory and industry to separate various minerals. Quantitative separations of the diamagnetic and paramagnetic minerals, and even sets within the latter, are possible through the use of specially designed electromagnets such as the Franz Isodynamic Separator.

Paramagnetic minerals include siderite, diopside, beryl, chalcopyrite. pyrite, and others, especially those which are iron-bearing. Diamagnetic minerals are calcite, zircon, wulfenite. and others. Research has shown that the oriented magnetic properties of minerals are analogous to the optical properties of the same paramagnetic mineral.

TASTE & TONGUE: The sensation of taste can be determined only for those minerals which are easily soluble. Halite yields a salty taste, trona the taste of soda, epsomite a bitter taste, chalcanthite the taste of sulphuric acid and is nauseating. A cooling taste comes from niter and an astringent taste for alum. Glauberite has a bitter, salty taste, borates are salty, and borax-kernite has a sweet alkaline taste.

Other minerals have the tenacity to stick to your tongue when licked due to the absorption of water (The sensation is similar to licking a lamp post during freezing temperatures, alas not as intense. The sample can be easily removed from your tongue unlike from lamp-posts in the dead of winter). The most common minerals exhibiting this phenomena are Kaolinite or China Clay ($Al_2Si_2O_5(OH)_4$) - a whitish clay mineral; Garnierite ($(Ni,Mg)SiO_3 \times nH_2O$) - a green nickel ore; and Chrysocolla ($CuSiO_3$) - a greenish-blue copper ore mineral.

ODOR: Except in rare cases minerals do not give off odors, yet if moistened with the breath, or in other cases rubbed or struck, some volatiles may escape and be detected by their odor. Some calcites, for example, when rubbed, and also some quartz, yield a very fetid odor, like rotten eggs. This may represent release of hydrogen sulfide which was originally trapped in small vacuoles in the minerals when they formed.

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Kaolinite and other clay minerals, when moistened with the breath give off the odor of moistened clay. If arsenopyrite is rubbed vigorously by a harder object, a garlic odor is produced. Likewise pyrite and other sulfides give off a sulphurous odor when rubbed vigorously by a hard object. Heat will produce sulphurous fumes from sulfides, horse-radish odor from selenium-bearing ores, and garlic odor from arsenic containing compounds.

TENACITY: Tenacity includes brittleness, sectility, malleability and flexibility. Briefly defined brittleness refers to the mineral separating into a powder or grain when cut. Sectility is demonstrated when the mineral may be cut and not produce a powder or grains, though it may be powdered by hammering. Malleability refers to slices which may be cut off and flattened by hammering. Flexibility suggests the bending of a mineral without breaking and remaining bent after the stress is removed. Elasticity refers to the bent mineral returning to its original shape and size. Mica illustrates the latter, and talc the former.

ELECTRICAL PROPERTIES: Some minerals, especially those of metallic luster, conduct electricity. while those of non-metallic luster generally are non-conductors. The Isometric system shows equal conductivity in all directions and varies with the crystal system for other minerals.

Some minerals develop positive and negative charges on different parts of the same crystal when the temperature of the mineral is changed. These are called pyroelectric minerals. The phenomenon occurs only among the non-conductors. The best development is among the low-order symmetry minerals. Tourmaline is a good example with opposite charges developing at either end of the elongate crystal. Quartz shows a positive charge on heating on three alternate prisms and negative charges on the three other prismatic faces.

A piezoelectric quality has been attributed for some minerals subject to pressure or tension. Quartz, for example, has excellent piezoelectric properties when subject to pressure or an electric field. This special characteristic lead to its use in radio transmitters during World War II. Well-developed, clear crystals were much sought after during this period. The development of synthetic quartz crystal grown in the laboratory caused a great drop in demand for the natural crystal.

The interesting semiconductor property of the mineral galena (PbS) to act as a rectifier when in contact with metal, led to the development of the crystal radio in the early 1900's. Crystal radios do not need an external power supply, such as batteries, but convert the radio wave back into electricity. A thin, stiff wire, called a "cat's whisker" is dragged across the galena surface until a rectifier site on the crystal is found and a good signal is received.

RADIOACTIVITY: The radioactive decay of uranium and thorium toward their final lead isotope product is a process of emission of alpha particles, electrons and gamma ray radiation. The quantity of the daughter product lead to that of U or to Th is used to determine the age of the mineral. The radioactive emissions can be detected by analytical instruments such as a Geiger counter, a scintillator, or gamma ray spectrometer. The latter instrument identifies and quantifies daughter products of thorium and uranium and gives equivalent uranium and thorium content for the rock or ore samples. The radioactive isotope of potassium can also be measured. The other two instruments measure total radioactivity only. Even cell phones can be converted into working radioactivity testers. The inexpensive http://www.hotray-info.de/html/radioactivity.html app will turn your phone into a sophisticated, working radioactivity counter. Other USB driven Geiger Counters are also available and can be attached to tablets and laptops. With one of these relatively inexpensive and easily operated instruments radioactivity can be determined and in turn those minerals which contain uranium, thorium, and in some cases, radioactive potassium identified in their composition. Instruments sensitive to radioactivity are widely used in the exploration for uranium.

A group of about 150 minerals, some very rare, contain uranium or thorium as a part of their chemical composition. Some other minerals have radioactive elements in small quantities in solid solution. Based on economic value, uraninite and pitchblende, carnotite, Tyuyamunite, coffinite, autunite, torbernite, brannerite, and davidite are the most important of the uranium minerals. Other minerals which contain smaller uranium content are quite numerous. Select minerals of this latter group include monazite, the niobate-tantalates, allanite, zircon and some apatite. Thorium minerals of value are thorite and thorogummite and the oxide thorianite.

STREAK: Most commonly a small tile of unglazed porcelain called a streak plate is employed for the test. A free version of a streak plate may be obtained by using the back side of discarded tile cuttings from home improvement stores. The mineral to be tested is rubbed against such an unglazed porcelain surface and the resulting powder color is called the streak. For many metallic luster minerals this is an important diagnostic feature.

FUSIBILITY: This term refers to the reaction of a mineral chip or mineral powder to a hot flame. Candles, Bunsen or alcohol burners and micro-torches (available at hardware stores ; powered by a simple cigarette lighter; figure 2.6) are employed for determining fusibility of a n Microtorches for this purpose should be rated for temperatures of around 2,500°F or rough test a fragment is held in the flame by forceps or a tweezer in the oxidizing flame area just internal, light blue flame. The preferred area and hottest part of the flame is depicted in figure arsenic and antimony compounds could injure the platinum plating in a high quality forceps set are classified on a scale of 1 - 7, with higher numbers denoting greater resistance to heat. A r melts in a candle flame has a fusibility of 1 and 2. Those fusing readily in a Bunsen burner fusibility of 2. Fusibility of 3-6 is also observed with the use of a Bunsen burner. A mineral which is readily fusible in a micro-torch has a fusibility of 3. When only the splinters and thin edges are readily rounded, micro-torch fusibility is 4. When the splinters and edges are difficult to round, fusibility is 5. If only the tips of splinters and the thinnest edges show micro-torch fusion with difficulty, fusibility is 6. When the mineral is not affected or is infusible, micro-torch fusibility is rated at 7. Table 2.5 summarizes fusibility standards and observations.

Figure 2.6 - Micro-Torch

Table 2.5 - Fusibility ratings of standard minerals

Fusibility	Mineral Standard	Reaction	Temperature
1	Stibnite Sb_2S_3	Fuses easily in candle flame to globule	525°C 977°F
2	Chalcopyrite $CuFeS_2$	Fuses slowly in a candle flame to globule. Fuses in closed tube at full red heat	~800°C ~1,500°F
3	Almandine $Fe_3Al_2(SiO_4)_3$	Fuses readily to globule near oxidizing flame in micro torch	~1,050°C ~2,000°F
4	Actinolite $Ca_2(Mg,Fe)_5(OH/Si_4O_{11})_2$	Fragment edges readily rounded by micro torch oxidizing flame. Only fine splinters fused to globule	~1,200°C ~2,200°F
5	Orthoclase $KAlSi_3O_8$	Edges rounded with difficulty by micro torch	~1,300°C ~2,400°F
6	Bronzite $(Mg,Fe)SiO_3$	Only finest points and thinnest edges rounded by micro torch	~1,400°C ~2,550°F
7	Quartz SiO_2	Infusible	> 1,400°C > 2,550°F

Fusibility tests can easily combined with Flame Tests and tests for thermoluminescence. It should be noted that unstable minerals or those containing water in their structure may react differently. Crystalline water may make a mineral spit, sputter or intumesce in an open flame. Those minerals that will liberate gases upon heating are known to decrepitate or to turn into a slag-like substance.

Figure 2.7 - Flame Temperature of Micro-Torch

Abundance of iron in a mineral exposed to a flame may generate a black residue or globule which is often magnetic, even though the original mineral has no magnetic response.

3 CHEMICAL TESTING FOR MINERALS

Wet chemical tests for mineral identification are usually destructive. Luckily only very small amounts of the mineral in question are needed and the smallest grains usually suffice. Scrape or break of a small unobtrusive flake of the sample to be tested. Often the powder generated on your streak plate is sufficient to perform the chemical tests, especially for soft minerals. Harder minerals can be powdered using an emery board, which is made from aluminum oxide grit, as long as no test for aluminum is needed.

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Care should also be taken in using any of the reagents listed. They are vigorously reactive to clothes, skin, lips and eyes. Keep the reagents away from the body and skin. Remember that if you mix acid and water, never pour the water into the acid. The heat generated by the mixing may cause the acid to boil explosively and violently from the container. Always pour the acid very carefully into water. It sinks slowly through it because of its greater density and the heat of reaction is absorbed by the water. Because of the small quantities of acid used, this should pose no serious problem, but may if larger volumes are utilized.

3.1 FLAME TESTS

Flame tests can be performed in several ways. A common method is the use of a set of special forceps to hold the mineral which is then placed in a hot open flame such as that from a micro-torch (see figure 2.6). Observe the color of the flame produced. This color relates to the reactive element in the mineral. If the mineral fragment is first digested in Aqua Regia (3 parts HCl + 1 part HNO_3) and the residue of the resulting evaporate is moistened with HCl (hydrochloric acid) or HNO_3 (nitric acid) before performing the flame test, a better flame color often results. Examples of elements which produce good flame color tests are barium - which produces a distinctive green yellow - or lithium which yields a deep red against a dark background.

If an abundance of the mineral is available, the powdered mineral may be sprinkled into the flame and the color observed. For small mineral fragments, the fragment may be placed in a small loop in a platinum or NiChrom wire which has previously been imbedded in a handle of some sorts.

3.2 CLOSED & OPEN TUBE TESTS

Closed Tube Test: A closed tube test can be made by taking a 7-inch length of 1/4 inch outside diameter tubing, heating the center of the tubing to orange-redness over a Bunsen burner flame and when it is softened, pulling it apart and closing the ends through reheating of the two parts. Small borate test tubes are cheap and will serve nicely for the same purpose. For use in mineral identification, the sample is heated in the closed end of the tube. Odors and sublimates in the tube are noted and compared to those which apply to the mineral being tested.

Open Tube Test: Open tube tests are conducted by using a glass tube about 3/8 inch in outside diameter and about 5-6 inches in length. To prepare an open tube, heat the glass tube and turn it slowly In the flame. When it is glowing orange-red bend it carefully into a 30° elbow, leaving 3/4 inch of the tube after the bend. After cooling, a small mineral fragment can be placed into the bend inside the tube and then heated carefully over the flame. The short end of the tube should be held nearly horizontal with the longer portion inclining upward. The objective is to form mineral oxides, hence heating should be slow but thorough to assure complete oxidation of the mineral sample. Best results are obtained when the mineral is finely powdered. Check for odors from the tube and watch for sublimates which might form on the inside of the tube. The mineral which you are attempting to confirm will indicate in its description if it is amenable to this type test. Open tube tests are very reliable assays when confirming S, As, Sb, Hg, Te, or Se in sulfides. Since open tubes are often impossible to clean after tests, you may want to manufacture several tubes at once.

3.3 SOLUBILITY TESTS FOR MINERAL IDENTIFICATION

The normal procedure is to first check for the solubility of a mineral fragment in water. If the mineral gradually disappears or dissolves in the water, it is considered soluble. It should fall within that category of the table.

If solubility is not visible in water, then a clear fragment of the mineral may be tested for solubility in 1 :7 (one volume of acid to 7 parts water) HCl (hydrochloric acid), which is your regular hydrochloric testing acid, used for field and laboratory tests. If bubbles arise in the liquid drop or strong foaming appears, the mineral is said to effervesce. The presence of effervescence indicates the generation of a gas. e.g., calcite ($CaCO_3$) effervesces vigorously in HCl acid and CO_2 (carbon dioxide) is generated and the remaining calcium element remains as ion in solution. If the test is made on a rock and the mineral fragment is small, the action of a reactive effervescent mineral can be heard by bringing the specimen close to, but not touching the ear. The slight “pop-pop” sound confirms the effervescence.
Other stronger acids may be used on mineral fragments not reactive to water or 1: 7 HCl. Minerals reactive to these acids and the character of the reaction are also shown in Table 3.4.

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One of the strongest acids to use in mineral identification is *Aqua Regia*, a mixture consisting of one part Nitric Acid (HNO_3) to three parts of Hydrochlorid Acid (HCl) by volume. *Aqua Regia* is so strong that it will dissolve even minerals and metals, such as gold or platinum, otherwise unscathed by caustic chemicals. It is a great tool in identifying certain minerals. However, mixing acids just for fun is not a good idea. Nitric Acid can react with organic residues creating explosive compounds which have mauled "want-to-be" scientists. Create only small quantities for immediate use and keep Nitric Acid away from organic materials.

3.4 FIELD CHEMICAL CHROMATOGRAPHIC ANALYSIS OF MINERALS

In many cases a more sophisticated test for mineral identification is desired. The systematic chromatographic chemical techniques described are a unique yet inexpensive and rapid measure to determine mineralogy even in the field and can be used as an additional confirmatory method of identification. Prerequisite is a good working knowledge of mineral chemical formulas in order to know what to test for.

Field Wet Paper Chromatography

Paper Chromatography is used to separate ions due to their varied travel times in various substances, as they migrate upward a absorbent material through capillary action. While often described as a teaching tool, A.S. Richie developed a field portable method in the early sixties to qualitatively distinguish between major ions in ore minerals. The test is relatively quick, inexpensive, and uses only a limited number of chemicals to obtain usable results.

In principle, a small amount of sample is digested and the resultant digest containing the dissolved ions is transferred close to the bottom of a chromatographic paper strip. A carefully formulated solvent is then placed at the bottom of the chromatography paper and is allowed to absorb upward through the paper. In the process, ions present in the digest will start to migrate with the solvent at varied rates, starting to separate into their constituent groupings. Spots will develop on the chromatograph at certain distances from the starting position of the digest, which are indicative of certain ions. Treating the paper with developer solutions afterwards will yield additional formerly invisible ion spots. Even more ions can be detected or confirmed when the chromatograph is viewed in UV light after developing.

The following ions are usually visible without developer solutions on the raw chromatographic paper strip, depending on the solvent used: Au, Bi, Cd, Co, Cr, Cu, Fe, Mn, Ni, Pd, Ti, and V. Adding developer solution will enhance detection of more ions. Developers are for the most part chemical chelating agents or dies that form colorful complexes when reacting with metal ions. The most powerful developer solutions are:

- 8 Hydroxyquinoline: 0.5% in ethyl alcohol, which can detect Al, Ag, Au, Ba, Be, Bi,Ca, Cd, Co, Cr, Cs, Cu, Fe, Ga, Ge, Hf, Hg, In, K, La, Li, Mg, Mn, Mo, Nb, Ni, Pb, Pd, Pt, Rb, Sb, Sc, Sn, Sr, Ta, Th, Ti, Tl, U, V, W, Zn, and Zr
- Alizarin: A saturated solution in ethyl alcohol, which determines Al, As, Bi, Ce, Cr, Cs, Cu, Fe, Hg, In, Li, Mg, Mn, Pb, Sb, Ta, Th, Ti, V, W, Y, Zn, and Zr

Even common household spices may work. Curcumin, the main ingredient of Turmeric, can detect Ag, Al, Au, B, Be, Cr, Cu, Fe, Li, Ni, Pt, Ta, Ti, V, W, and Zr. For this a 0.1% solution of Turmeric in ethyl alcohol is employed.

Next to specific coloration of the chromatography spots for elemental identification, either developed or undeveloped, is their position or travel distance. This distance is expressed as the Retention Factor or R_f value and is calculated as a ratio between the distance of the ion spot from the starting baseline and the distance of the solvent traveled along the length of the chromatography paper visible as a wetting front. If the value is 1, then the ion has traveled with the wetting front and no retention is experienced. If the value is 0 then the solvent had no affect and the ion did not move. Most ions, however, do travel at least some distance and therefore will have an R_f value. Note that the R_f value is influenced by the solvent used and temperature. Therefore care should be taken when interpreting the results and some deviation from published values is to be expected.

A materials list for a Chromatography Assay Kit is given in Table 3.5 and the indicated supplies are suggested to keep results consistent. Sample digestion of unknown material is carried out either on glass slides or heat retardant crucible. The microtorch is used to heat the glass slides or crucible and the digestive reagents. The capillary tubes are used to transfer the digest to the chromatography paper.

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Table 3.5 - Reagents & Supply List for the Chromatography Assay Kit

Digestive Chemicals and Equipment	Chromatographic Chemicals and Equipment
☐1:1HCl ☐1:1 HNO_3 or ☐NH_4Cl ☐NH_4NO_3 ☐Microscope Slide ☐High T crucible ☐Microtorch ☐1:7 HCl ☐Toothpicks ☐Capillary tubes	☐Solvent3 solution: 15mL Ethanol + 15mL Methanol + 20mL 2N HCl ☐NH_4OH solution in Spray Bottle ☐0.5% 8-Hydroxyquinoline in ethanol developer solution in Spray Bottle ☐UV Light ☐Ruler ☐2cm x 12cm Chromatography Paper Strips ☐12cm tall Plastic vessel w/ lid for chromatography

The chromatographic procedures used for elemental and mineral determination are those outlined by A.S. Richie (1961, A Paper Chromatographic Scheme for the Identification of Metallic Ions; 1962, the Identification of Metal Ions in Ore Minerals by Paper Chromatography. Part I. Opaque Ore Minerals) and J. Crawford (2009, Chemical Tests for Small Specimens). The great majority of the elements are identified on the basis of R_f values and spot color development. In order for spots to show a good color, the reagent used to bring about the precipitation or color change should be more concentrated than the unknown ion to be tested. If concentrations are too low, no color will form.

Keep in mind that the chromatographic tests suggested are not necessarily diagnostic of the mineral itself, but rather of the elements present in the mineral. For example, a green raw spot and developed orange spot with an R_f of 0.65 is indicative of Cu, which of coarse may belong to a whole list of Cu bearing minerals. Impurities, such as pyrite (FeS_2) in chalcocite (Cu_2S) may result in misinterpreted positives for mineral identification. In many cases the chromatographic testing may not singularly identify one mineral in preference to another. Other properties of the minerals used in conjunction with elemental composition, however, will permit identification.

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Experimentation and experience with the use of digestion, solvents and developing solutions will yield predictable results for your procedures and identifications. By taking care of your cleanliness of your glass slides, the rate and amount of heat applied in order to dissolve the mineral, the concentration of your digest, and other related factors, you can markedly influence the end results of the chromatography tests. Positive results are attainable. The important point is to take care in the steps of the method used.

DISCLAIMER: This manual is intended for use by persons with a basic knowledge of inorganic chemistry, they are advised to follow strictly the safety instructions. The author does not accept liability or responsibility for any injury or damage to persons or property incurred by performing the experiments described in this manual, nor for the content of any outside material referred to in this manual, including linked websites.

4 SOPHISTICATED INSTRUMENTAL GEOCHEMICAL ANALYSIS

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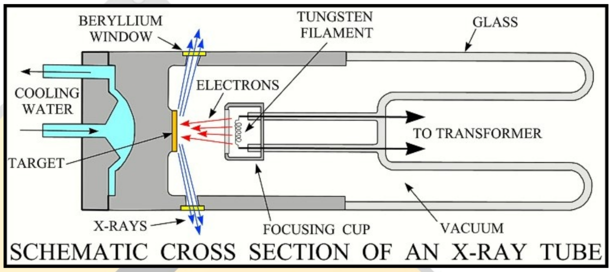
5 X-RAY DIFFRACTION

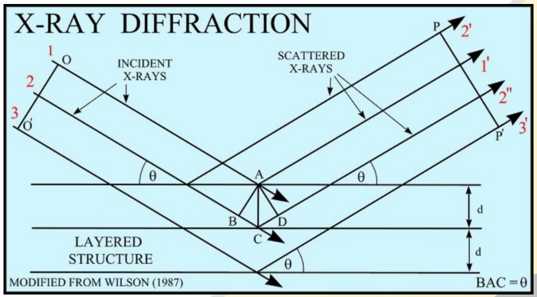
The most common routine approach for mineral identification of fine-grained particles are X-ray diffractive (XRD) methods. The following is a reprint from the U. S. Geological Survey Open-File Report 01-041 - A Laboratory Manual for X-Ray Powder Diffraction by L.J. Poppe, V.F. Paskevich, J.C. Hathaway, and D.S. Blackwood. This public domain edition was specifically written for laboratory application. Text and graphics are taken verbatim from the publication. Reprinted under the public domain agreement from http://pubs.usgs.gov/of/2001/of01-041/ .

X-Ray Diffraction Primer

X-rays are electromagnetic radiation similar to light, but with a much shorter wavelength. They are produced when electrically charged particles of sufficient energy are decelerated. In an X-ray tube, the high voltage maintained across the electrodes draws electrons toward a metal target (the anode). X-rays are produced at the point of impact, and radiate in all directions. Tubes with copper targets, which produce their strongest characteristic radiation (Kα1) at a wavelength of about 1.5 angstroms, are commonly used for geological applications.



SCHEMATIC CROSS SECTION OF AN X-RAY TUBE


X-RAY DIFFRACTION

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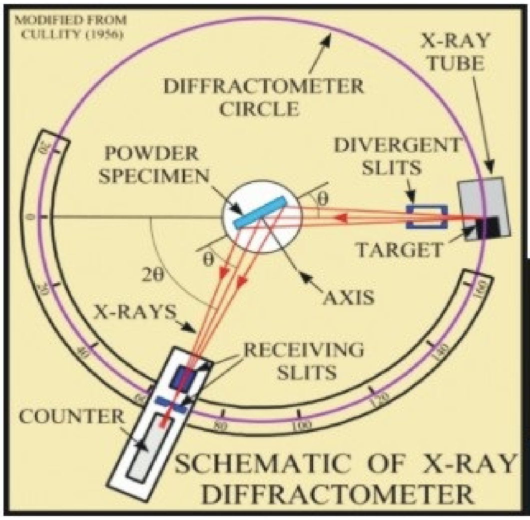
If an incident X-ray beam encounters a crystal lattice, general scattering occurs. Although most scattering interferes with itself and is eliminated (destructive interference), diffraction occurs when scattering in a certain direction is in phase with scattered rays from other atomic planes. Under this condition the reflections combine to form new enhanced wave fronts that mutually reinforce each other (constructive interference). The relation by which diffraction occurs is known as the Bragg law or equation. Because each crystalline material has a characteristic atomic structure, it will diffract X-rays in a unique characteristic pattern.

The basic geometry of an X-ray diffractometer involves a source of monochromatic radiation and an X-ray detector situated on the circumference of a graduated circle centered on the powder specimen. Divergent slits, located between the X-ray source and the specimen, and divergent slits, located between the specimen and the detector, limit scattered (non-diffracted) radiation, reduce background noise, and collimate the radiation. The detector and specimen holder are mechanically coupled with a goniometer so that a rotation of the detector through 2x degrees occurs in conjunction with the rotation of the specimen through x degrees, a fixed 2:1 ratio.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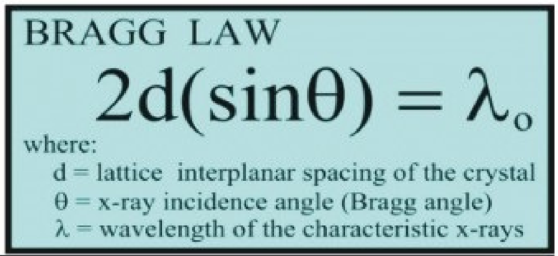
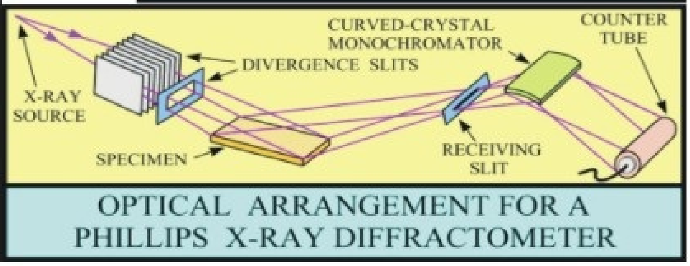
SCHEMATIC OF X-RAY DIFFRACTOMETER


BRAGG LAW

$$2d(\sin\theta) = \lambda_o$$

where:

d = lattice interplanar spacing of the crystal
θ = x-ray incidence angle (Bragg angle)
λ = wavelength of the characteristic x-rays



OPTICAL ARRANGEMENT FOR A PHILLIPS X-RAY DIFFRACTOMETER

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A curved-crystal monochromator containing a graphite crystal is normally used to ensure that the detected radiation is monochromatic. When positioned properly just in front of the detector, only the $K\alpha$ radiation is directed into the detector, and the K_β radiation, because it is diffracted at a slightly different angle, is directed away. The signals from the detector are filtered by pulse-height analysis, scaled to measurable proportions, and sent to a linear rate meter for conversion into a continuous current. Common output devices include strip-chart recorders, printers, and computer monitors.

Preparatory and Interpretive Procedures

Note: While many preparatory techniques are listed, only the techniques required in this course are reproduced.

SMEAR SLIDE SAMPLE MOUNTS FOR X-RAY POWDER DIFFRACTION

Samples may be ground and smeared on the surface of a glass slide as a rapid means of preparing samples for X-ray powder diffraction. This mounting method seldom produces acceptable random orientation of the crystallites, resulting in diffraction maxima with relative intensities that are not accurately reproduced. Although not useful for semi-quantitative analysis, this method is useful for rapidly determining bulk mineralogy, especially if the operator knows which mineral phases are likely present.

PREPARATION OF SMEAR-SLIDE MOUNTS

Materials Required:
weighing paper (4"x4")
mortar and pestle
spatula
glass rod
marker pen
glass slide cut to fit the diffractometers's sample holder
acetone in a squeeze bottle
lab tissues
diamond or carbide scribe to cut slides

Grind the dried sample thoroughly with the mortar and pestle. The particles should be much finer than .062 mm to avoid fractionation of the minerals. The finer the powder the greater the opportunity for obtaining an adequate number of particles with random orientation and the less likely that surface roughness will reduce low-angle intensities.

Use the spatula to loosen any sample that has stuck to the mortar and to homogenize the powder.

Place a small amount of sample in the center of the slide and add an amount of acetone sufficient to wet the sample (2-3 drops). Distilled water may be used, but the sample dries quicker with acetone.

Spread the sample into a thin layer with the glass rod. Gently blow on the slide to rapidly evaporate the acetone. Use the spatula or a finger to remove any excess sample sticking to the edges of the slide. Use a lab tissue to remove any sample stuck to the back of the slide.

Label the slide with a marker pen. Brush the sample from the mortar onto the weighing paper and transfer the remaining sample back into a sample vial.

The output of the XRD unit will be presented as a chart showing the theta angles on the bottom and peaks of various heights representing d-spacings. A typical resultant chart of a mixed clay mineral sample is presented in figure 5.1. While students had to sort comparative d-spacing tables to identify fine grained minerals, most modern XRD units include automated data bases, preselecting the most appropriate mineral phases.

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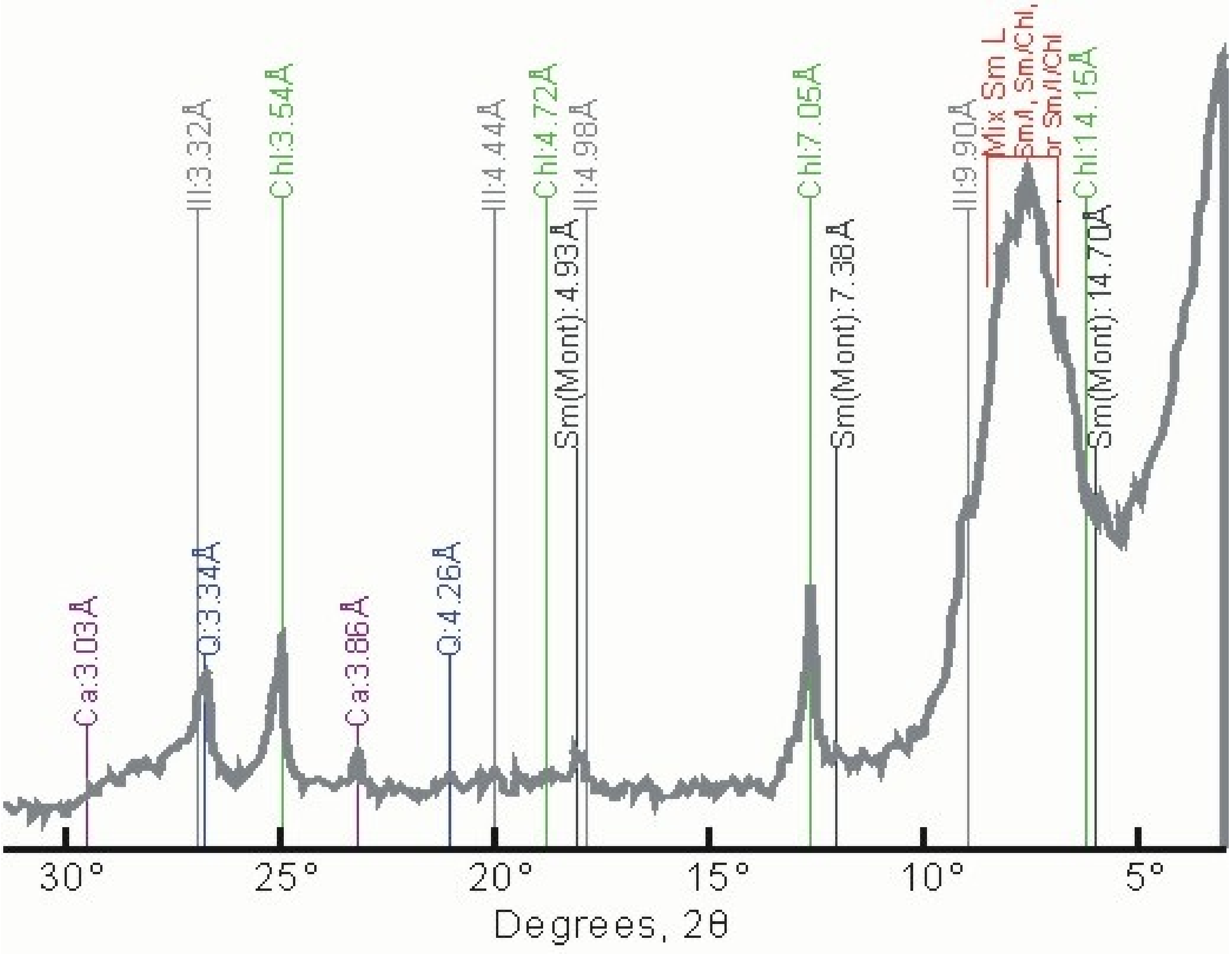

Figure 5.1 - X-ray pattern of untreated sedimentary sample, particle size <2µm. Identification of major peaks: Ill=Illite, Chl=Chlorite, Sm=Smectite, Mont=Montmorillonite, Q=Quartz, Ca=Calcite

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6 OPTICAL MINERALOGY

Optical mineralogy uses a light transmission microscope and thins sections of rock ground to a thickness of 30µm or 0.03mm. (In comparison, human hair has a thickness from 17 to about 180µm depending on race and hair color, with black hair being the thickest and blonde or flaxen hair the finest.) Polarized light is channeled through the now translucent mineral crystals which start to refract the light according to their specific optical properties. This refraction can be observed and individual mineral grains can be identified according to their refractive behavior. The refractive properties are tightly controlled by the crystallography of the mineral, hence a sound understanding of crystal classes is essential for the identification processes.

6.1 THE PROPERTIES OF LIGHT

In order to use the optical properties for mineral identification one must have a grasp of the basic properties of light and optical physics. Visible light comes in various wave indicative of a specific color with 790nm (nanometer = 10^{-9} m or 10^{-6} mm) at the red or long wavelength visible boundary and 390nm at the violet or short wavelength visible boundary. For all purposes the speed of light is the fastest entity known to men, usually indicated as $c = 2.99792458 \times 10^8$m/s or 7 ½ times around the earth in one second for comparison. This velocity is generally rounded to 3.00×10^8 m/s or 300,000 km/s.

Light travels with different speed in different media, like water or glass. This change in speed is generally termed “refraction”and the amount by which light slows in a given material is described by the index of refraction, n. The index of refraction of a material can be calculated as

$$n = \frac{c}{v}$$

which is the speed of light in vacuum *c* divided by the speed of light through the material *v*. Since light is always fasted in a vacuum, the refractive index of materials will always be greater than 1. The following table 6.1 gives a summary of refractive indices for various materials and minerals.

Table 6.1 - Some refractive indices *n* for selected materials

material	n	material	n	material	n	material	n
Vacuum	1.000	Air	1.0003	Water	1.33	Pyrex glass	1.470
Fluorite	1.434	Halite	1.544	Grossularite	1.734	Diamond	2.418
Sphalerite	2.369	Spinel	1.719				

To simplify matters we will call the index of refraction of air the same as the index of refraction in a vacuum or 1. There is a relationship between the density of the mineral and the indicate index of refraction. In general, the denser the mineral the greater the index of refraction. Gemstones usually must have a high index of refraction, preferably > 2.0 to display brilliance or fire.

The refractive index of minerals is not necessarily the same in all directions. The only exception are isometric minerals where the velocity of light will propagate equally through the crystal. They are said to be optically isotropic.

All other crystal classes display anisotropic behavior with more than one refractive index. Table 6.2 summarizes the crystal classes and their optical behaviors.

Table 6.2 - Crystal classes and optical behavior of minerals

Optical Behavior	Crystal Classes	Examples
Isotropic (1 refractive index)	Isometric	Garnet, Fluorite, Diamond
Anisotropic - Uniaxial (2 refractive indices)	Hexagonal Tetragonal	Quartz, Calcite
Anisotropic - Biaxial (3 refractive indices)	Triclinic Monoclinic Orthorhombic	Orthoclase, Barite

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Polarization of Light

Light may vibrate in all directions and planes unless it passes through a polarization filter. Polarized light is comprised of all light rays vibrating parallel to each other in the same direction. Light which is contraire to the polarization lense will simply be absorbed. If a second polarizer is placed perpendicular over the first, all light will be absorbed and any image projected will appear black. An optical microscope for mineral identification has two polarization filters, also called polars. When both polars are engaged or crossed and no mineral specimen is placed on the stage, the field of view appears black. Isotropic minerals are easily identified by an optical microscope since they do not interfere with the vibration of polarized light. Isotropic minerals always appear black under crossed polars, but are nicely visible when the polars are uncrossed.

Anisotropic minerals on the other hand do interfere with the polarize light and actually split the income rays into two beams, a fast one with a refractive index of *n* and a slow beam with a refractive index of *N*. Hence anisotropic minerals present an interesting light show under crossed polars when the microscope stage and with it the minerals specimens are rotated. Anisotropic minerals will appear dark or extinct about every 90° of stage rotation. If the microscope stage is rotate further, the mineral will become light again. This dual property of N-n is called birefringence which gives rise to interference colors. These colors can be plotted on a birefringence chart relating directly to thinsection thickness and mineral identification (figure 6.1).

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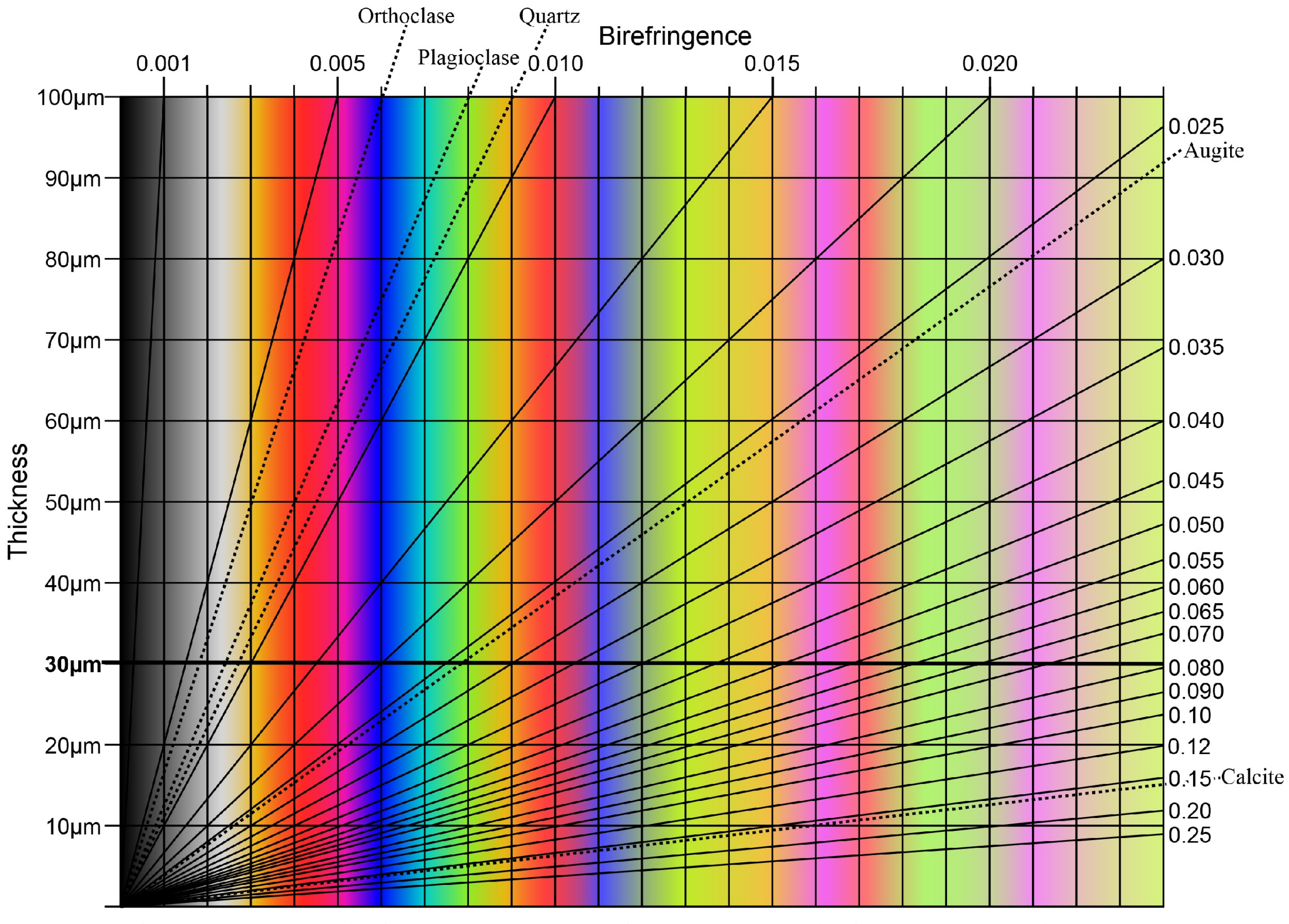

Figure 6.1 - Optical mineralogy birefringence chart

6.2 OPTICAL MICROSCOPY & THIN SECTIONS

A great detail about analyzing thin sections is available as an excellent FREE Open Access Publication titled:

Guide to Thin Section Microscopy

Michael M. Raith, Peter Raase, and Jurgen Reinhardt (2012, second edition) 127 pp. ISBN 978-3-00-037671-9 (English); 978-3-00-036420-4 (German); 978-3-00-040623-2 (Spanish); 978-3-00-046279-5 (Portuguese).

This work is available in several language and can be downloaded at NO COST as indicated below:

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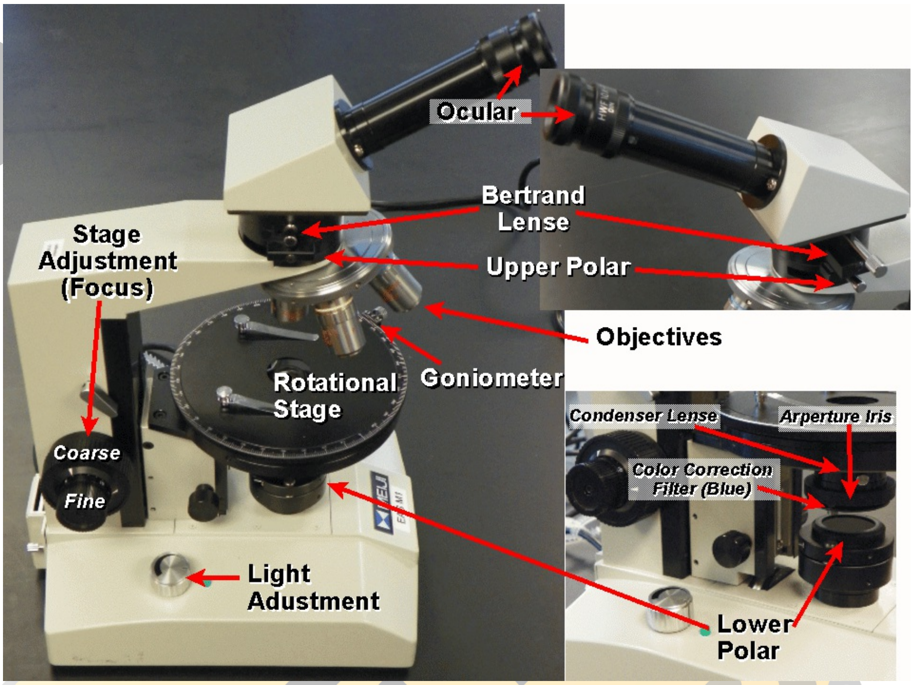
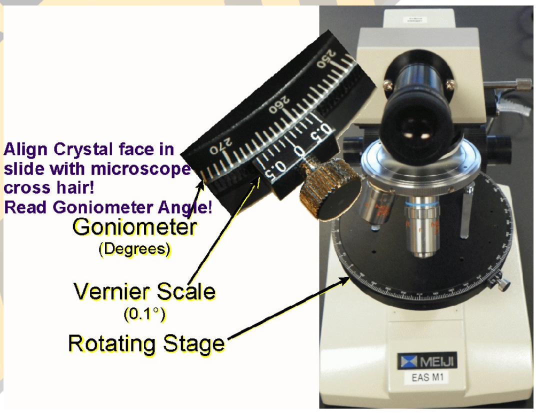
6.3 THE OPTICAL MICROSCOPE

The graphic on the right is a quick overview of a simple Polarized Light Microscope (PLM). Next to the obvious use of polarized light, these microscopes have rotational stages which are necessary for proper mineral identification. Precise measurements of angles are possible with the printed goniometer scale on the circumference of the stage.



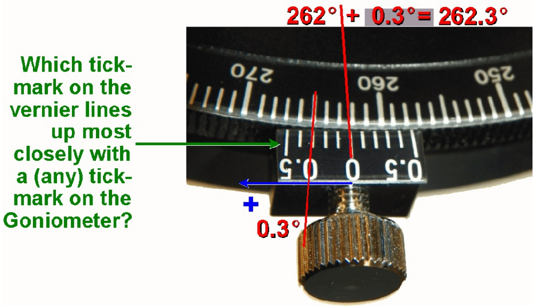
6.3.1 ANGULAR MEASUREMENT USING THE PLM GONIOMETER

The goniometer and associated vernier scale on a PLM allows for the precise measurement of angles, on crystal faces for example. A crystal face edge is aligned with the cross hair in the objective and the exact angle is noted. The stage is then rotated until the next crystal face aligns in the cross hairs and another angle reading is performed. The difference of these two angles gives the precise angular intersection of the two crystal faces.



How to read precise angles using the vernier scale in conjunction with the goniometer on a PLM? Follow the example as pictured:

1. Use the “0" tick mark on the vernier scale to identify which angle value tick mark is closest on the goniometer. In the example the value falls between 262° and 263°. The exact value is therefore 262° plus “some” degree decimals.
2. To read the degree decimals, use the vernier scale. Which tick marks on the vernier scale match most closely with ANY tick marks on the goniometer, even outside 262° or 263°. In the example the best match lines up with the 0.3 mark on the vernier, on the left side of the vernier “0" mark. We therefore add 0.3° to the 262° reading. Should a tick mark on the right side of the “0" align, then we subtract the decimal value from the higher number of our initial “between” readings.
3. In the example our reading is 262° + 0.3°, which is precisely an angle of 262.3°.



7 IGNEOUS ROCKS

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7.1 VOLCANIC (EXTRUSIVE) IGNEOUS ROCKS

Volcanic rocks solidify from a lava (magma) on the earth's surface and form lava flows. The rock which results from the crystallization of a lava is determined by the original composition and its cooling history. In contrast to a lava flow, a burled or partly burled magma, under high vapor pressure or its solidified equivalent may be explosively ejected into the air. The resultant broken fragments or breccia and fine particles fall through the atmosphere and form ash deposits such as *tuff,* or ash and fragmental material called *tuff breccia.* In certain types of volcanic eruptions, partly solidified magma with included gases and molten magma may blow out almost horizontally from a vent and form glowing avalanches or *ash-flows.* Some of these eruptions which are very violent in character may cover tens of thousands of square miles in a series of eruptions. In the rare case where a magma, or lava or ash flow, arrives on the surface and is very rapidly cooled, a *glass,* or *vitrophere* (if crystals are visible in the glass matrix) may result. If very rapid loss of vapor occurs from a magma and it also cools quickly, a frothy, glassy rock results and *pumice* forms. This may also be blown explosively into the atmosphere and fall to the earth and be incorporated into other volcanic material to form *pumice tuff breccias,* or if the explosively generative particles are sand size or less, a *pumicite* may form.

Table 7.1 - The More Abundant Minerals of Igneous and Sedimentary Rocks

Mineral	**Igneous Rocks**	**Sedimentary Rocks**
Feldspars	50 %	16 %
Quartz	21 %	35 %
Pyroxene, Amphibole, Olivine	17 %	-
Micas	8 %	15 %
Magnetite	3 %	-
Clay Minerals	-	9 %
Dolomite	-	9 %
Chlorite	-	5 %
Calcite	-	4 %
Hematite, Limonite	-	4 %
Others*	1 %	3 %

*Others may include: Titanite, Ilmenite for igneous rocks; and Apatite, Topaz, Zircon for sedimentary rocks.

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7.2 PLUTONIC (INTRUSIVE) IGNEOUS ROCKS

More often, the magma does not reach the surface of the earth. In this environment it cools or solidifies below the surface in the host rock it has invaded or partly assimilated. Because of its much lower rate of heat loss in this setting, the crystallization process has more time, nucleation of crystals begin, seed minerals are able to grow larger, and the final solidified rock mass has a crystalline aggregate of minerals readily visible to the eye in a much finer-grained matrix. With longer cooling time, and possibly greater depth of the cooling magma below the surface, the mineral growth may finally be almost equigranular, medium to coarse-grained, and of *phaneritic* texture. These mineral crystals are called *phenocrysts* and the texture of the igneous rock is *porphyritic.* Thus, a rhyolite porphyry is a rock whose composition is very much like granite, but whose visible minerals are few to as much as 25% of the total rock volume. The minerals include quartz and feldspar, the latter mainly of the orthoclase variety.

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7.3 IDENTIFYING IGNEOUS ROCKS

The general color of the solidified igneous rock is not only a guide to its name, but also suggests the chemical composition, and in turn the minerals of the rock. The light- colored or felsic rocks are made up of an abundance of quartz and feldspars, either megascopic (visible) or microscopic in size. Rhyolite and light-colored pumice and granite are in this group. The darker colored, more iron-rich, or mafic (magnesium- ferrum (Latin: Iron) rich) igneous rocks consist of the more calcium-rich feldspars and the magnesium-iron bearing biotite, amphiboles, pyroxenes, and olivine. These include extrusive basalts and the intrusive equivalent gabbro rocks of the crust. An intermediate colored and intermediate compositional group lies between the felsic and mafic groups. It includes the diorites and andesites, which are among the very abundant igneous rock types in the world.

The most simple classification of igneous rocks would be to describe them as light- colored, medium-colored and dark-colored. In essence we do this by listing them as felsic, intermediate, and mafic, and even ultra-mafic. The classification thus denotes feldspar-silica-rich rocks, through those which are made up of increasing contents of magnesium, calcium, and iron to those which consist essentially of high content of iron-magnesium-calcium with low silica content. The textural characteristic of these rock groups indicate the crystallization history of the rock.

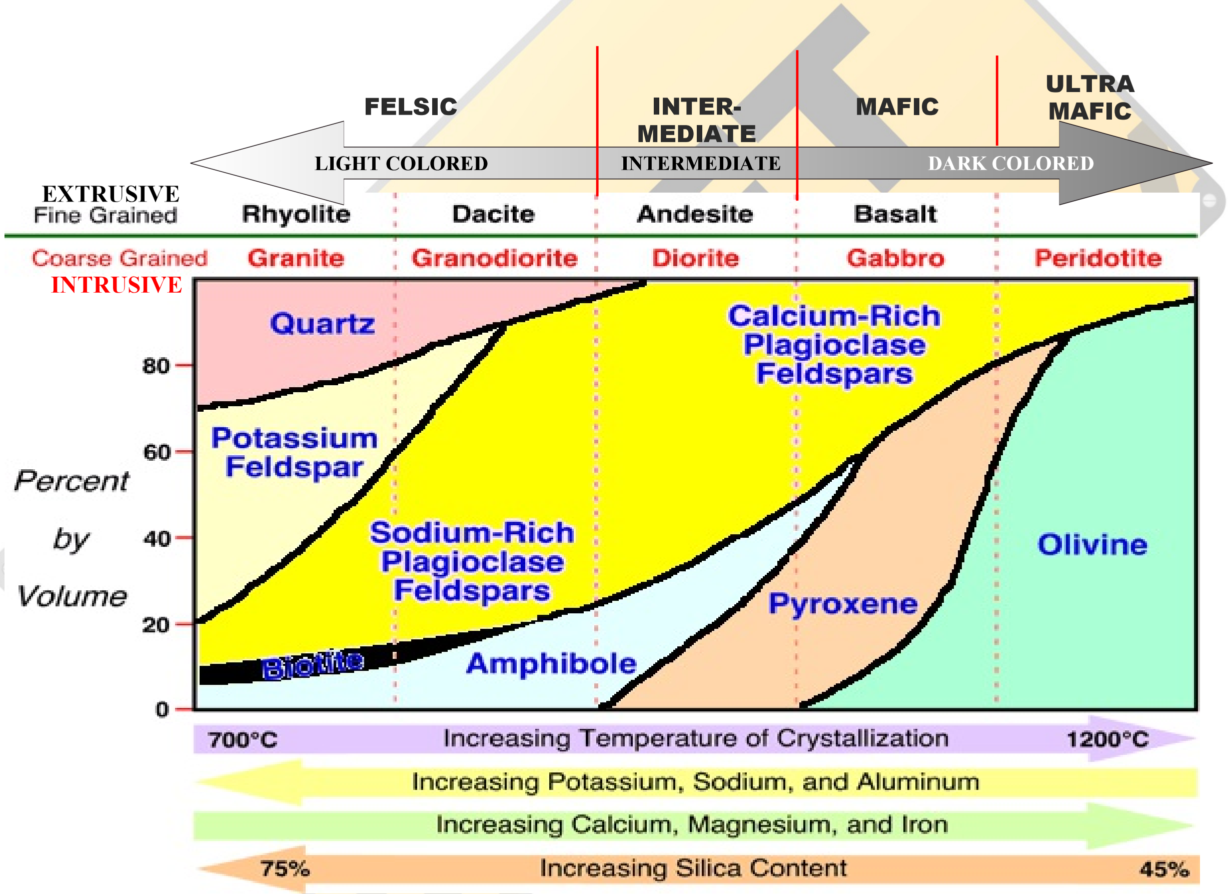

Figure 7.1- Classification chart of igneous rocks

The classification chart of the igneous rocks (figure 7.1) includes the more important igneous rock types. It is based on texture, or grain size of the included minerals (coarse grained = phaneritic = intrusive; fine grained = aphanitic = extrusive), color (light = felsic; dark = mafic), and mineral composition. If the visible minerals can be identified in the field, the naming of the rock is a relatively simple matter.

Obviously, more refined methods are used to give an exact classification name to a rock type. This involves mounting a rock slice on a glass slide, grinding it to a very thin thickness until it is transparent, then studying the minerals and texture under a petrographic microscope, identifying them, and estimating their percentage, eventually arriving at a classification of the rock. New instrumental methods now permit rapid determination of the chemical composition of a rock at a reasonable cost. The chemical information is also used to classify the rock. Yet, even with these sophisticated methods, the field classification of an igneous rock can, if carefully done, yield an accurate mineral and chemical composition for the rock type and a specific name.

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Table 7.2 - Igneous Rock ID Table

Texture	**Light Color**		**Intermediate Color**	**Dark Color**
Fragmental	Pumice, Tuff and Tuff Breccia			
Glassy & Phenocrysts	Obsidian – Tachylyte Vitrophyres			
Aphanitic Rare Pheno-crysts in very fine-grained matrix	Trachyte	Rhyolite	Andesite	Basalt (May be smooth vesicular, ropy amygdaloidal or breccia)
Porphyritic Fine-grained Matrix	Trachyte Porphyry	Rhyolite Porphyry	Andesite Porphyry	Basalt Porphyry
Phaneritic somewhat equigranular	Syenite	Granite and Variants	Diorite	Gabbro Pyroxinite Hornblendite Peridotite Dunite Olivine-Pyroxene
	Pegmatite (Very Coarse-grained)			
	Essential Visible Minerals			
	Feldspar Some Dark Minerals Occ. Nepheline	Quartz Feldspar Minor Dark Minerals	Feldspar (with striae) 20%-40% Dark Minerals	40%-70% Dark Minerals Biotite Hornblende Pyroxene Olivine Magnetite

A simple classification matrix is given in table 7.2. To classify an igneous rock first identify the visible minerals, note the color of the rock, refer to the general position of this color on the igneous rock ID table, identify the column for the visible mineral composition, then find the row which best matches the texture of the rock. The chart block lists the rock name. As an example of the procedure, if the rock is light colored, has quartz and feldspar and minor biotite in a matrix of finer-grained or aphanitic material, and the phenocrysts comprise about 20 percent of the volume, the rock would be a rhyolite porphyry.

Another helpful aid in estimating percentages of minerals in a sample is the use of a percentage estimation chart as presented in figure 7.2.

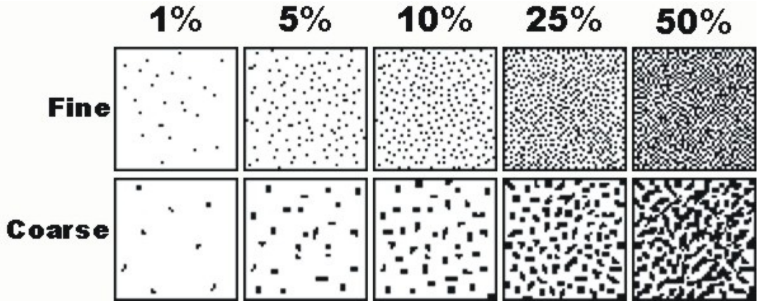

Figure 7.2- Percentage Estimation Chart

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DETAILED IGNEOUS ROCK DESCRIPTIONS

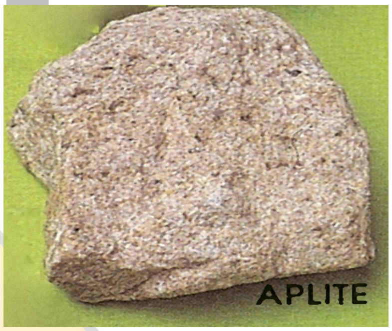
APLITE *SI-CONTENT:* > 69% *ROCK GROUP:* Felsic Intrusive Igneous Rock (Dikes, Light Color)

TEXTURE: Aphanitic., with distinctive sugary texture. Composition is mainly quartz. potassium feldspar and some plagioclase. ***DESCRIPTION:*** Aplites occur as dike-like bodies cutting granites and country rock. The dikes range from inches to feet in thickness and may extend to thousands of feet in length. Generally grayish to pinkish in color, sometimes medium red. ***COMPOSITION:*** Similar to granite pegmatite.

MAJOR MINERALS: Quartz, Potassium Feldspar (Orthoclase, Microcline, Albite), Mica.

MINOR MINERALS: Tourmaline. Plagioclase and others.

OCCASIONAL MINERALS: N/A



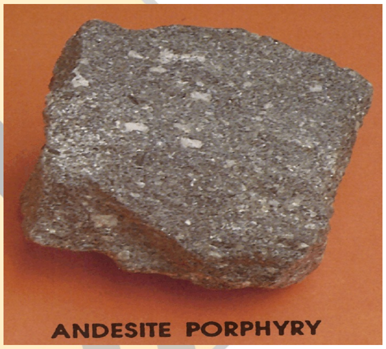
ANDESITE *SI-CONTENT:* ~ 60% *ROCK GROUP:* Intermediate Extrusive Igneous Rock (Intermediate Color)

TEXTURE: Aphanitic, porphyritic (common). ***DESCRIPTION:*** While phenocrysts are not abundant in this intermediate gray porphyritic rock, they are mainly of plagioclase. The mafic mineral content is relatively low and the matrix is dense of darker gray tone. The rock is part of a lava flow. ***COMPOSITION:*** The usual andesite contains more mafic minerals, usually hornblende with more or less biotite and/or augite. Calcium content slightly exceeds potassium and sodium content combined.

MAJOR MINERALS: Plagioclase (Anorthite < 59%), Biotite

MINOR MINERALS: Magnetite, Ilmenite, Quartz (<10%), Hornblende, Pyroxene

OCCASIONAL MINERALS: Orthoclase, Anorthoclase, Olivine



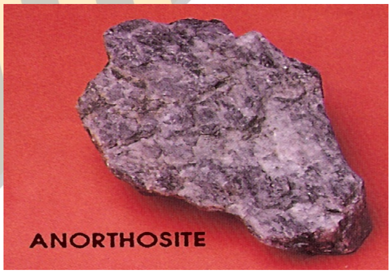
ANORTHOSITE *SI-CONTENT:* ~ 50% *ROCK GROUP:* Ultramafic Intrusive Igneous Rock (Dark Color)

TEXTURE: Phaneritic. Interlocking Plagioclase. ***DESCRIPTION:*** Intermediate gray. Ilmenite-Magnetite ore bodies are worldwide associated with Anorthosite. ***COMPOSITION:*** Mainly Plagioclase of Ca rich Anorthite and Labradorite.

MAJOR MINERALS: Plagioclase (Anorthite ~50%, Albite ~23%)

MINOR MINERALS: Diopside

OCCASIONAL MINERALS: Olivine, Apatite



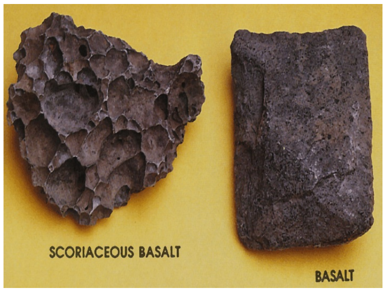
BASALT *SI-CONTENT:* ~ 50% *ROCK GROUP:* Mafic Extrusive Igneous Rock (Dark Color)

TEXTURE: Aphanitic, occasionally porphyritic. ***DESCRIPTION:*** A few ferromagnesian phenocrysts, mainly augite, in a dense, dark colored matrix. The increased density over the lighter colored igneous rocks is apparent. Vesiculation (gas bubble cavities) common due to accelerated escape of large quantities of gas during the rapid cooling of large lava flows (Scoriaceous Basalt). ***COMPOSITION:*** Normally basalt consists of dark gray plagioclase feldspar in a dark-colored matrix. Augite and olivine are commonly present, while hornblende and biotite are scarce. Some magnetite may be visible. Calcium oxide is about 10 percent, content of K and Na is less than 3 percent total. Weathering may produce Clay and Zeolites in vesicles.

MAJOR MINERALS: Plagioclase (Anorthite >50%), Pyroxene

MINOR MINERALS: Glass, Olivine, Magnetite, Apatite, rarely quartz

OCCASIONAL MINERALS: Leucite, Nepheline, Sodalite, Mica (Phlogopite)



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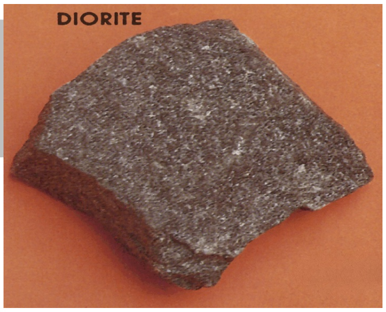
DIORITE *SI-CONTENT:* ~ 57% *ROCK GROUP:* Intermediate Intrusive Igneous Rock (Intermediate Color)

TEXTURE: Phaneritic (smaller grains). ***DESCRIPTION:*** Holocrystalline (all minerals are crystalline and visible to the unaided eye) often resulting in a "Salt & Pepper" appearance. Light to dark mineral content is about 65% / 35%. Composite color is usually dark grey. ***COMPOSITION:*** Hornblende is the dark mineral. Content of K_2O + Na_2O to CaO is about 5.5 / 6.7 % respectively.

MAJOR MINERALS: Plagioclase (Anorthite < 50%), Hornblende

MINOR MINERALS: Quartz (< 10%), Allanite (dark silicate containing Ce, Th, La, & Y), Ilmenite

OCCASIONAL MINERALS: Pyroxene, K-feldspar, Biotite



DUNITE *SI-CONTENT:* ~ 38% *ROCK GROUP:* Ultramafic Intrusive Igneous Rock (Dark Color)

TEXTURE: Phaneritic. Sugary grained. ***DESCRIPTION:*** Pale green to brown. ***COMPOSITION:*** Almost monomineralic consisting of Olivine (~67%).

MAJOR MINERALS: Olivine

MINOR MINERALS: Chromite

OCCASIONAL MINERALS: Apatite, Ilmenite

GABBRO *SI-CONTENT:* ~ 48% *ROCK GROUP:* Mafic Intrusive Igneous Rock (Dark Color)

TEXTURE: Phaneritic. Equigranular. ***DESCRIPTION:*** Dark gray to blackish brown color results from approximately 50% light and 50% dark mineral content. ***COMPOSITION:*** The lighter-colored minerals are mainly plagioclase feldspar, the dark minerals mainly pyroxene. Calcium oxide is about 11 percent, content of $Na_2O + K_2O$ is about 1/3 that of the calcium in the rock.

MAJOR MINERALS: Plagioclase (Anorthite >50%), Pyroxene

MINOR MINERALS: Olivine, Magnetite, Apatite, rarely quartz

OCCASIONAL MINERALS: Leucite, Nepheline, Sodalite, Mica (Phlogopite)

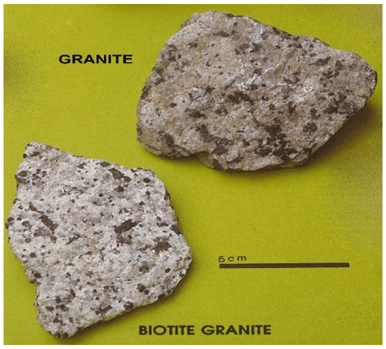
GRANITE *SI-CONTENT:* > 70% *ROCK GROUP:* Felsic Intrusive Igneous Rock (Light Color)

TEXTURE: Phaneritic. Equigranular. ***DESCRIPTION:*** Light to intermediate colors, generally grays to reds with a uniform mix of minerals. ***COMPOSITION:*** Quartz, Orthoclase feldspar, lesser Plagioclase, and even lesser ferromagnesian minerals. A *Biotite Granite* contains biotite. If black, elongate hornblende is visible, the granite is called a *Hornblende Granite*.

MAJOR MINERALS: Quartz, Potassium Feldspar (often pink)

MINOR MINERALS: Plagioclase, Mica, Hornblende

OCCASIONAL MINERALS: Magnetite, Ilmenite.



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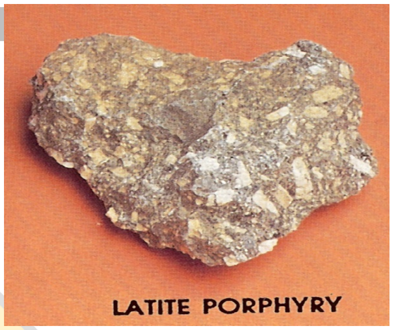
LATITE *SI-CONTENT:* ~ 56% *ROCK GROUP:* Intermediate Extrusive Igneous Rock (Intermediate Color)

TEXTURE: Porphyritic. Conspicuous two generation plagioclase phenocrysts (small & large) intermixed with orthoclase. ***DESCRIPTION:*** Dense gray matrix with phenocrysts. Rock is part of lava flow. ***COMPOSITION:*** K_2O / Na_2O about 8%, CaO about 6%.

MAJOR MINERALS: Sanidine, Plagioclase, Augite, Hornblende.

MINOR MINERALS: Magnetite, Ilmenite, Anorthoclase, Olivine, Feldspathoids.

OCCASIONAL MINERALS: N/A



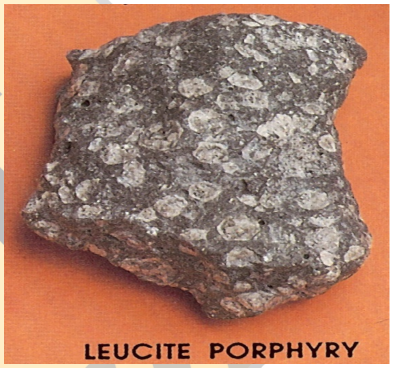
LEUCITE *SI-CONTENT:* ~ 58% *ROCK GROUP:* Intermediate Extrusive Igneous Rock (Intermediate Color)

TEXTURE: Porphyritic. Distinctive dodecahedral form of Leucite phenocrysts. ***DESCRIPTION:*** Growth of the large phenocrysts must have taken place in a slow-cooling environment with the early leucite growing from nucleation centers and achieving an almost perfect crystal shape. The isometric form is readily recognized. The magma then moved upward to the surface environment where the dense matrix formed from much more rapid cooling. If orthoclase is present in greater amounts, Leucite rocks become Phonolites. ***COMPOSITION:*** Ferromagnesian minerals are rare. K_2O about 6%.

MAJOR MINERALS: Leucite

MINOR MINERALS: Nepheline, Sodalite

OCCASIONAL MINERALS: Magnetite



MONZONITE *SI-CONTENT:* ~ 56% *ROCK GROUP:* Intermediate Intrusive Igneous Rock (Intermediate Color)

TEXTURE: Phaneritic. Equigranular. ***DESCRIPTION:*** Intermediate gray tones. Density about 2.82. ***COMPOSITION:*** Equal abundance of both orthoclase and plagioclase and about 15% of ferromagnesian minerals (Hornblende). Quartz is not visible. Oxides of Ca, K, Na ~6.4 to 6.5%.

MAJOR MINERALS: Orthoclase, Labradorite (Plagioclase), Pyroxene, Hornblende.

MINOR MINERALS: Quartz, Biotite, Titanite, Magnetite, Ilmenite.

OCCASIONAL MINERALS: Olivine, Nepheline.

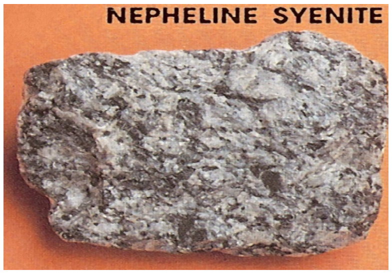
NEPHELINE SYENITE *SI-CONTENT:* ~ 55% *ROCK GROUP:* Intermediate Intrusive Igneous Rock (Intermediate Color)

TEXTURE: Phaneritic. Equigranular. ***DESCRIPTION:*** Light to intermediate gray. Developed in a slow cooling environment, various minerals were able to grow with mutually interlocking boundaries.. ***COMPOSITION:*** Relatively high in K and Na oxides ~14% and 20% Al_2O_3. Intrusive equivalent of Phonolite.

MAJOR MINERALS: Orthoclase, Plagioclase, Amphibole, Nepheline.

MINOR MINERALS: Pyroxene, Quartz, Biotite, Magnetite, Ilmenite.

OCCASIONAL MINERALS: Olivine, Corundum.



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OBSIDIAN *SI-CONTENT:* variable *ROCK GROUP:* Volcanic Glass, Extrusive Igneous Rock (usually Dark Color)

TEXTURE: Volcanic glass with bright vitreous luster and very distinctive conchoidal fracture. ***DESCRIPTION:*** Formed as the result of sudden cooling or quenching of the original hot magma. Time for solidification was insufficient to permit crystal growth and glass is the final product. Reddish coloration is probably the result of hematitic iron content. Black obsidian may relate to dispersed, very fine-grained magnetite. ***COMPOSITION:*** Water content of obsidian is usually < 1%. Specific gravity ranges from 2.37 for rhyolitic obsidian to 2.77 for basaltic glass. Most natural glasses have a composition approaching the quartz-alkali feldspar cotectic.

MAJOR MINERALS: N/A

MINOR MINERALS: N/A

OCCASIONAL MINERALS: N/A

PEGMATITE *SI-CONTENT:* variable *ROCK GROUP:* usually Felsic Intrusive Igneous Rock (Light Color)

TEXTURE: Phaneritic. Very large crystals (inches to feet). ***DESCRIPTION:*** Usually light colors. Very large grain size, most commonly of granitic composition, but also known for gabbro. Simple pegmatites have a granite composition. The same is true for Layered pegmatites. Zoned pegmatites exhibit a concentric pattern with the core carrying unique assemblages of rare minerals. ***COMPOSITION:*** Large crystals of spodumene (Li bearing pyroxene), Amblygonite (Li bearing phosphate), Columbite-Tantalite (Ta, Nb rare earth oxide), and other rare earth minerals. Major minerals of pegmatites can reach incredible crystal dimensions of several feet.

MAJOR MINERALS: Quartz, Potassium Feldspar, Mica (usually Muscovite)

MINOR MINERALS: Tourmaline, Beryl, Zircon. Apatite, and other rare minerals.

OCCASIONAL MINERALS: N/A

PHONOLITE *SI-CONTENT:* ~ 52 -63% *ROCK GROUP:* Intermediate Extrusive Igneous Rock (Intermediate Color)

TEXTURE: Aphanitic, porphyritic (common). ***DESCRIPTION:*** Light to medium gray, dense matrix. The rock has an unusual ring to it when struck by a hammer. Occurs as lava flows and dikes. ***COMPOSITION:*** Occasional phenocrysts of feldspar and feldspathoids (nepheline, leucite, and others) and minor ferromagnesian minerals (usually hornblende). K + Na oxide content is about 14% and Al near 21 %.

MAJOR MINERALS: Sanidine, Nepheline, Acmite and Hornblende (Riebeckite or Artvedsonite)

MINOR MINERALS: Anorthoclase, Albite, Augite, Apatite, Titanite, Magnetite, Sodalite, Analcite, Zeolites and Melanite garnets.

OCCASIONAL MINERALS: N/A

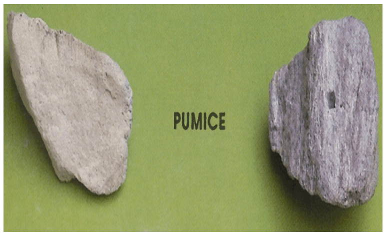
PUMICE *SI-CONTENT:* variable *ROCK GROUP:* usually Felsic Extrusive Igneous Rock (Light Color)

TEXTURE: Volcanic frothy glass. Under hand lens fine fibers surrounded by a multitude of open pores and vesicles visible. ***DESCRIPTION:*** Unique texture results from rapid escape of gas in a surface or near surface environment with extreme vesiculation and rapid cooling of the magma during the degassing period. The high pore space imparts a low specific gravity to the volume of rock which is known to float on water. Light colored, generally white to grayish brown. ***COMPOSITION:*** The oxide composition approaches that of rhyolite, but some is basaltic.

MAJOR MINERALS: N/A

MINOR MINERALS: N/A

OCCASIONAL MINERALS: N/A



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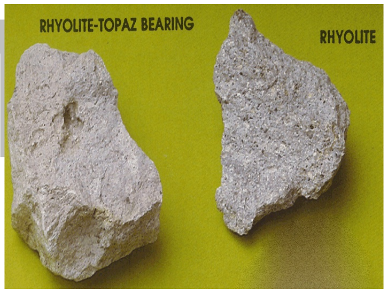
RHYOLITE *SI-CONTENT:* > 72% *ROCK GROUP:* Felsic Extrusive Igneous Rock (Light Color)

TEXTURE: Aphanitic. Sometimes with distinctive, conspicuous banding or layering, often intricately folded as a result of planar development in the cooling of the original lava. Gas may develop pore space within some rhyolites, which may contain minerals (Topaz common) ***DESCRIPTION:*** Dense matrix with minor phenocrysts of feldspar and quartz. Generally light in color in shades of gray, sometimes red. ***COMPOSITION:*** Next to SiO_2, ~13.5% Al_2O_3, $Na_2O + K_2O$ about 8%, 1% CaO.

MAJOR MINERALS: Quartz, Potassium Feldspar (Sanidine)

MINOR MINERALS: Glass, Biotite, Plagioclase (Albite), some Magnetite, Ilmenite

OCCASIONAL MINERALS: Topaz, Amphibole, Diopside.



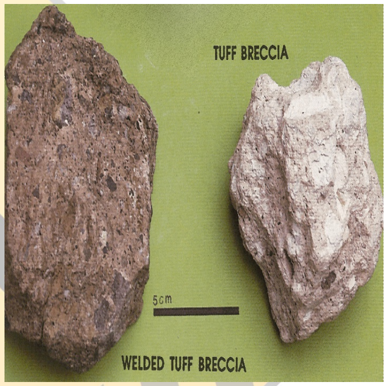
TUFF BRECCIA *SI-CONTENT:* variable *ROCK GROUP:* Volcanic Rock Fragments (variable)

TEXTURE: Angular lithic or rock fragments included in a dense matrix with broken phenocrysts. ***DESCRIPTION:*** Form from explosive ejecta from a main venting area in which solids, semi-molten magma, gas, and water are violently erupted as dense pyroclastic flows with velocities of 100 mph. Almost instantaneous geologic time areas extending for thousands of square miles are covered by these "glowing cloud" masses. Fragment size may range to inches in diameter and may be a function of distance from the explosive center. Coloration is light gray to moderately dark red. ***COMPOSITION:*** *Welded Tuff Breccia*: Solid, dense matrix of rock with volcanic fragments literally welded in place by rapid cooling. *Tuff Breccias:* resemble welded tuff breccias, but lack the dense, more solid matrix. Tuff breccias are quite friable, may powder and erode easily, and have a lower specific gravity.

MAJOR MINERALS: N/A

MINOR MINERALS: N/A

OCCASIONAL MINERALS: N/A



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8 SEDIMENTARY ROCKS

Sedimentary rocks are derived from the weathering and erosion of pre-existent rocks whether igneous, metamorphic, or sedimentary, or from a combination of these. Such an origin suggests a broad spectrum of composition for the sedimentary rock class. Rocks formed by sedimentary processes develop at or near the surface of the earth. Two major sub-classes are recognized: (1) clastic-type sedimentary rocks and (2) chemical precipitates. A less abundant, though economically important organic plant accumulation such as peat, lignite and coal also occurs, as well as fossil animal deposits such as those made up of skeletons of invertebrates. The accumulation of sediments, chemical precipitates, organic deposits and evaporites by deposition is but a first stage of many which result in the final product we call a sedimentary rock. Burial, compaction, dehydration, induration, diagenetic changes and other processes are involved before a sediment can become a rock we can heft and classify.

8.1 CLASTIC SEDIMENTARY ROCKS

The clastic sedimentary rocks are derived from the erosion or weathering away of pre-existent rocks. The small and large derived particles are transported by water, wind or ice to the site of deposition. The type of transporting agent to a large extent determines the texture and structure of the rocks formed. Ice is a very poor sorter of rock materials. As it melts and retreats it leaves behind aggregates of fine and coarse particle sizes all heterogeneously arranged. If this sedimentary material is buried, consolidated, and indurated, the rock product formed is called tillite.

Wind is at the other extreme and is a very good sorting agent. Only the finer particles are transported and deposited by this medium and even these are well- sorted by the wind currents. Fine-grained sands, silts and even finer clay particles are part of the sediment load which is deposited. The burial, compaction and induration of these fine-grained sediments produces sandstones, siltstones. and shales. The massive cliffs of Zion National Park, Utah are a striking example of wind deposited sandstones with their conspicuous internal cross-layered structures.

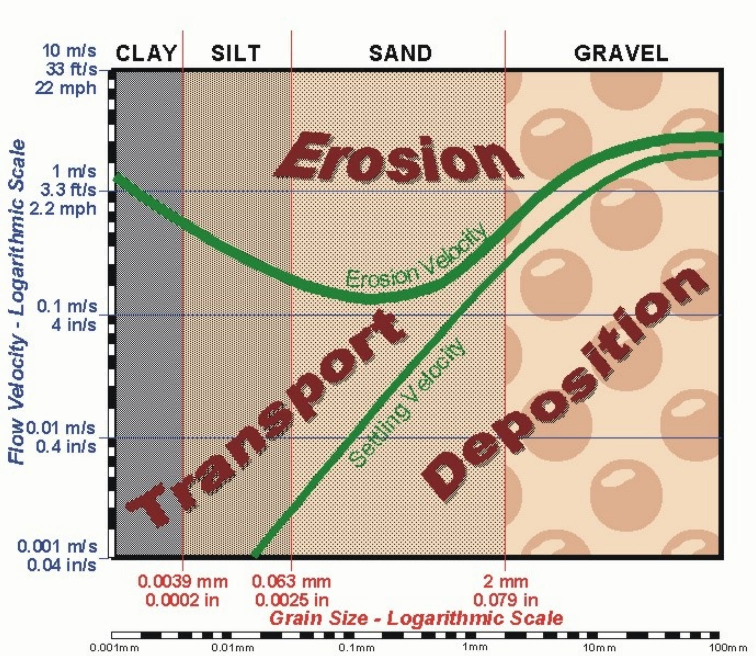

Figure 8.1- Sediment transport velocities in relation to erosion & deposition

Stream waters, one of the main erosional and transporting media, mechanically erode particles of rock, carry them in suspension and as bed load, and redeposit them in the same or a different environment. As an example, it has been estimated that 500,000,000 tons of dissolved and clastic material are carried into the Gulf of Mexico by the mighty Mississippi River each year. A stream is potentially able to carry in suspension and traction an amount of sediment equivalent to the energy available to stream flow. The greater the velocity of the stream. the larger the amount of sediment transported and also the larger the particle size carried and later deposited.

Lake waters and ocean basins are also very important agents in the formation of sedimentary rocks. Within such basins large quantities of fine-grained sediments can be deposited through streams carrying the sediments into the lakes and these waters dispersing the particles across the lake floors. Similarly in the oceans, sediments from the land through stream transport and also from wave erosion and current transport result in extensive clastic deposits.

8.2 CHEMICAL SEDIMENTARY ROCKS

Large quantities of the chemical weathering products of rocks are transported in solution in water. These are not mechanical or visible particles, but dissolved material in the water and invisible to the eye. They are part of the surface waters of the earth. Material in solution in stream waters is estimated at about 30% of the stream load. If this applies to the Mississippi River, about

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150.000,000 tons per year of dissolved matter is carried to the Gulf of Mexico each year. The total probably exceeds this amount. To gain a better grasp of the quantity of this dissolved material in the rivers, it would be the equivalent of 1,500,000 freight cars carrying 100 tons each year or about 4,000 per day, or 166 per hour passing a set point near the mouth of the river and chemically adding this large quantity to the waters of the Gulf. It is obvious that the amount of materials in solution transported by the rivers of the world is prodigious. In both lacustrine and marine environments chemical precipitates can form. Witness the extensive marine limestone of the mountains and plains area and the great saline deposits of New Mexico, Kansas, Michigan, New York and elsewhere-the result of both extensive chemical and evaporative processes in ancient seaways.

The minerals calcite and dolomite are chemical precipitates of calcium and magnesium carbonates. These may result from direct chemical precipitation from water or be precipitated by organic processes in a marine environment, such as secretions by corals or clams. In some cases, large calcareous skeletons, such as shells, accumulate to form coquina, a type of sedimentary rock.

Chert which is a locally abundant siliceous rock may result from the accumulation of great quantities of siliceous shells of certain planktonic life forms or be a chemical precipitate from water.

A third, and locally economically important sub-class of chemical precipitates are the evaporites or salt deposits. These are the residues from the evaporation of sea or lake waters. They consist mainly of sodium chloride, or salt, gypsum and anhydrite. Much rarer are the trona, sodium bicarbonate deposits, and the borates which include borax, ulexite and other borax-bearing bodies.

Peat and lignite are formed from the compaction and chemical changes on organic accumulations of plant materials. These are plant fragments which have been uniquely preserved from erosion and decay in special sedimentary environments such as swamps, some lagoons, and even some unusual stream areas.

8.3 CLASSIFICATION & IDENTIFICATION OF SEDIMENTARY ROCKS

Clastic sedimentary rocks are classified according to particle size into gravels and breccias of greater than 2 mm in diameter, sand 2 mm down to 1/16 mm, silts 1/16 - 1/256 mm, and as clay or mud for those which are less than 1/ 256 mm in diameter. The particle size also suggests an energy environment for the sedimentary deposits. Sorting and roundness of grains are also helpful in deciphering the transport history and the environment of deposition of grains (see figure 8.2).

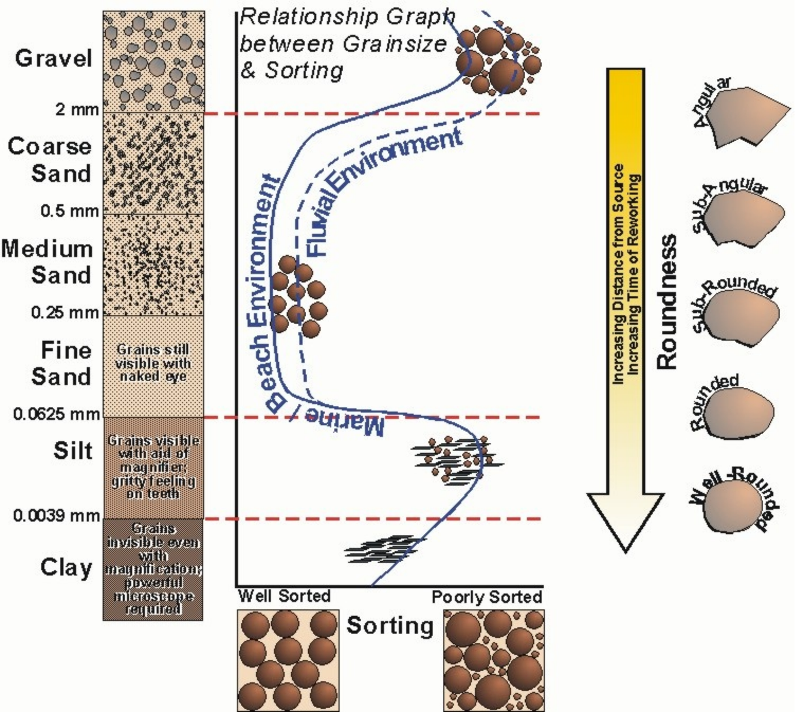

Figure 8.2 - Grainsize, rounding and sorting chart for clastic sediments (from Jones & Jones, 2003)

In short, the following transport and depositional assumption usually apply:

Grain size:
Small grains - low velocity, quiet depositional environment
Large grains - high velocity, high energy depositional environment

Rounding:
Angular - short transport
Well-rounded - long transport

Sorting:
Poorly sorted - fluctuating transport energies, usually short transport
Well sorted - uniform transport energies, usually longer transport

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Organic and chemical sedimentary rocks do not lend themselves to the ready classification of the clastic sub-class. In many classifications the type of origin is often used.

In order to distinguish between clastic and chemical sedimentary rocks, the following approximate flow chart (figure 8.3) might be helpful.

Once clastic or chemical / organic sedimentary rocks are distinguished, classification table 3.1 can be used to identify sedimentary rock types. It is based on texture, cement, and composition.

1. Step: Identify Clastic or Chemical / Organic rock (figure8.3)

2. Step: Identify *"Main Constituents"* in the rock (table 8.1)

3. Step: Identify the *"Texture"* of the rock (figure 8.2 & table 8.1)

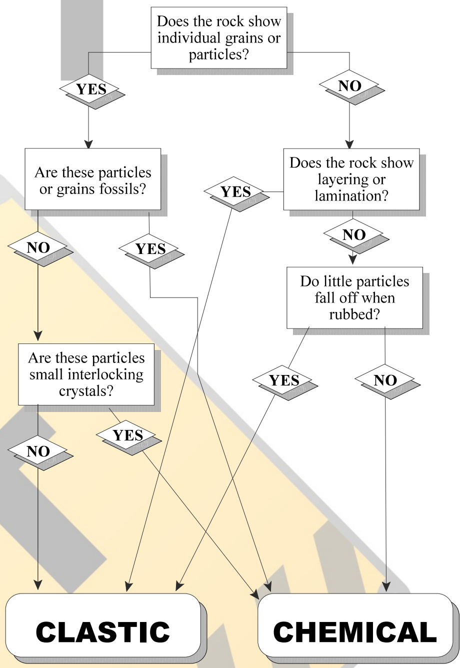

Figure 8.3 - Clastic / Chemical Sedimentary Rock ID flowchart

Table 8.1 - Classification of common sedimentary rocks

TEXTURE Size & Structure of grains & particles composing the rock	**CEMENT** Composition of material holding grains together	**CLASTIC ROCKS**			**CHEMICAL ROCKS**				**ORGANIC ROCKS**	
		MAIN CONSTITUENTS			MAIN CONSTITUENTS				MAIN CONSTITUENTS	
		Quartz &/or Chert	Quartz;, Chert, Mica, Rock Fragments	Quartz & Feldspars Some Clay	Silica gel	Calcite or Dolomite	Gypsum	Halite	Organic Materials	Carbon
CRYSTALLINE	N/A					Crystalline Limestone/ Dolomite	Selenite	Rock salt/ Halite		
COARSE Grain size > 2 mm	Silica	Quartz/Chert Conglomerate	Breccia/Fan Conglomerate	Breccia/Fan Conglomerate						
	Calcite & Other Carbonates	Quartz/Chert Conglomerate	Breccia/Fan Conglomerate	Breccia/Fan Conglomerate						
	None	Gravel	Tillite/ Gravel	Tillite/ Gravel			Selenite/ Gypsum	Rock salt/ Halite		
MEDIUM Grain or Crystal 2 mm - 1/16 mm	Silica	Clean Sandstone	Argemite/ Graywacke	Arkose Sandstone		Siliceous Limestone/Dol.				
	Calcite & Other Carbonates	Calc. or Dolo. Sandstone	Argemite/ Graywacke	Cal. or Dolo. Arkose Sandstone		Dolomite/ Limestone				
	None	Clean Sand-stone/Sand	Argemite/Gray-wacke/Sand	Arkose Sand-stone/Sand		Dolomite/ Limestone	Selenite/ Gypsum/ Anhydrite	Rock salt/ Halite		
FINE & VERY FINE Grain or Crystal < 1/16 mm	Silica	Siltstone/Claystone			Opal/ Chert	Porcellanite/ Silic. Limestone				
	Calcite & Other Carbonates	Calcareous or Dolomitic/Siltstone/ Claystone/Marly Siltstone			Cal. or Dol. Chert	Micrite/ Limestone/Dol.				
	None	Siltstone/Claystone/Mud			Opal/ Chert	Micrite/Fine-grain Limestone	Anhydrite	Rock salt/ Halite		Cannel coal
Open Fibrous Fossil Remains	N/A					Fossiliferous Limestone COQUINA			PEAT COQINA (with carbonate cement)	
Dense	N/A					(see fine & very fine grain or crystal size)	Anhydrite	Rock salt/ Halite		COAL: Lignite (compact) Anthracite (very dense)

8.4 ENVIRONMENT OF DEPOSITION

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In the investigation of the earth's history, sedimentary rocks are the ultimate story tellers. By being observant, one cannot only estimate under what prehistoric conditions the rock was deposited, it is also possible to tell climate characteristics, distance from the source rock, and many other useful features. If one finds a sandstone containing clam shell fossils, it is easy to surmise that we are looking at an ancient beach. A limestone full of corals indicates warm, tropical shallow oceans. Coal seams point toward swamps with massive amounts of dead plant detritus. Sometimes features in the rock unit itself pinpoints the environment. Ripple marks in sandstone point to flowing or agitated water. In fact, even ancient stream directions maybe derived from such keen observations. The following figure 8.4 might be helpful in deciphering the story behind the rocks.

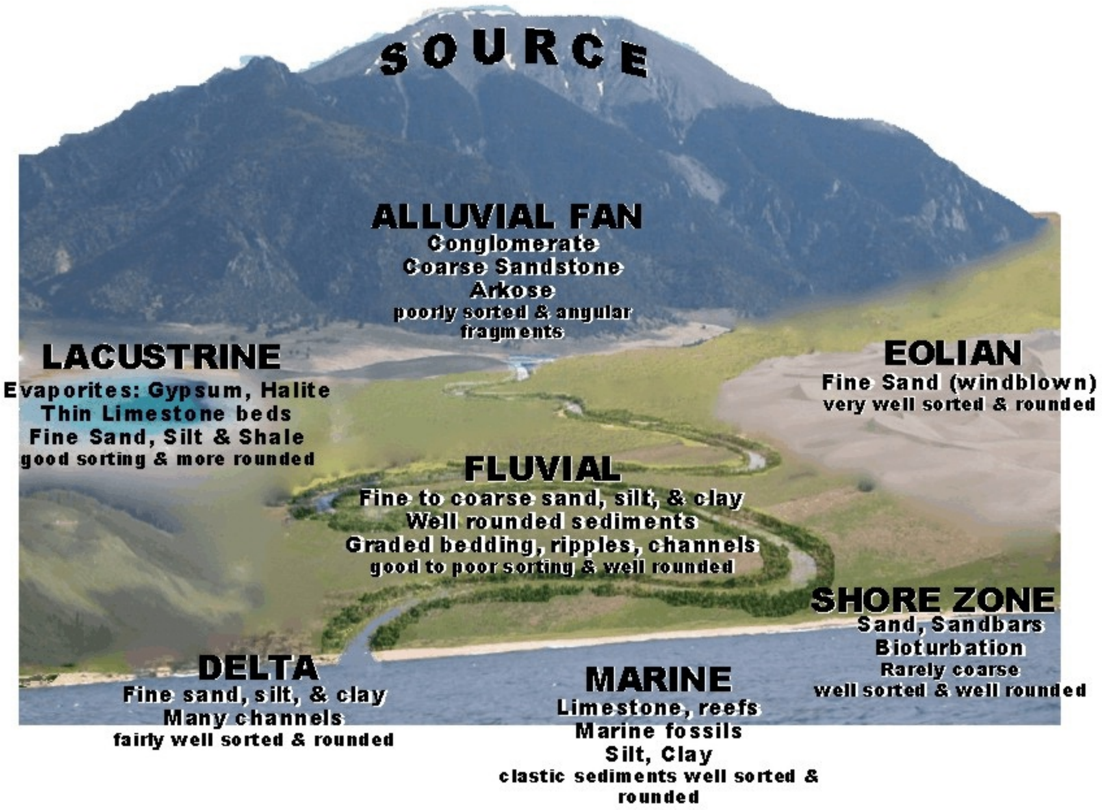

Figure 8.4 - Common depositional environments with simplified sedimentary rock types & features

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DETAILED SEDIMENTARY ROCK DESCRIPTIONS

CHERT

ROCK GROUP: Chemical/Biological Sedimentary Rock

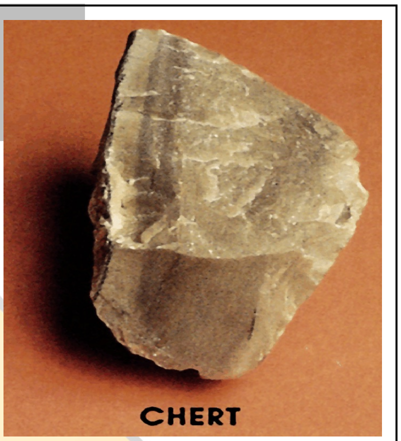
TEXTURE: Brownish gray, slight1y banded, crypto crystalline silica. Hardness of 7. conchoidal to splintery fracture.

ENVIRONMENT OF DEPOSITION:.As bands, beds, layers or nodules in other sedimentary units. See Formation History.

FORMATION HISTORY: Theories proposed for the origin of cherts include: 1) chemical sediment precipitate, 2) volcanic activity and silica release to ocean waters and later deposition, 3) normal sediment from activity of silica-secreting organisms and precipitation as a sediment.

RESISTANCE TO EROSION: Usually very high. Would make impressive ledges, but usually too thin.

COMPOSITION: Made entirely of silica (Quartz). Will scratch glass. Often used as arrow head material by aborigines. May also spark when struck together (flint).



COAL

ROCK GROUP: Chemical/Biological Sedimentary Rock

TEXTURE: Various textures from the fibrous organic matter of the pre-coal stage (peat) to the rich, black coal with vitreous luster and conchoidal fracture.

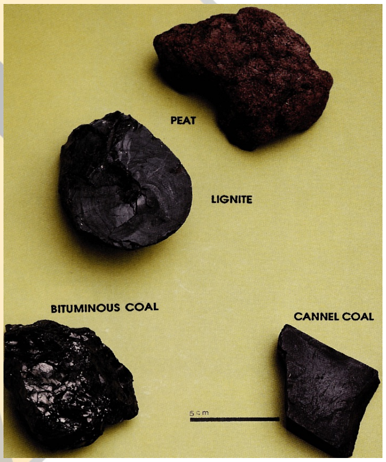
Peat - fibrous plant material (peat moss). Pre-coal stage.
Lignite - brown coal. Retains organic materials, lacks porosity of peat.
Bituminous - Black, glossy or dull, finely granular to homogenous

ENVIRONMENT OF DEPOSITION:.Swamps, tropical swamps, peat bogs.

FORMATION HISTORY: Coal formation begins with peat, which is compacted into lignite and upon deeper burial processes into bituminous coal and eventually the high quality anthracite, also called king of coals.

RESISTANCE TO EROSION: Not very high. Coal is soft, but it is astonishingly resistant to chemical erosion.

COMPOSITION: Made of organic plant materials which are changed into carbon over time. Bituminous coal may include small nodules of yellow sulfur. All coals are known to contain abundant plant fossils.



CONGLOMERATE

ROCK GROUP: Clastic Sedimentary Rock

TEXTURE: Coarse grained, poorly sorted, rounded particles in matrix of finer grains.

ENVIRONMENT OF DEPOSITION: Close to source. Alluvial fans. Mountain streams of youthful geomorphologic stage. Layers in river sediments during high energy transport events.

TRANSPORT HISTORY: Rounding of grains suggests transport in water.

RESISTANCE TO EROSION: Usually high when cemented. Ledge former common.

COMPOSITION: Larger particles can be any rock type (sedimentary, igneous, metamorphic) of some resilience. Finer material often sand or silt. Arkose common.

DOLOMITE *ROCK GROUP:* Chemical/Biological Sedimentary Rock

TEXTURE: Finely crystalline, dull luster. This rock looks like limestone but is a calcium / magnesium carbonate and will not react with acid unless powdered or hot acid is used. It is very often mistaken for limestone and even trained geologists have been fooled, mapping whole dolomite rock units as limestone, because a quick and simple acid test was not applied.

ENVIRONMENT OF DEPOSITION:.Dolomite is believed to replace Limestone over time through secondary alteration. Some evidence suggests that primary dolomite may occur from biochemical activity. Dolomite is very common. The most notable occurrence is in the Alps where whole mountain ranges (The Austrian / Italian Dolomites) are comprised of the carbonate.

FORMATION HISTORY: Result of the dolomitization of limestone. The process may occur during or soon after sedimentation or later after rock is completely solidified. May also result from hydrothermal (hot water) alteration near plutonic igneous rocks.

RESISTANCE TO EROSION: Usually high. Ledge former.

COMPOSITION: Made of dolomite - $(Ca,Mg)CO_3$. Reaction with acid is almost imperceptible. Will effervesce when dolomite is powdered or hot acid is applied.

GYPSUM *ROCK GROUP:* Chemical/Biological Sedimentary Rock

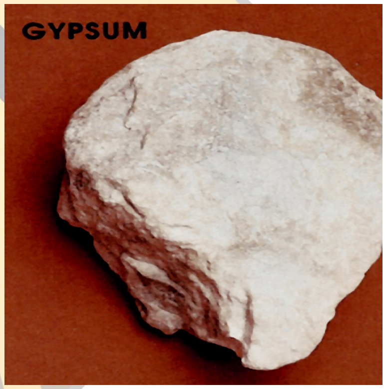
TEXTURE: Usually white to gray, finely crystalline. May show fine to coarse-layering, sometimes intricately folded. Some texture and properties as the mineral gypsum.

ENVIRONMENT OF DEPOSITION:.Typical evaporite deposit indicating hot & dry paleo-conditions. Forms sedimentary beds of wide areal extent. Well-layered, often thin to massive-bedded with thicknesses of up to 100 feet, but generally much less.

FORMATION HISTORY: Water from mountain run-off may collect in salt flats, pans, or shallow flood areas in hot climates where it quickly evaporates. If the concentration of calcium and sulfate ions in water is high enough, gypsum will precipitate.

RESISTANCE TO EROSION: Very soft. Erodes easily.

COMPOSITION: Made of the mineral gypsum. Very soft, can be scratched by fingernail (H:2). Finds extensive use in the building trade, being a major component of plaster and sheet rock.



LIMESTONE *ROCK GROUP:* Chemical/Biological Sedimentary Rock

TEXTURE: Dense to grainy to crystalline to fossil bearing.

Crystalline Limestone - medium to coarse-grained, tanish cream, uniform
Coquina - clastic limestone made entirely of mainly shells.
Fossiliferous Limestone - contains abundant fossils &/or fossil fragments
Lithographic Limestone - compact, very fine & even-grained, tanish cream
Oolitic Limestone - comprised of small, somewhat spherical particles
Travertine - hot spring deposit, often banded

ENVIRONMENT OF DEPOSITION: Often marine deposits within the Limestone facies (e.g. coral reefs), but can also form in fresh water (e.g. stromatolitic), or in hot springs (travertine) or evaporite deposit.

FORMATION HISTORY: Commonly secreted by biological organisms (shells, clams) & microorganisms. Chemical formation occurs around mineral springs, in evaporite settings or for oolitic limestone.

RESISTANCE TO EROSION: Usually high in dry climates, where it can form ledges. Erodes easily in wet areas or environments exposed to acid. Cave formation very common.

COMPOSITION: Made predominantly of calcite or aragonite. Often intermixed with some dolomite. Limestones will effervesce with acid.

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MUDSTONE, CLAYSTONE, SHALE

ROCK GROUP: Clastic Sedimentary Rock

TEXTURE: Grains are invisible to the naked eye or magnifying glass and are smaller than 1/256 mm.

Shale - shows laminations
Mudstone or Claystone - massive

ENVIRONMENT OF DEPOSITION: Very low energy environment, such as quiet flood plains, lake bottoms or deep oceans.

TRANSPORT HISTORY: Usually transported in water. Deep marine sediments (red clays) are windblown from continents into the oceans.

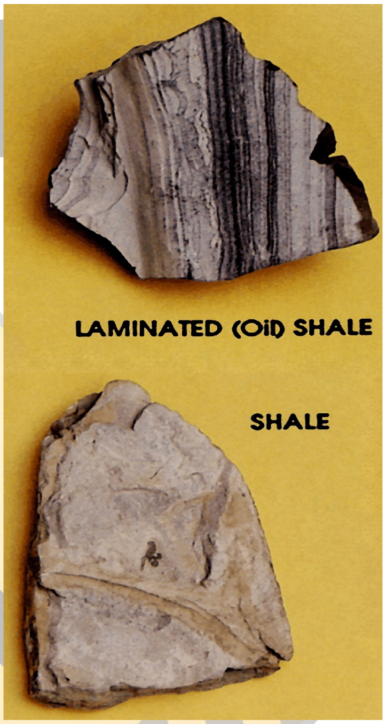
RESISTANCE TO EROSION: Very low. Forms extensive valleys or slopes.

COMPOSITION: Impossible to determine without instrumentation. The average mudstone contains about 30 % quartz and slightly more clay minerals. The other third is made up of other minerals including carbonates, feldspars, iron oxides and organic matter. Generalizations about mudstone composition can be made by simple tests or observations:

Black shale - high carbon content, frequently in the form of kerogen (hydrocarbons), thus often called "oil" shale. Much research has been done to effect an economic recovery of oil from these rocks, of which the higher content shales carry as much as 25-100 gallons / ton.

"Popcorn" weathering - a popcorn like appearance of the weathered clay or mudstone indicates the presence of the mineral bentonite, a swelling clay. Bentonite forms from the decomposition of volcanic ash.

Marl - Claystone or shale that contains calcite, hence it will effervesce with acid.



SANDSTONE

ROCK GROUP: Clastic Sedimentary Rock

TEXTURE: Grains are sand size, ranging from 1/16 mm for very fine sand to 1 mm for coarse sand.

ENVIRONMENT OF DEPOSITION: Varied. Ancient beaches, sand dunes, river channels and sandbars. Bedding, layering, or varving may occur.

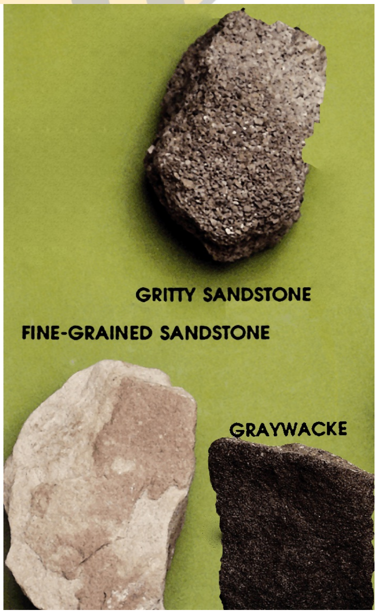
TRANSPORT HISTORY: Usually transported in water. Very well sorted sandstone often indicative of aeolian (wind) environment.

RESISTANCE TO EROSION: Usually high when cemented. May form ledge or hogbacks.

COMPOSITION: Varied. Common sandstone variations are named as:

Arenite - clean sandstone, predominantly quartz, with < 10% clay in matrix
Arkose - sandstone rich in K-spar
Litharenite - contains mainly fragments of preexisting rock.
Wacke - sandstone with 10% - 50% clay content

More than 60% of mineral grains in sandstone are usually quartz. Presence of accessory minerals augite, chromite, ilmenite, or topaz grains are indicative of igneous source rock. Grains of actinolite, andalusite, diopside, epidote, garnet, kyanite, rutile, sillimanite, staurolite or tremolite point to a metamorphic source. Greenish glauconite (mica) grains in sandstone are interpreted as having originated as marine fecal pellets.



SILTSTONE *Rock Group:* Clastic Sedimentary Rock

Texture: Grains are very small and range in size from 1/32 mm to 1/256 mm, about the size of very fine sandpaper, often outside the visible range. Siltstone will feel gritty on teeth, while shale or mudstone will be smooth (finer grains).

Environment of Deposition: Low energy environments such as lake beds, river deltas, flood plains, continental shelves or submarine basins. Fine scale bedding permits the use of "laminated" to be added to the rock name. Siltstones are part of the red-bed series of the Colorado Plateau.

Transport History: Transported in water.

Resistance to Erosion: Usually low, forms valleys or slopes. Occasional protruding siltstone ledges may occur when well cemented.

Composition: Difficult to determine visually because of small particle size. Instrumentation required. Intermixing with clay or mudstone customary. Quartz is a probable common component.

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9 METAMORPHIC ROCKS

Have you observed what happens to various substances when heat is applied or when other components are mixed with the original substance and heat is applied? The material may change in form. It may become more fluid and tend to flow, or it may combine with the added components and form a new solid substance. In nature rocks behave in a somewhat similar fashion. Given a certain rock type, the addition of heat and or pressure of increasing intensity over a period of time plus the presence or addition of fluids to this environment causes the rock to adjust to these new conditions. A new equilibrium is established through the formation of more stable minerals. How much change in form occurs is dependent on the magnitude of the stress and temperature and the reaction of the fluid components, plus the duration of the reaction time. Aqueous solutions are of utmost importance in the metamorphic process. ***"Dry" rocks hardly metamorphose!*** Simple rocks such as shale, a sedimentary rock mainly composed of clay minerals and some quartz, maybe radically changed. The end product is a metamorphic rock, often times completely different in appearance. It may become a phyllite, a rock much hardened and with a visible sheen of micaceous minerals developed within it. Under stress the original shale may be converted to a slate. Where once the shale broke along bedding planes, now the rock parts along cleavage planes, often times completely different than the still visible bedding planes of the shale. Under the microscope one can observe the orientation of minute mineral grain parallel to this new breaking of cleavage surface. With even greater heat and stress the original shale may be changed to a schist, a metamorphic rock with clearly visible foliation planes and platy minerals aligned parallel to the schistose structure. Under extreme heat and pressure all of the components of the rock may be reorganized, even flowage may occur, and dark and light minerals tend to segregate together to form a banded-rock called a gneiss.

As in the case of the origin of the deeply burled igneous rocks, we have no way of observing these changes as they occur. We must base our conclusions on field observations and the inferred conditions of temperature and pressure. Fortunately, abundant and well-controlled laboratory experiments have been performed which permit well-documented conclusions as to the effects of temperature, pressure and fluids on various rock materials and the end products produced. The development of the ultra-high pressure tetrahedral press by Tracy Hall and his associates has resulted in productive research at pressure over 1,000,000 PSI and at temperatures up to 1500 degrees Centigrade.

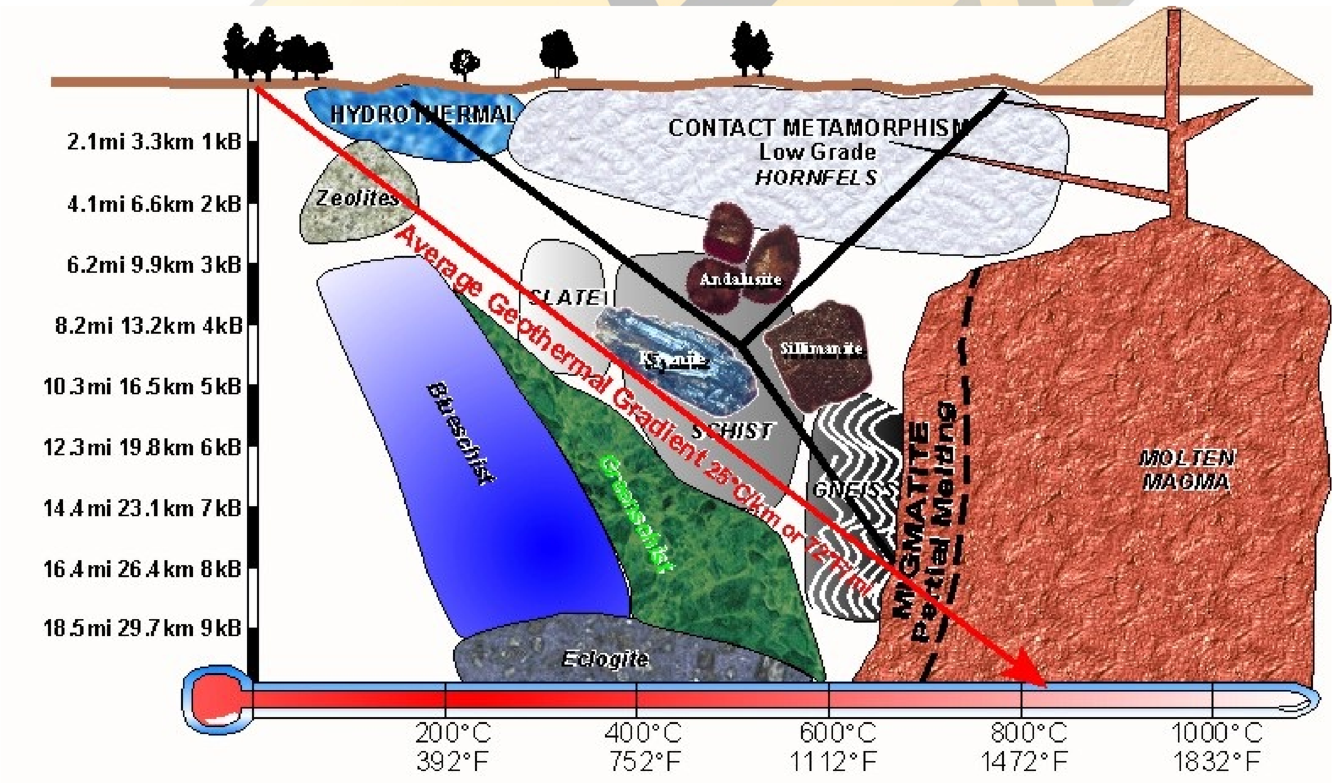

Figure 9.1 - Temperature / Pressure Diagram for metamorphic environments and most common rock types

Two stress environments for metamorphic rocks are recognized: 1) static or load metamorphism under elevated temperatures, and 2) dynamic stress environment with increased temperature. A third environment of metamorphism is that of increased temperature. The latter may act on the rock in both stress environments and locally only as heat near a cooling magmatic body. Figure 9.1 displays a temperature and pressure diagram with related metamorphic classes and rock types.

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With these factors operating in nature certain visible changes are wrought by metamorphism. The constituents of the rock may be completely reorganized to form a new more stable mineral assemblage. Grain size of the minerals is generally increased with a resultant coarser texture. The internal structure of the rock may change, and commonly a parallel arrangement of the constituent minerals may take place. In m any cases the overall composition of the rock may remain the same, while in others constituents are lost or others added. The terms non-additive and additive metamorphism have been given to these processes.

Because the metamorphic processes take place at high pressure and high temperature, many of the minerals which form are also common to igneous rocks. Exceptions are the silica deficient feldspathoids such as leucite, nepheline and sodalite. Table 9.1 includes those minerals common in metamorphic rocks.

Minerals within metamorphic rocks can also be used as a geothermometer showing the temperature of metamorphism as seen in figure 9.2. Thus formation temperatures can be bracketed and metamorphic conditions estimated.

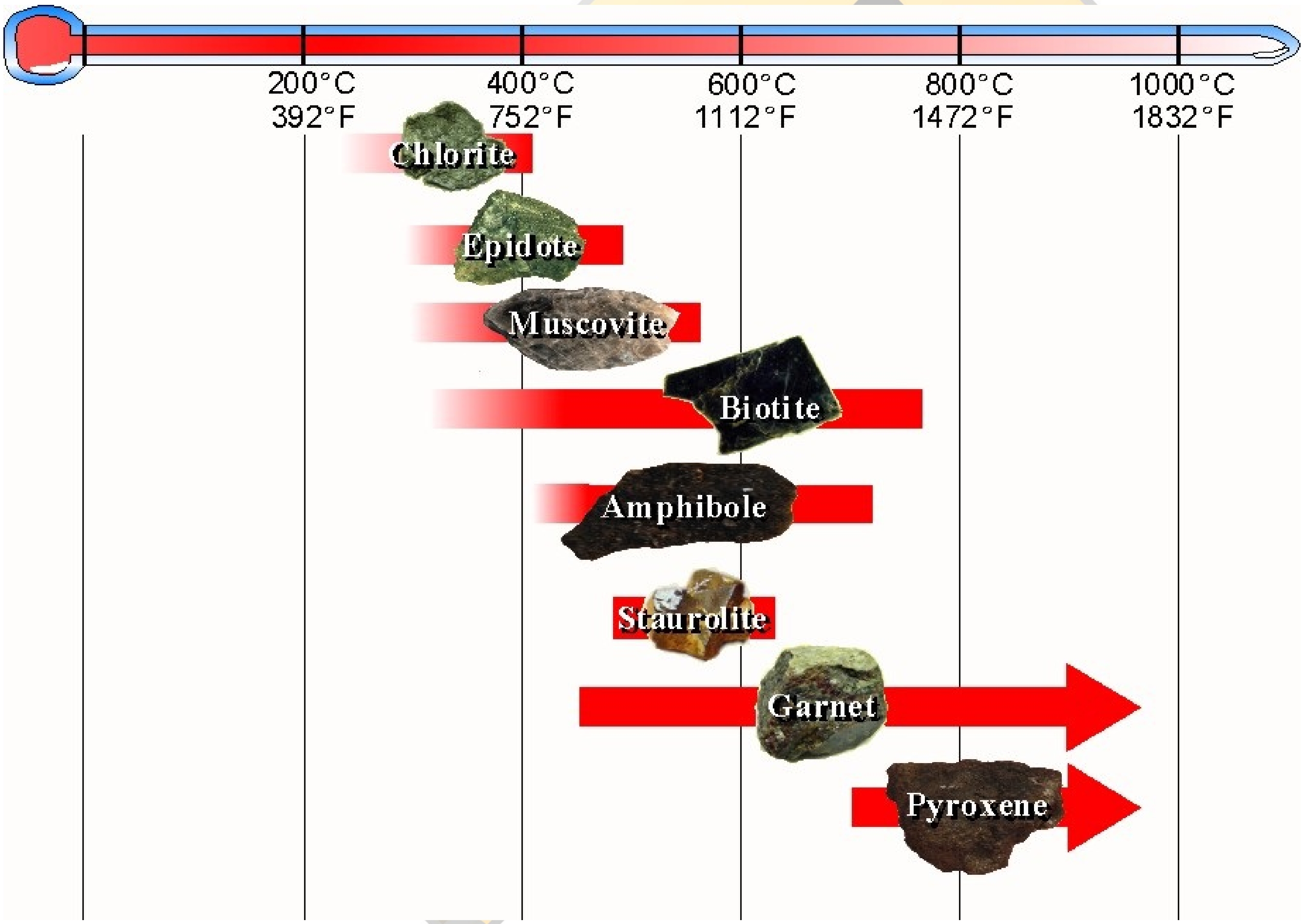

Figure 9.2 - Geothermometer minerals in metamorphic rocks.

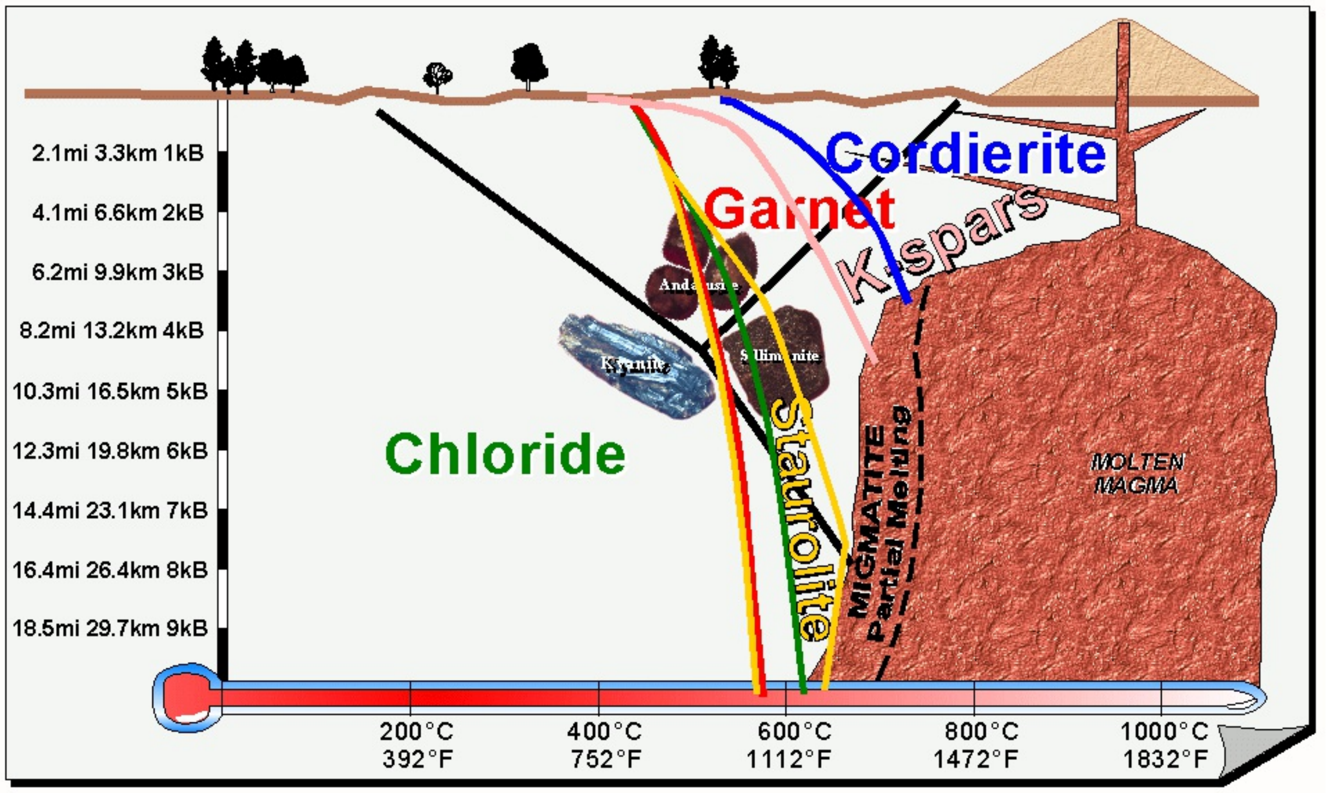

Figure 9.3 - PT Diagram of common metamorphic mineral species.

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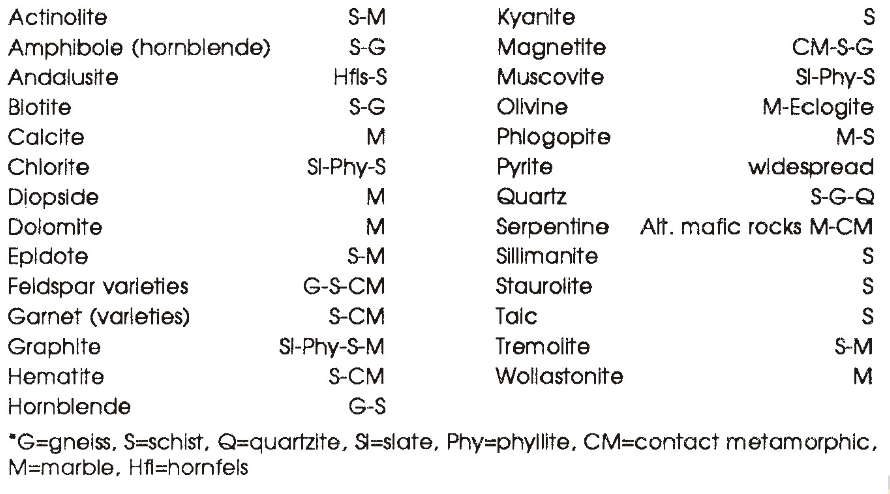
Table 9.1 - Common Minerals in Metamorphic Rocks with most likely Rock Hosts



Actinolite	S-M	Kyanite	S
Amphibole (hornblende)	S-G	Magnetite	CM-S-G
Andalusite	Hfls-S	Muscovite	Sl-Phy-S
Biotite	S-G	Olivine	M-Eclogite
Calcite	M	Phlogopite	M-S
Chlorite	Sl-Phy-S	Pyrite	widespread
Diopside	M	Quartz	S-G-Q
Dolomite	M	Serpentine	Alt. mafic rocks M-CM
Epidote	S-M	Sillimanite	S
Feldspar varieties	G-S-CM	Staurolite	S
Garnet (varieties)	S-CM	Talc	S
Graphite	Sl-Phy-S-M	Tremolite	S-M
Hematite	S-CM	Wollastonite	M
Hornblende	G-S		

*G=gneiss, S=schist, Q=quartzite, Sl=slate, Phy=phyllite, CM=contact metamorphic, M=marble, Hfl=hornfels

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9.1 TEXTURE OF METAMORPHIC ROCKS

Recognition and naming of a metamorphic rock requires identification of the major minerals and the textures and structures present. The list of minerals includes most of those common to this rock class. The processes of metamorphism generally produce increase in mineral size, whether by recombination of elements and new mineral development, or by simple crystal growth of preferred minerals. Distinct differences in textures of igneous, sedimentary and metamorphic rocks exist. For most metamorphic rocks crystals are readily visible, and while somewhat interlocked form a mosaic often with rough to well-developed layering characteristics. Some minerals tend to develop good crystal form and are called *idioblastic* with bounding crystal faces. *Xenoblastic* crystals are those without crystal form. A *porphyroblast* is a crystal much larger than the matrix and may contain other mineral inclusions. It is of late stage development in contrast to phenocrysts of igneous rocks which are formed early in the crystallization process. Porphyroblasts may lie athwart the general internal structures of the rock or may show structures bending around them, as if deformed by the developing mineral.

9.2 STRUCTURES OF METAMORPHIC ROCK

Three types of structures are recognized in the metamorphic rock class: 1) planar-parallel planes or flat surfaces of minerals parallel or roughly so; 2) a series of linear or line-like features, such as elongate minerals parallel in their long direction which may or may not lie within the planar structure; 3) massive linear or planar features not visible. One or more of these may be used in the naming of the metamorphic rock.

9.3 CLASSIFICATION AND NAMING OF METAMORPHIC ROCKS

Utilizing texture, structure, and mineral composition we can classify or name the metamorphic rock. In some cases the name may be derived solely from the structural characteristics, such as schist. In others it may be based on the mineral composition, as for quartzite or marble. Additional qualifying terms may give a more specific name to the rock. For a rock with schistose structure and clearly visible biotite, garnet and quartz present, the name would be "garnet biotite quartz schist" based on the relative abundance of each of the mineral phases present, the least abundant first. The panacea of a name which would describe origin, original rock type and intensity of metamorphism and resultant mineralogy, texture and structure has not been derived. We must depend on visible minerals, textures and structures which may or may not denote the rock history.

The chart of Table 9.2 is general in its make up but includes the main criteria for naming of this important rock class. It is self explanatory and is based on typical structures, textures, and minerals.

Table 9.2 - Classification of Metamorphic Rocks

			TEXTURE	MINERALS	ROCK NAME
STRUCTURE	MASSIVE TO FAINTLY LAYERED		Very fine grain	Identifiable under microscope	Hornfels
			Sugary texture, interlocking grains (Faint to distinct layering)	Mainly quartz	Quartzite
			Fine to coarse grains, with or without dark-streaking. Sometimes breciated.	Calcite Dolomite	Marble
		GRANOBLASTIC	Variable grain size	Silicates of calcium, iron, magnesium in or with limestone or dolomite, and near igneous contact, includes garnets, epidote, pyroxene, amphibole and others.	Tactite
			Medium to coarse-grained rock, faintly foliated	Plagioclase, amphiboles, possible garnet, quartz, epidote	Amphibolite
				Feldspar, pyroxene, garnet, kyanite and other silicates	Granulite
	FOLIATED		Very fine-grained, well foliated. Remnant sedimentary structures visble.	Mineral grains very small, microscopic	Slate
			Faint foliation to schistose and well-developed layering	Mica and quartz when recognizable	Phyllite
				Chlorites, plagioclase, epidote	Chlorite schist
				Muscovite, quartz, biotite	Mineral qualifier, as mica schist, biotite, muscovite, sericite schist,or sillimanite mica schist, kyanite, etc.
				Amphibole, plagioclase	Amphibole schist
			Gneissic-banded (dark-light bands)	Feldspar, amphibole, quartz, biotite	Gneiss
			Coarsely-banded mixture or metamorphic & igneous rocks	Feldspar, amphibole, quartz, biotite interlayered and interfingered	Migmatite

DETAILED METAMORPHIC ROCK DESCRIPTIONS

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GNEISS

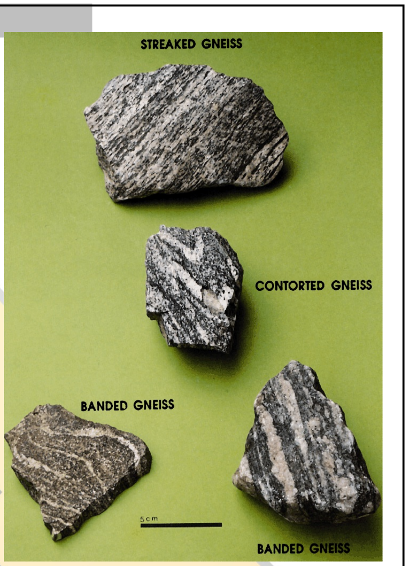
Rock Group: Foliated Metamorphic Rock
Grade of Metamorphism: High; Regional Metamorphism

Texture: The term gneiss applies to a group of rocks with banded or coarsely foliated structure. These are coarse grained, often irregularly banded. Some varieties occur near intrusive masses intricately interlayered in bedded metamorphic rocks and are known as migmatites. They are believed to have formed from the emanations from the partial melting and recrystallization of the original rock under high temperature conditions.

Banded Gneiss - Mainly composed of grainy biotite flakes.
Contorted Gneiss - Sharply folded well-developed dark & light bands.
Streaked Gneiss - distinct foliation, but streaky. Some layers end rather abruptly.

Mineralogy:. Consist mainly of quartz and feldspar with some biotite.

Protolith:. Paragneisses - Metamorphic sequence shale -> slate ->schist -> gneiss.
Orthogneisses - Metamorphism of primary igneous rocks.



HORNFELS

Rock Group: Non-Foliated Metamorphic Rock
Grade of Metamorphism: Usually Low; Contact Metamorphism

Texture: Dark colored rock, mostly brownish black. May show original bedding or other structures of the protolith (usually shale). However, Hornfels is NOT foliated since it breaks readily across these laminations and not only along them. Usually fine to very fine grained (aphanitic).

Mineralogy:. Quartz, Feldspars and Biotite mica (often responsible for the dark color). The presence of Andalusite may form spots.

Protolith:. Any; shale, sandstone, limestone, even basalt.

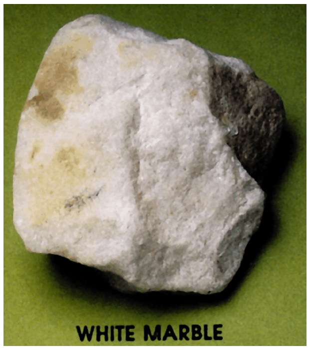
MARBLE

Rock Group: Non-Foliated Metamorphic Rock
Grade of Metamorphism: Variable; Contact & some Regional Metamorphism

Texture: Light colored rock, often white. Has increased grain size and much lighter color than the original rock. Grains are almost equal in size. Bedding planes have been essentially obliterated and the rock considerably changed in appearance.

Mineralogy:. Predominately Calcite or Dolomite. Some biotite may form from impurities in the protolith.

Protolith:. Limestone / Dolomite.



QUARTZITE

Rock Group: Non-Foliated Metamorphic Rock
Grade of Metamorphism: Variable; Contact & Regional Metamorphism

Texture: Various colors. Pink, white, dark, even green. Relic bedding or layering may still be visible. These are NOT foliations indicated by their very parallel & linear arrangement. The mineral grains fracture when the rock is broken.

Mineralogy:. Quartz. Impurities of clay in the protolith will form muscovite.

Protolith:. Sandstone.

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SCHIST *Rock Group:* Foliated Metamorphic Rock

Grade of Metamorphism: Intermediate; Regional Metamorphism

Texture: Crystalline, somewhat crenulated. and showing distinct foliate structure.

Biotite Schist	- Mainly composed of grainy biotite flakes.
Garnet Schist	-contains fine muscovite and brownish garnet metacrysts.
Muscovite Schist	- Mainly composed of grainy muscovite. Silvery appearance.
Staurolite Schist	- shows visible mineral crosses of staurolite within mica matrix.

Mineralogy: Grain size for this rock is far larger than that of the slate. Platy and other minerals are clearly visible to the eye.

Protolith: Slate.

SERPENTINITE *Rock Group:* Foliated Metamorphic Rock

Grade of Metamorphism: Low; Contact or Regional Metamorphism

Texture: Frequently green or green gray, almost wax-like in appearance with a slippery feel. Very soft, may be easily scratched or carved with a knife (soapstone).

Mineralogy: Commonly composed of hydrated magnesium silicate with an approximate composition of $3MgO*2SiO_2*2H_2O$. Other minerals such as Chlorite, Talc and Graphite may also be present. Sometimes contains fibrous veins of chrysotile (asbestos) or similar heat resistant minerals.

Protolith: Mafic or ultramafic igneous rocks, such as Basalt, Gabbro or Peridotite. Forms often at divergent plate boundaries.

SLATE *Rock Group:* Foliated Metamorphic Rock

Grade of Metamorphism: Low; Regional or Contact Metamorphism

Texture: Homogeneous appearance, very fine-grained texture and composition. May show difference in color banding. Schistosity or cleavage. is independent of bedding. The rock can be cleaved to relatively thin plates parallel to the finely developed schistosity. Numerous types of slates have been described, but the use of color terms such as gray slate or red slate dominates.

Mineralogy:. Has microscopic alignment of platy minerals such as mica and chlorite which are oriented in the plane of cleavage as a result of stress. Compositional terms such as graphitic and calcareous are also used if these minerals are present.

Protolith:. Original shale or fine siltstone.