Sedimentary Rocks (ii)

For Third Year Students (Geology, Geophysics and Geology-Chemistry)

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Sediments and sedimentary rocks

The material which was produced through the erosion and weathering of rocks exposed is called ***sediment***s, sediment transported by rivers eventually finds its way into a standing body of water. When rivers enter standing bodies of water (e.g., the Gulf), the sediment load that they are carrying is dropped and deposition occurs. Usually deposition forms more or less parallel layers called ***strata***. Given time, and the processes of compaction and cementation, the sediment may be lithified into sedimentary rock.

Sedimentary rocks cover some 80 percent of the earth's crust. They form at low temperatures and pressures at the surface of earth owing to deposition by water, wind, or ice. By contrast, igneous and metamorphic rocks form mainly below earth's surface where temperatures and pressures are higher than those at the surface.

Sedimentary rocks are characterized particularly by the presence of layers, and by distinctive textures and structures. Many sedimentary rocks are also distinguished from igneous and metamorphic rocks by their mineral and chemical compositions and fossil content.

classification of sedimentary rocks

The problem of how to explain sedimentary rocks is closely related to the problem of how to classify them. The properties of sedimentary rocks provide a basis for classification according to similarities in texture, composition, or other characteristics. However, it is generally impossible to select a single group of characteristics that allows classification of all sedimentary rock types. There are two main classifications:

1- The broad general classifications that embrace all sedimentary rocks. These are the most different classifications as most sediments are mixtures of two or more end members.

2- The specialized classifications which may include the classification of sandstones only, or carbonates only etc…

The broad and general classifications are dealt with in this section and the specialized classifications will be dealt with when describing each type of rock separately.

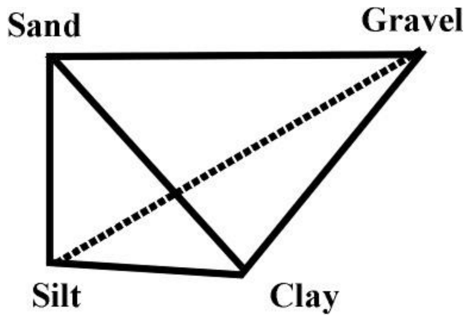
Although there have been many attempts to discuss general classifications for all sedimentary rocks non have achieved widespread acceptance. In general, sedimentologists distinguish the mojor groups of rocks one of the following two main bases:

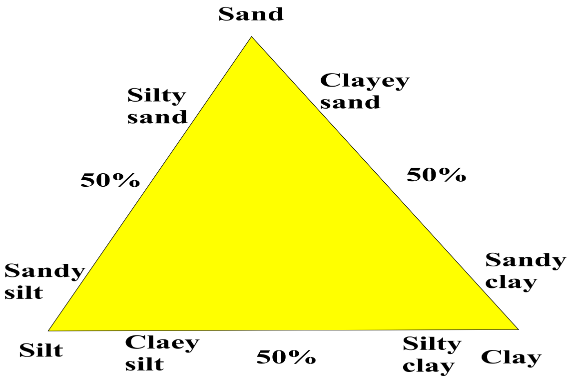
1- Descriptive basis: in this case the descriptive classification either use the similarities in textural (e.g. grain size) or in the composition as base for classification.
2- Genetic basis: and in this case the genetic classifications take the sedimentary origin into their account.

1-Descriptive classifications

A- based on texture

A tetrahedron may be used to indicate the textural relations of the sediments.

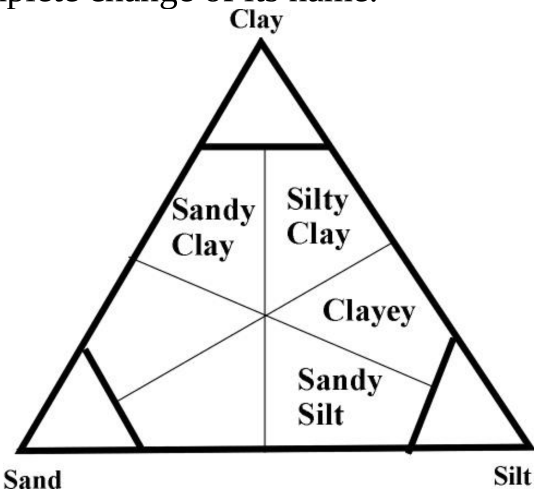




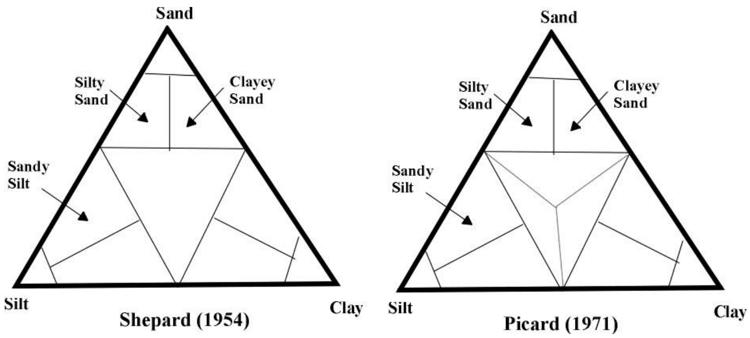
The figure shows a tetrahedron with gravel, sand, silt and clay end members (i.e. different sediments with different grain sizes). These may be mixed in any proportions to form typical classes of detrital rocks.

Some of the transitional groups are shown in the sand-silt-clay face of tetrahedron. The admixture of silt with sand for example yields such intermediate textures as silty sand and sandy silt.

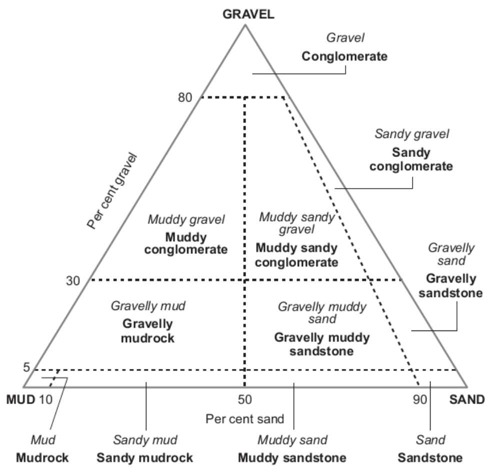
The textural triangle includes three end-member sediments plus six intermediate kinds. In this general case the six intermediate classes would meet at the centre of the triangle, as shown by illustration. Note that for sediments containing approximately equal amounts of sand, silt and clay, a slight compositional difference in the sediment results in a complete change of its name.



To overcome this disadvantages, Shepard (1953) and Picard (1971) developed a modified end-Member triangles having a triangle area in its centre. These modifications provide convenient separate class for those sediment that tend to fall within the central part of the end-member triangle an additional advantage is that the six intermediate mixtures have more restricted allowing more discrimination between them.



Another mode of textural Classification is given by Folk (1954) as follows:

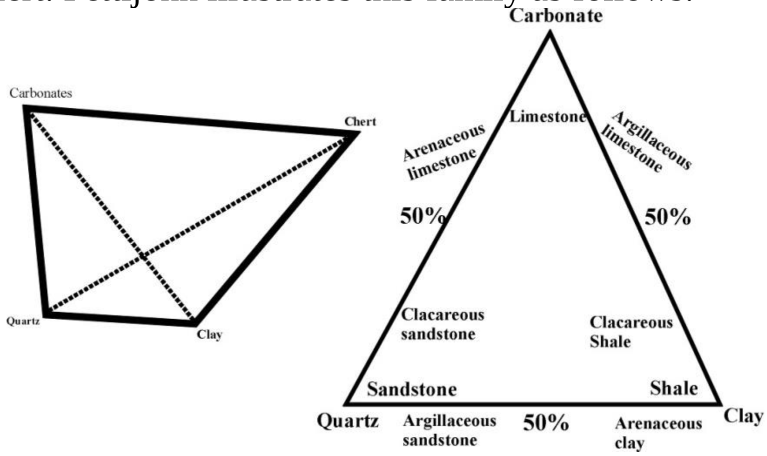


Folk'S nomenclature for mixtures of gravel, sand and mud in sedimentary rock.

In this classification the median size of the deposits specified wherever possible. Also the composition of mud, silty, muddy, or clayey is also specified wherever present. For Example: Sandy gravel, Muddy sandy gravel.

B- based on composition

A tetrahedron may also used to show the composition of sediments. One of the commonest major families of sedimentary rocks is based on the very commonly occurring end members quartz, clay, carbonate and chert. Pettijohn illustrates this family as follows:



2-Genetic (Fundamental) classifications

At a more advanced level than the descriptive classifications are these called "genetic" which group together sedimentary rocks that have a common origin, and on the basisof the dominant processes operating. There are difficulties in constructing such classification because almost all sedimentary rocks have complex origin.

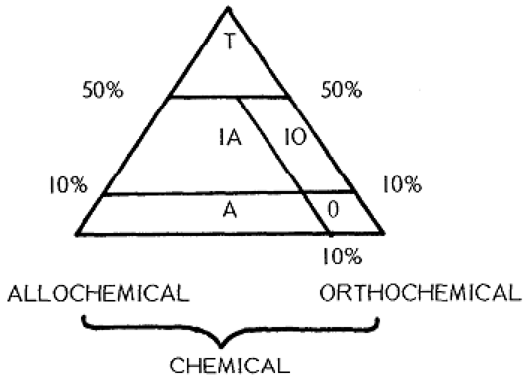
As an example of the genetic classifications Folk,S classification (1974) is here given. Folk belives that sediments consist fundamentally three components which may be mixed in nearly all proportions.

A)- Terrigenous components are those substances derived from erosion of a land area outside the basin of deposition, and carried into the basin as som examples: quartz or feldspar sand, heavy minerals, clay minerals, chert or limestone pebbles derived from erosion of older rock outcrops.

B- Allochemical constituents (Greek: “allo” meaning different from normal) are those substances precipitated from solution within the basin of deposition but which are “abnormal” chemical precipitates because in general they have been later moved as solids within the basin; they have a higher degree of organization than simple precipitates. Examples: broken or whole shells, oolites, calcareous fecal pellets, or fragments of penecontemporaneous carbonate sediment torn up and reworked to form pebbles.

C)- Orthochemical constituents (Greek: “ortho” meaning proper or true) are “normal” chemical precipitates in the customary sense of the word. They are produced chemically within the basin and show little or no evidence of
significant transportation or aggregation into more complex entities. Examples: microcrystalline calcite or dolomite ooze, probably some evaporites, calcite or quartz porefi llings in sandstones, replacement minerals.

Class B and C are collectively referred to as chemical constituents, Classes A and B can be collectively termed " fragmental". Folk adds that some people use " detrital" or 'clastic" as a collective term including both "terrigenous" and "allochemecal". Because of this confusion Folk states that it is best that the terms detrital and clastic be dropped.



T : Terrigenous Rocks. Example: most mudrocks, sandstones, and con glomerates. Comprise 65% to 75% of the stratigraphic section. Most terrigenous rocks occur in the shaded area.

IA :Impure Allochemical Rocks. Example: very fossiliferous shales; sandy fossiliferous or oolitic limestones. Comprise I O-15% of the stratigraphic section.

IO: Impure Orthochemical Rocks. Example: clayey microcrystalline lime stones. Comprise 2-5% of the stratigraphic section.

A: Allochemical Rocks. Example: fossiliferous, oolitic, pellet or intraelastic limestones or dolomites. Comprise 8-15% of the stratigraphic section.

O: Orthochemical Rocks. Example: microcrystalline limestone or dolomite; anhydr i te; chert. Comprise 2-8% of the strat igraphic section

Collectively, “IA” and “10” are classed as Impure Chemical Rocks, and “A” and “0” as Pure Chemical Rocks.

Accordingly, Folk classified the sedimentary rocks into three basic classes based on the proportions of these three fundamental end members as shown in the triangular diagram. These are:

1. Terrigenous siliciclastic rocks

Siliciclastic rocks are composed of broken rock fragments formed during erosion of bedrock . They comprise a diverse group of rocks, ranging from fine-grained mudrocks (shales), through sandstones to the coarser-grained conglomerates and breccias. The sediments are composed largely of grains (clasts), or aggregates of minerals . The minerals are mainly silicates such as quartz, feldspars, and micas. The rock fragments are clasts of igneous, metamorphic, or older sedimentary rock that are also composed dominantly of silicate minerals derived from pre-existing igneous, metamorphic and sedimentary rocks. The clastic grains are released through mechanical and chemical weathering processes, and then transported to the depositional site by a variety of mechanisms, including wind, glaciers, river currents, waves, tidal currents, debris flows and turbidity currents.

Grain Size (mm)	Sediment Name		Rock Name	
Coarse ↑	boulders	GRAVEL	CONGLOMERATE (rounded clasts) or BRECCIA (angular clasts)	
64	cobbles			
16	pebbles			
2.00	v. coarse sand	SAND	SANDSTONE	
1.00	coarse sand			
0.50	medium sand			
0.25	fine sand			
0.125	very fine sand			
0.063	silt	MUD	SILTSTONE	MUD-STONE
0.004	clay		CLAYSTONE	
Fine				

Grain size characterization of terrigenous, siliciclastic sedimentary rocks

2. Chemical and biochemical rocks

This group of sedimentary rocks are produced through biochemical or chemical precipitation. The most volumetrically important are composed of carbonate minerals (***limestone and dolostone***). Other important of chemical of this group are ***chert, ironstones and the evaporites***. Coals are a special variety of biochemical sedimentary rocks.

For example: Limestone is composed of the carbonate minerals calcite and aragonite ($CaCO_3$). These minerals occur in two basic forms: 1) fragmented grains of living organisms such as oysters, corals, etc., and 2) as a chemical cement precipitated between the grains that hold the rock together. It is for these reasons that limestones are classified as *"**biochemical**"* sedimentary rocks. Chemically and biochemically produced sediment is frequently deposited in the exact same place it is produced. This is known *as* ***in situ deposition***.

3- Volcaniclastic rocks

Volcaniclastic deposits constitute a fourth category and consist of lava and rock fragments derived from penecontem-poraneous volcanic activity. constitute a fourth category and consist of lava and rock fragments derived from penecontem-poraneous volcanic activity.

Importance of sedimentary rocks

Sedimentary rocks cover roughly three-fourth of earth's surface. They have special genetic significance because their textures, structures, composition, and fossil content reveal the nature of past surface environments and life forms on Earth. Thus, they provide available clues to evolution of earth's landscapes and life forms through time. In addition, many sedimentary rocks contain minerals and fossil fuels that have economic significance. Petroleum, natural gas, coal, salt, phosphorus, sulfur, iron and other metallic ores, uranium, building materials and many other essential raw materials are examples of economic products that occur in sedimentary rocks. Also, environments and processes of deposition and palaeogeography and palaeoclimatology can all be deduced from studies of sedimentary rocks.

COARSE GRAINED TERRIGENOUS ROCKS (Conglomerates and Breccias)

Siliciclastic sedimentary rocks that consist predominantly of gravel-size (>2mm) clasts are called conglomerates and breccias. The term rudite is also sometimes used for these rocks. Conglomerates are common rocks in stratigraphic sequences of all ages, but make up less than about one percent by weight of the total sedimentary rock mass.

The framework grains of conglomerates are composed mainly of rock fragments (clasts) rather than individual mineral grains. These clasts may consist of any kind of rock. Some conglomerates are composed almost entirely of highly durable clasts of quartzite, chert, or quartz. Others are composed of a variety of clasts, some of which, limestone and shale clasts. Conglomerates may contain various amounts of matrix, which commonly consists of clay- or sand-size particles or a mixture of clay and sand.

The gravel-size material in conglomerates consists mainly of rounded to subrounded rock fragments (clasts). By contrast, breccias are aggregates of angular, gravel-size fragments. The particles in breccias are distinguished from those in conglomerates by their sharp edges corners. Many breccias, such as volcanic and tectonic breccias, are non sedimentary in origin.

Classifications of conglomerates and breccias

1-based on origin

Major types	Subtypes	Origin of clasts
Epiclastic conglomerate and breccia	Extraformational conglomerate and breccia	Breakdown of older rocks of any kind through the processes of weathering and erosion; deposition by fluid flows (water, ice) and sediment gravity flows
	Intraformational conglomerate and breccia	Penecontemporaneous fragmentation of weakly consolidated sedimentary beds; deposition by fluid flows and sediment gravity flows
Volcanic breccia	Pyroclastic breccia	Explosive volcanic eruptions, either magmatic or phreatic (steam) eruptions; deposited by airfalls or pyroclastic flows
	Autobreccia	Breakup of viscous, partially congealed lava owing to continued movement of the lava
	Hyaloclastic breccia	Shattering of hot, coherent magma into glassy fragments owing to contact with water, snow, or water-saturated sediment (quench fragmentation)
Cataclastic breccia	Landslide and slump breccia	Breakup of rock owing to tensile stresses and impact during sliding and slumping of rock masses
	Tectonic breccia: fault, fold, crush breccia	Breakage of brittle rock as a result of crustal movements
	Collapse breccia	Breakage of brittle rock owing to collapse into an opening created by solution or other processes
Solution breccia		Insoluble fragments that remain after solution of more soluble material; e.g. chert clasts concentrated by solution of limestone
Meteorite-impact breccia		Shattering of rock owing to meteorite impact

Fundamental genetic types of conglomerates and breccias

From the table above, c.onglomerate clasts can be grouped into intraformational and extraformational conglomerates. The intraformational type means clasts derived from within the basin of deposition, whereas the extraformational type means pebbles are derived from beyond the area of sedimentation.

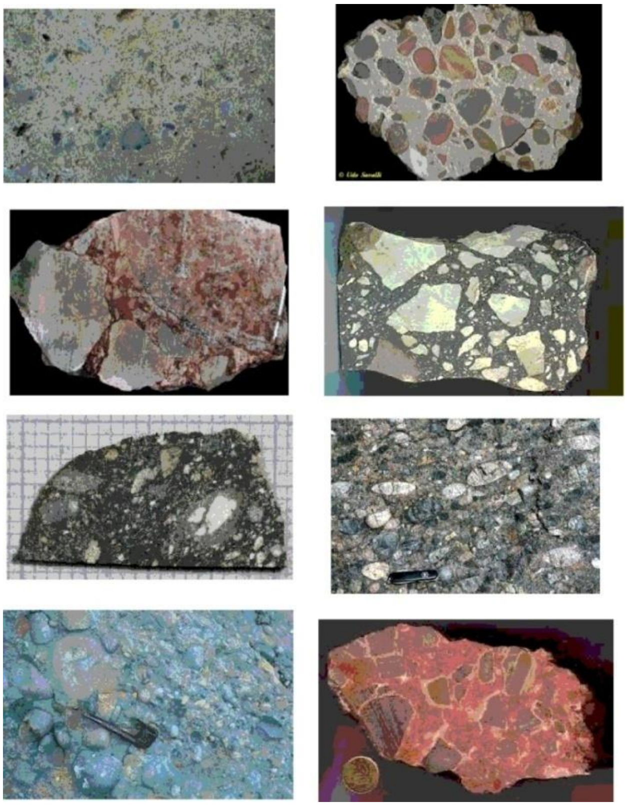
Particular types of breccia include slump breccia, consisting of broken and brecciated beds derived from downslope slumping, collapse breccia, resulting from the dissolution of evaporites and the collapse of overlying strata, and palaeokarstic breccia, produced by dissolution of limestone, formation of caves, fracture and collapse. Carbonate megabreccias, with very large clasts (>4m diameter), may have a sequence stratigraphic significance and relate to a relative sea-level fall

2-based on composition of framework clasts

Gravel-size particles are the framework grains of conglomerates. Conglomerates may contain gravel-size pieces of igneous, metamorphic, or sedimentary clasts.The term oligomict conglomerateis (mono- or oligomictic conglomerates) often applied to stable conglomerates composed mainly of a single clast type. Polymict conglomerates (polymictic conglomerates, or breccias) made up of a mixture of clasts such as igneous, limestone, shale, and metamorphic clasts.

3- based on fabric support

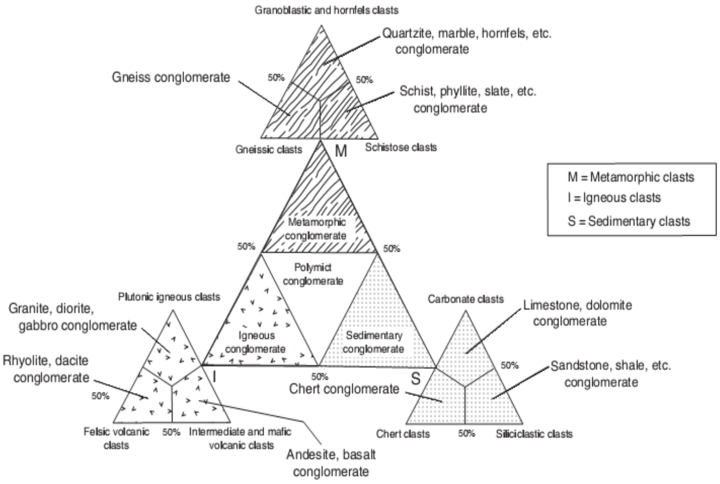
If mud or sand matrix is abundant within rudites or gravelly sediment, the fabric is commonly referred to as matrix-supported framework (paraconglomerates). This also called diamictites. Rudites or gravelly sediments that contain little matrix that the gravel-size framework grains touch and thus form a supporting framework are called clast-supported (orthoconglomerates).



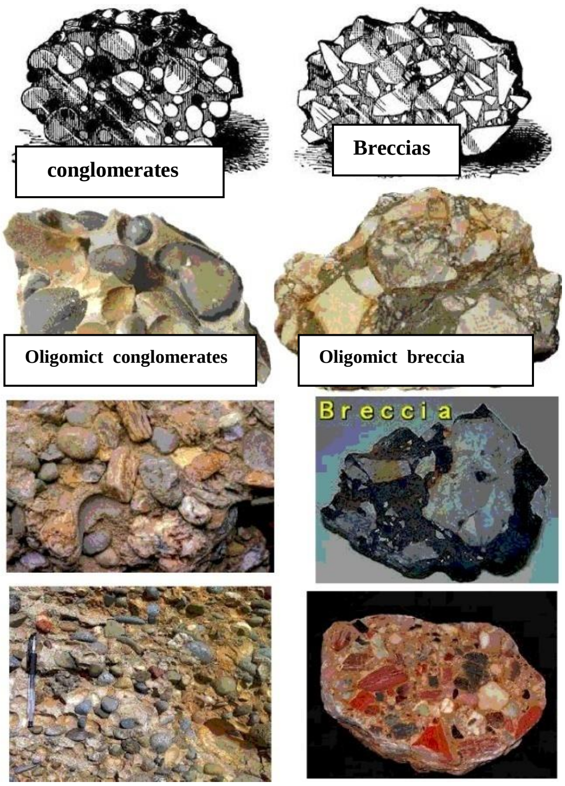
Matrix-supported Breccias

4- based on clast lithology,

Conglomerate clasts can be grouped into three fundamental kinds: igneous, metamorphic, and sedimentary. Accordingly, conglomerates can be classified on the basis of these end-member clast types into four kinds of conglomerates: metamorphic (clast) conglomerates, igneous (clast) conglomerates, sedimentary (clast) conglomerates, and polymict conglomerates. The term polymict can be applied informally to all conglomerates containing clasts of mixed lithology.



Classification of conglomerates on the basis of clast lithology.



5- based on origin Quartzose conglomerates

Quartzose conglomerates consist dominantly of metaquartzite, vein-quartz, or chert clasts, are derived from metasedimentary, sedimentary, and some igneous rocks. the concentration of chert clasts in a conglomerate implies destruction of large volumes of chert-nodule limestone to yield a chert-clast concentrate. Some chert clasts may, of course, be derived also by erosion of bedded chert deposits.

The source rocks that yield quartzose clasts are most likely to occur in recycled orogen or continental block provenances. They appear to be largely of fluvial, particularly braided-stream, origin, but marine, wave-worked quartzose conglomerates also exist.

Petromict conglomerates

The volume of ancient petromict conglomerates is far greater than that of quartzose conglomerates. The petromict conglomerates form the truly great conglomerate bodies of the geologic record, and they may reach thicknesses of thousands of meters. Petromict conglomerates contain significant amounts of metastable rock fragments. They may be derived from many different types of plutonic igneous, volcanic, metamorphic, or sedimentary rock. Petromict conglomerates can accumulate in any tectonic provenance (continental block, recycled orogen, or magmatic arc) where the requisite con-ditions that allow their preservation aremet. They are deposited in environments ranging from fluvial through shallow-marine to deep-marine, although the bulk of the truly thick petromict conglomerate bodies are probably non marine.

Conglomerate properties and depositional environments

The goal of many conglomerate studies is to identify the paleoenvironments of conglomerates on the basis of these characteristic properties. Gravels are deposited in modern environments ranging from fluvial to deep-marine by a variety of fluid-flow, ice-flow and sediment-gravity-flow processes. We assume that ancient conglomerates were deposited in similar environments and that the characteristics of ancient conglomerates and modern gravels deposited in similar environments are also similar.

The conglomerates are subdivided into nine fundamental types: sheetflood (braided-stream) conglomerate, stream flow conglomerate, wave-worked conglomerate, wave-, storm-, and current-worked conglomerate, tide-worked conglomerate, melt out conglomerate, subaqueous meltout conglomerate, subaerial debris-flow conglomerate, and resedimented conglomerate. Resedimented conglomerates are further divided into subaqueous debris-flow conglomerates, subaqueous grain-flow conglomerates, and turbidite conglomerates.

How Does Breccia Differ From Conglomerate?

Breccia and conglomerate are very similar rocks. They are both clastic sedimentary rocks composed of particles larger than two millimeters in diameter. The difference is in the shape of the large particles. In breccia the large particles are angular in shape but in conglomerate the particles are rounded. This reveals a difference in how far the particles were transported. Near the outcrop where the fragments were produced by mechanical weathering the shape is angular. However, during transport by water away from the outcrop the sharp points and edges of those angular fragments are rounded. The rounded particles would form a conglomerate**.**

Talus Slopes: Scene of a mountain environment where talus, the angular mechanical weathering debris that might form breccia, is produced in abundance.

(Model In Egypt)
Structural control on syn-rift conglomerates, northwestern Red Sea margin, Egypt
(Khalil* and McClay, ;Mahran et al., 2014)

Wadi Sharm El Bahari is located about 30 km south of Quseir . In this Wadi the conglomerates of the Ranga Formation form a fan delta which has wedge-like to sheet-like geometry and an east to southeastwards progradation direction. The length of the fan in the progradation direction is 3 km and its average strike length is 2 km. The maximum height of the fan is about 200 m above the wadi floor and it decreases eastwards (at distal areas) to less than 50 m high. This fan sequence is 60–100 m in thickness.

The conglomerate sequence is characterized by large-scale foresets which are 20–40 m high. The average dip of these foresets is 22, but this decreases gradually in both updip and downdip directions to less than 10 , producing sigmoidal to gently oblique clinoforms. The foreset beds are moderately stratified, with bed thickness ranging from 0.5 to 3 m. These units are composed of chert-dominated conglomerate which are both clast supported and sand-matrix supported. Clasts are 2–10 cm in diameter, rounded to subrounded and inplaces have an imbricated fabric. When tracing the foresets updip, the beds become more sandier, shallow dipping, and contain shell fragments and patch reefs. These are onlapped by coral and algal carbonates of Um Mahara Formation.

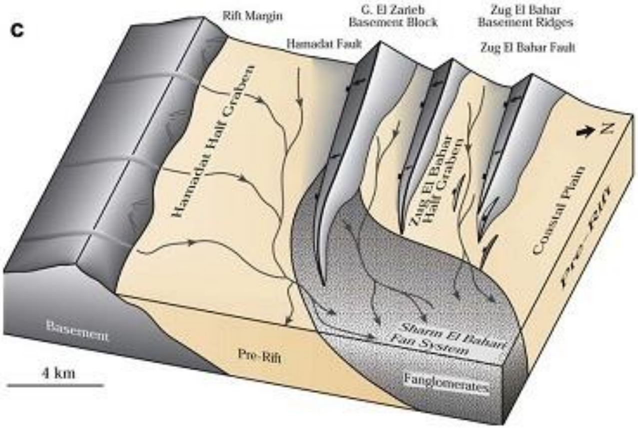
The conglomerate facies of the fan sequence in Wadi Sharm El Bahari is interpreted to indicate the dominance of fluvial processes, with the coarse sediments transported to the foreset slopes by sheet flows. Foreset bed dips and imbricated clast fabrics indicate ESE to SE palaeoflow directions.

Model for the structural control on the Miocene fan delta system

The localization of the fan delta in Sharm El Bahari appears to have been controlled by the WNW and NW-trending, WSW-dipping faults (Hamadat fault) that bound the Hamadat half graben.

The plunge of the southern corner of the fault block and the associated with the decrease of the displacement and low topographic relief led to open the half graben basins southwards (The Early Rift-Climax stage), and acted as pathway for clastic sediments derived from the uplifted rift margins to the west and northwest. Also, the basement footwall blocks of these faults (Gebel El Zarieb ridge) are interpreted to have formed barriers against the eastward transportation of the clastics supplied by the basement of he rift shoulder in the west. The Hamadat fault offset and lose displacement as they are traced south-eastward (towards Sharm El Bahari area). This basement block has their lowermost structural relief at the locality of Sharm El Bahari. Due to these combined structural and morphologic elements, the Miocene clastics, therefore, instead of being transported eastwards towards Zug El Bahar in the coastal area, are considered to have been carried southeastward, towards theWadi Sharm El Bahari sediment-input point,

The chert-dominated conglomerates of the Wadi Sharm El Bahari fan are interpreted to have been derived from the pre-rift Cretaceous-Eocene stripped from the fault blocks in Wadi Hamadat and probably also from the pre-rift strata stripped from the basement highlands further to the west.

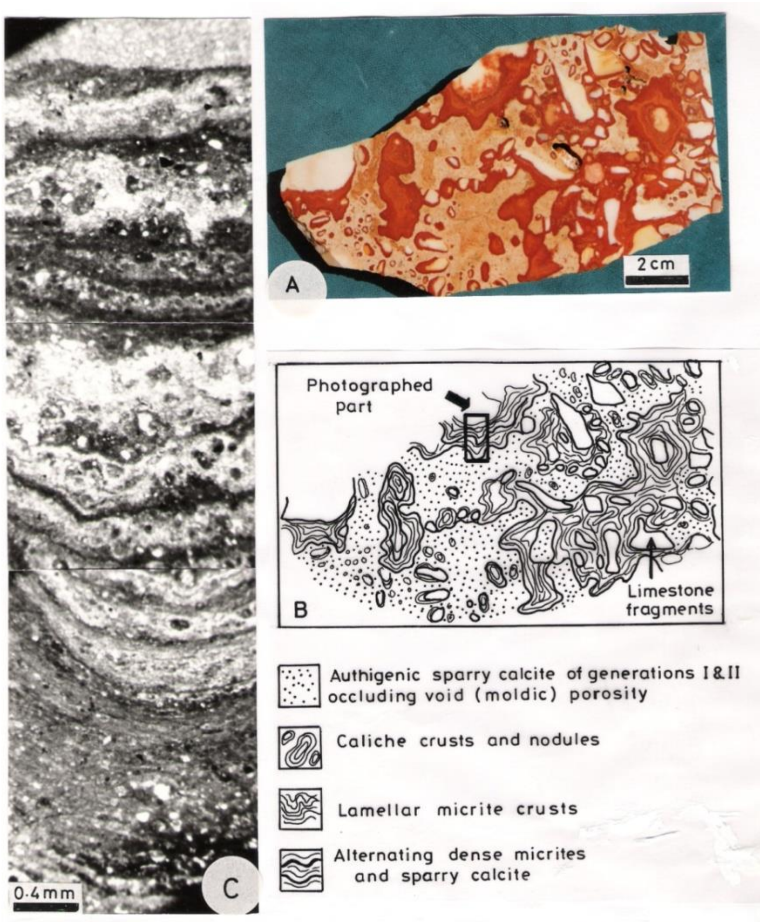


Red Breccias

- The post-Eocene breccias outcropping around Sohag extend as a separate belt occupying the proximity of a fault-controlled escarpment.
- Based on the stratigraphic level and the textural characteristics these breccias are subdivided into three types, clearly separated by paleosol horizons. 1- Breccia 1, composed of angular, massive and crushed deposits accumulated as fault breccias. Its distribution is structurally-controlled. It occurs along the N140, N40-60 and N10-20 fault intersections which most probably resulted during the formation of the Nile gulf in post Eocene time..
- B- Breccia 2, occurs as talus composed of scree and landslides and collapsed breccias flanking the Eocene escarpment.
- C- Breccia 3, are recorded at the top of the Issawia Formation. They are intercalated with white breccias and palustrine-lacustrine carbonates. They accumulated by running water as debris flow-dominated alluvial fans . They comprise four generations of fans sourced from reactivated fault scarps and easterly and northeasterly trending wadis

– Field and petrographic studies have shown that the sediments of red breccias became hard and indurated as a result of the development of caliche carbonates that cemented and encrusted the lithoclasts. This intergrain caliche crusts and cements have possibly resulted from calcimorphic paleosols cover that separated the red breccia types, as well as due to dissolution and reprecipitation of calcareous materials. These crusts later absorbed the red ferrugineous coloration.

It is belived that the calcimorphic paleosol and the associated caliche crusts as well as red iron pigmentation aroubd breccia clasts have genetically developed during periods of exposure which alternated with intense rain fall and evaporation



MEDIUM GRAINED TERRIGENOUS ROCKS

Sandstones

Sandstones make up nearly one-quarter of the sedimentary rocks in the geologic record. The term sandstone was rongly used by beginners to identify rocks that are mainly composed of quartz grains. For a specialist, on the other hand, the term sandstone means consolidated clastic material of sand size.

Sandstone varieties

Classification of sandstones

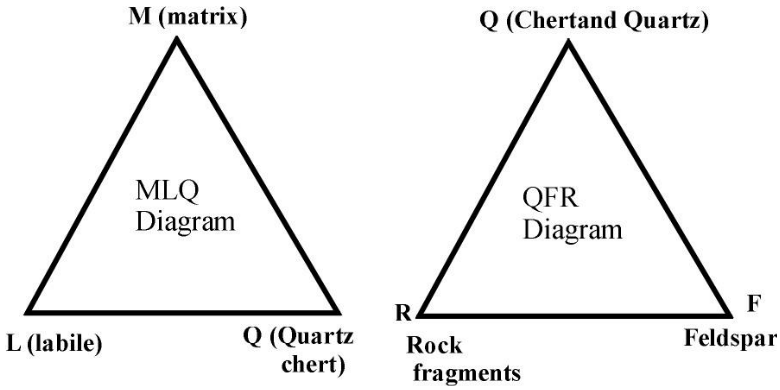
Generally, sandstones can be separated into two groups: epiclastic and volcani-clastic. Epiclastic deposits are formed from fragments of pre-existing rocks derived by weathering and erosion. Thus, they are composed mainly of silicate minerals and various kinds of igneous, metamorphic, and sedimentary rock fragments. Volcaniclastic deposits are those especially rich in volcanic debris, including glass. Many volcaniclastic deposits consist principally of pyroclastic materials such as ash or lapilli, derived directly through explosive volcanism

Epiclastic sandstones

The framework grains in most sandstones are dominated by quartz, feldspars, and rock fragments. Some sandstones contain matrix in addition to sand-size framework grains. The matrix content of sandstones may range from zero to several tens of percent.

The classification of a sandstone is based on microscopic studies and requires an assessment of the percentages of the various grain types present. There are several classification schemes available and most use a triangular diagram with end members of quartz (Q), feldspar (F) and rock fragments (L).

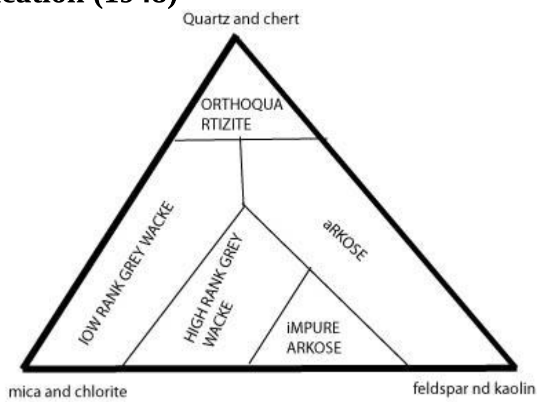
<u>MLQ and QFR diagrams:</u> Two types of triangular diagrams have been widely employed in the classification of sandstones. Pettijohn (1949) and Packham (1954) have used MLQ diagrams which have percentages of matrix (M), Labiles (feldspar, rock fragments, ferromagnesians, chlorite), (L), and quartz plus chert (Q) as parameters.



Gilbert (1955) on the other hand, utilized QFR diagrams, whose parameters are percentages (on a matrix- and cement-free basis) of (quartz plus chert(Q), feldspar (F), and rock fragments plus other labiles (R).

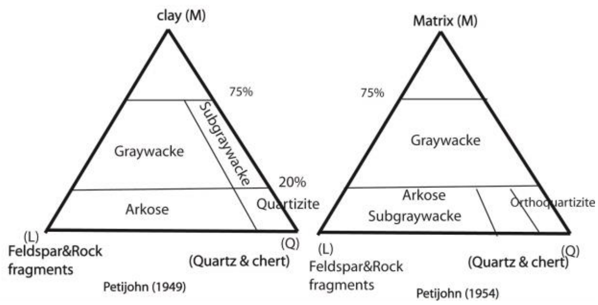
In the following page, the most important proposed sandstone classification are given very briefly.

Krynine classification (1948)



Pettijohn's classifications

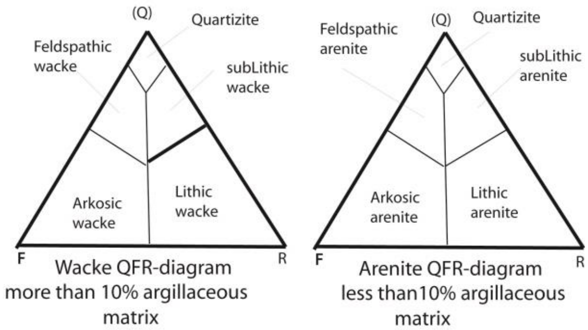
Pettijohn (1954) recognized 3 factors to be of greatest genetic significance: a maturity factor ratio of quartz+chert to labiles on a matrix-free basis); a provenance factor (ratio of sand detritus to matrix). His classification, which is shown in the form of an MLQ in the following figure is based on these factors.



Pettijohn (1954) gave a slightly different classification, although based on the same 3 factors mentioned above. His classification is shown in the following table.

<u>Gilbert's classification</u> (1955)

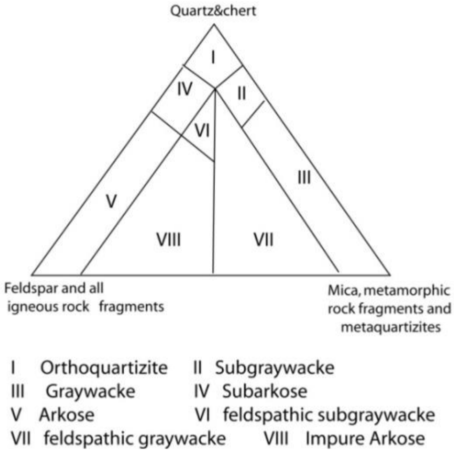
Gilbert,s classification is basically similar to Pettijohn,s 1954 classification, it employs the same three factors, but uses two diagrams of the QFR type. The first is the wacke diagram" for arenites with more than 10% detrital matrix. Thus it is the equivalent of those parts of Pettijohn,s (1954) MLQ diagram which lie above the 10% level of matrix. The second QFR diagram is the " arenite diagram", for sandstones with less than 10% detrital matrix. It is the equivalent of the remainder of the MLQ diagram of Pettijohn,s. Combining the matrix parameter with composition (QFL, seeFig. 4.12for explan-ation) yields six kinds of sandstones: quartz arenites and wackes, feldspathic arenites and wackes, and lithic aenites and wackes



Folk,s classification(1954)

Folk,s classification does not utlize either QFR or MLQ diagram, but uses a ternary diagram; quartz+ chert; feldspar+all igneous ock fragments, and mca plus metamorphic rock fragments plus metaquartizite as parameters. The diagram is subdivided in the manner to the figure. Unlike the classification already discussed Folk,s scheme does not utilize matrix-content as a parameter, and the names given to arenites in his scheme are thus uninfluenced by the matrix-content of the rock.

Folk,s scheme does not attempt the mode of sedimentation the parameters being chosen with the intent to classify arenites according to provenance. The three parameters given are held to measure (a) detritus of sedimentary origin , or detritus which has undergone a prolnged period of chemical and for physical modification, (b) detritus of igneous origin, and (c) detritus of metamorphic origin.



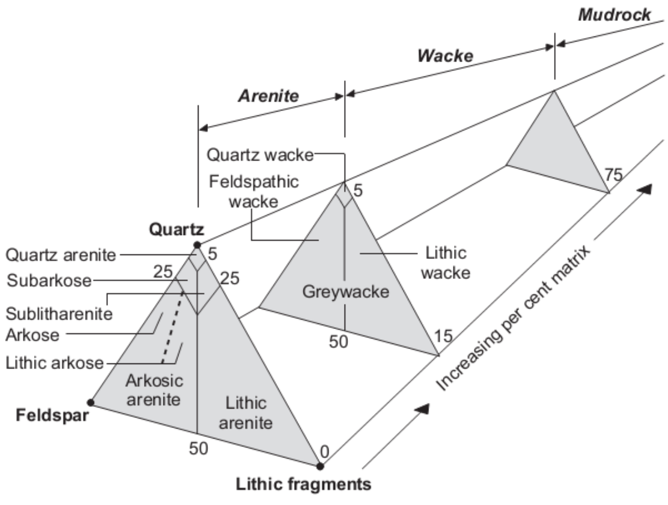
The Pettijohn classification of sandstones(Pettijohn 1975)

The widely used simple Pettijohn sandstone classification combines textural criteria, the proportion of muddy matrix, with compositional criteria, the percentages of the three commonest components of sandstone: **quartz, feldspar and lithic fragments**. The triangular plot has these three components as the end members to form a ‘Q, F, L’ triangle. The matrix is the silt and clay material that was deposited with the sand grains. If the amount of muddy matrix present is less than 15% the rock is called an arenite, between 15% and 75% it is a wacke and if most of the volume of the rock is fine-grained matrix it is classified as a mudstone.

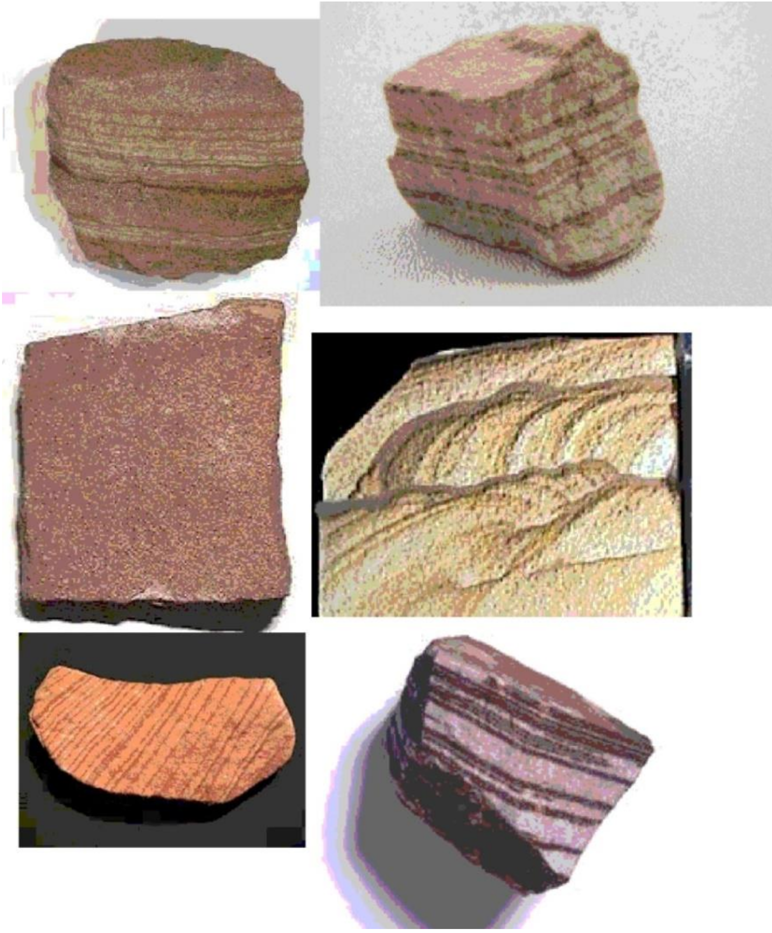
Quartz is the most common grain type present in most sandstones so this classification. Only 25% feldspar need be present for the rock to be called a feldspathic arenite, arkosic arenite or arkose. Also 25% of lithic fragments in a sandstone make it a lithic arenite. Over 95% of quartz must be present for a rock to be classified as aquartz arenite, a rock type for-merly referred to as orthoquartzite (quartzite is the low-grade metamorphic equivalent). Arkosic arenite refers to an arenite with more than 25% feldspar, which exceeds the rock-fragments content, and litharenite is applied where the rock-fragment content exceeds 25% and is greater than feldspar. The arkosic arenites can be divided into arkose sand lithic arkoses. Sandstone with intermediate percentages of feldspar or lithic grains is called subarkosic arenite and sublithic arenite.

The wackes are the transitional group between arenites and mudrocks. The most familiar is the greywacke and two types are distinguished: feldspathic and lithic wacke. The term arkosic wacke is used for arkoses with a significant proportion of matrix.

The term **greywacke** has been used in the past for a sandstone that might also be called a feldspathic or lithic wacke. They are typically mixtures of rock frag-ments, quartz and feldspar grains with a matrix of clay and silt-sized particles.



The Pettijohn classification of sandstones(Pettijohn 1987), on the basis of three mineral components: Q = quartz, chert, quartzite fragments; F = feldspars; L = unstable lithic grains (rock fragments)



Petrography and origin of principal sandstone types

The four common types of sandstone are quartz arenite, arkose, litharenite and greywacke. They are typical of certain depositional environments, but because of the provenance control on sandstone composition, they are not restricted to a par-ticular depositional setting. They reflect the geology of the source area to a greater or lesser extent, depending on weathering and relief.

Quartz arenites

Sandstones with at least 95% quartz grains are the most compositionally mature of all sandstones . In addition, they usually consist of well-rounded and well-sorted grains so that textural maturity is also very high. Cements are typically quartz overgrowths, but calcite also is common, normally a poikilotopic cement. Monocrystalline quartz grains dominate, because the less stable, undulatory and polycrystalline grains have been eliminated. Common heavy minerals are rutile, tourmaline, zircon and ilmenite. Very pure quartz sands with little cement are referred to as glass sands. Quartz arenites commonly show the effects of pressure dissolution, with sutured grain contacts. In many cases quartz arenites are the products of extended periods of sediment reworking, so that all grains other than quartz have been broken down by mechanical abrasion. A warm humid climate in the source area can also play a major role in producing quartz arenites, and will lead to the removal of many unstable grains, and if this is coupled with low relief and slow sedimentation rates, the quartz will dominate the detritus. Many quartz grains in these arenites could be second cycle, derived from pre-existing sediments. Evidence for this is provided by grains with abraded overgrowths from an earlier diagenetic cycle.

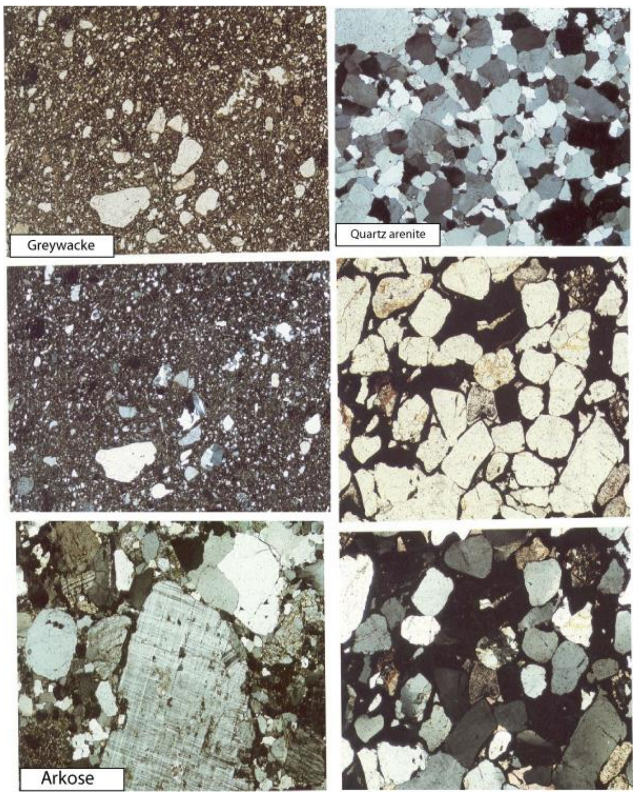
Provenance , tectonic setting of Quartz arenites

In many cases , quartz arenites as the product of extensive periods of sedimrnt reworking, so that all grains other than quartz have been brocken down.

The majority of quartz grains are second cycles, derived from pre-existing sediments. Quartz arenites of this type often deposited on shallow marine shelves, with little tectonic movements, and basin subsides very slowly.

Arkoses

Arkoses contain more than 25% feldspar, much quartz and some rock fragments. Detrital micas are also present and some fine-grained matrix. The feldspar is mainly potassium feldspar(mostly microcline). The feldspar usually is fresh, although some may be altered to kaolinite and sericite. Polycrystalline quartz and quartz/ feldspar rock fragments are common. Arkoses are typically red or pink colour. The texture of the arkose is typically poorly sorted to well sorted, with very angular to subrounded grains . Grain-supported arkoses are cemented by calcite or quartz, whereas others are cemented by a matrix .Arkoses are clearly derived from feldspar-rich rocks, particularly K-feldspar-bearing granites and gneisses. Arkoses are affected by paleoclimate..

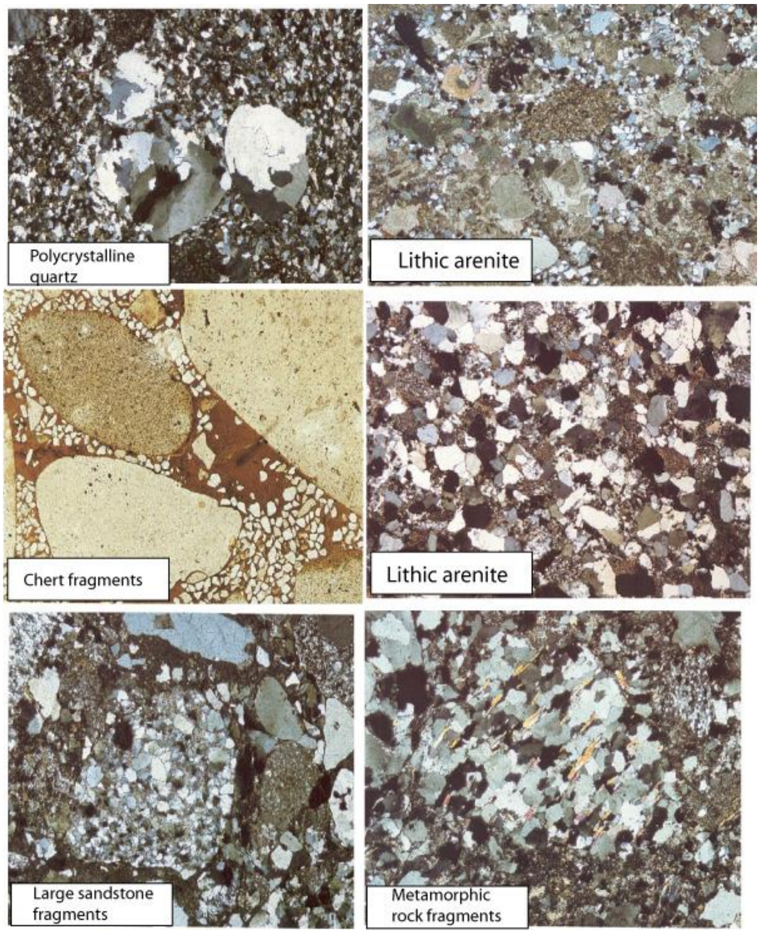


Provenance , tectonic setting of arkoses

Arkoses are derived from granites and gneisses and vary from in-situ weathering products, to stratified and cross bedded arkoses where there has been substantial sediment transport. Arkoses are clearly derived from feldspar-rich rocks but apart from an appropriate provenance geology, climae and source areas relief are also important factor. Under humid conditions, feldspars altered to clay minerals, so that semi-arid and glacial climates favour arkose formation. If erosion is very rapid, however, particularly the case where the source area has a high relief, then arkosic detritus can be pro-duced in spite of intense chemical weathering. Many arkoses were deposited in fluvial environments

Litharenites

These sandstones are characterized by a rock-fragment content that is in excess of feldspar. These are chiefly fragments of mudrock and their low-grade metamorphic equivalents, and volcanic grains; other components are flakes of mica, some feldspar and much quartz. Cements usually are either calcite or quartz, and authigenic clays are common. The nature of rock fragments in litharenites may change through time reflecting uplift in the source area, and the availability of different rock types.



Provenance , tectonic setting of Litharenites

Litharenites account for some 20% to 25% of all sndtones. Their immature composition and implies high rates of sediment production from supracrustal sources followed by short to moderate transport distances. Many fluviatile and deltaic sandstones are litharenites.

Greywackes

The characteristic feature of greywackes is the fine-grained matrix, which consists of an intergrowth of chlorite, sericite and silt-sized grains of quartz and feldspar. Quartz dominates over rock fragments and feldspar. Fine - grained sedimentary and metasedimentary rock types dominate. Igneous rock fragments are common in some greywackes, especially grains of more acidic and andesitic extrusives. Feldspar grains are chiefly fresh sodic plagioclase. Greywackes are dark grey or black rocks, usually well indurated; they can look like dolerite. Geochemically, Greywackes differ from arkoses in the dominance of FeO over Fe 2O3, MgO over CaO and Na2O over K2 O.

Provenance , tectonic setting of Greywackes

Many greywackes were deposited by turbidity cur-rents in basins of various types, usually off continental margins, in back-arc and fore-arc basins, and in asso-ciation with volcanics.

Frequently greywackes are transported by masses of water highly charged with suspended sediment. Because of the suspended matter, the mass is denser than surrounding water and moves alng the sloping sea floor or down submarine canyons as a turbidity current.

Graywackes sediments are characteristically accumulates in deep sea fans at the base of the continental slope.

Sedimentary rocks	Source area lithology	Paleo-climate	Tectonic activity	Energy levels	Time
Quartz SS, well orted, well rounded	Granite	Humid	Passive	High and consistent	Long
Arkose. poorly sorted, poorly rounded	Granite	Arid	Active	Inconsistent	Short
Quartz SS, muddy, angular poorly sorted,	Granite	Humid	Passive	Inconsistent	Short
Arkose. well sorted, well rounded	Granite	Arid	Active	High and consistent	Long
litharenite, poorly sorted, no mud, angular	Basalt, Gneiss other mafic rocks	Arid or temperate	Active	Inconsistent	Short

Provenance and tectonic setting of sandstones

In addition to siliciclastic sandstones, there are many hybrid sandstones. These types are : calcarenaceous, glauconitic, phosphatic sandstones, **Flysch and Molasse, Tuffs and Tuffaceous**

. In glauconitic sandstones, the glauconite occurs as sand-sized pellets (see Section 6.4.4). With phosphatic sandstones, the phosphate may be present as coprolites, faecal pellets and bone fragments.

Calcarenaceous sandstones contain up to 50% CaCO3 as carbonate grains. The latter are chiefly ooids, commonly with quartz nuclei, and skeletal frag-ments (bioclasts). Calcarenaceous sandstones occur in carbonate-producing areas where there is a large influx of terrigenous clastics. They will pass laterally into limestones or into purer sandstones towards the source of the siliciclastic sediment. Siliciclastics ce-mented by calcite have been referred to as calcareous sandstones

Flysch and Molasse: These terms are particular kinds of sandstones, associated with conditions of occurrences.

Flysch is commonly referred to as the greywacke suite, because its apparent occurrences in many geosynclines.

Molasse refers to a subgraywacke suite, a product of the erosion of essentially penecontemporaneous uplifts during orogenic cycles.

Tuffs and Tuffaceous sandstones: an important subdivision of sandstone includes those of volcanic origin . tuff is a stratified rock composed of sand-size particles expelled from volcanceous and deposited on the landsurface or in water. Tuff is composed igneous rock fragments, volcanic glass, and crystals or fragments of quartz, plagioclase, biotite and hornblende.

There are a complete gradation from tuff, composed wholly of volcanic material to detrital sandstone intermediate variations are called tuffaceous sandstones, and receive specific names based on the relative abundance of volcanic, and detrital grains. Tuffaceous variaties are typically associated with greywacke and subgraywacke.

It is suggested to call the rock " **tuff**" if it contains more than 90% volcanic material . "**tuffaceous** " rocks contain 10-50% volcanic material.

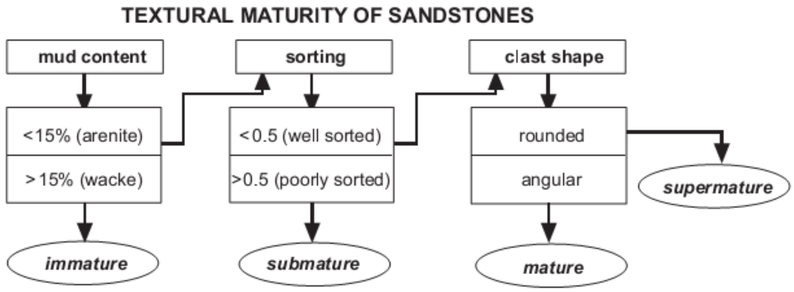
Genetic implications of sandstone composition

A-Maturity of sandstones

1-Textural maturity

The determination of the textural maturityof a sediment or sedimentary rock can best be represented by a flow diagram. Textural Maturity. The texture of sediments has two aspects: (I) description of properties, i.e., determination of the grain-size parameters and grain-size name, measurement of grain shape and description of surface features, and (2) integration of these properties into an assumed sequential development.

Using this scheme for assessing maturity, any sandstone that is classified as a wacke is considered to be texturally immature. Arenites can be subdivided on
the basis of the sorting and shape of the grains. If sorting is moderate to poor the sediment is considered to be submature, whereas well-sorted or very well-sorted sands rounded to well-rounded are considered mature.

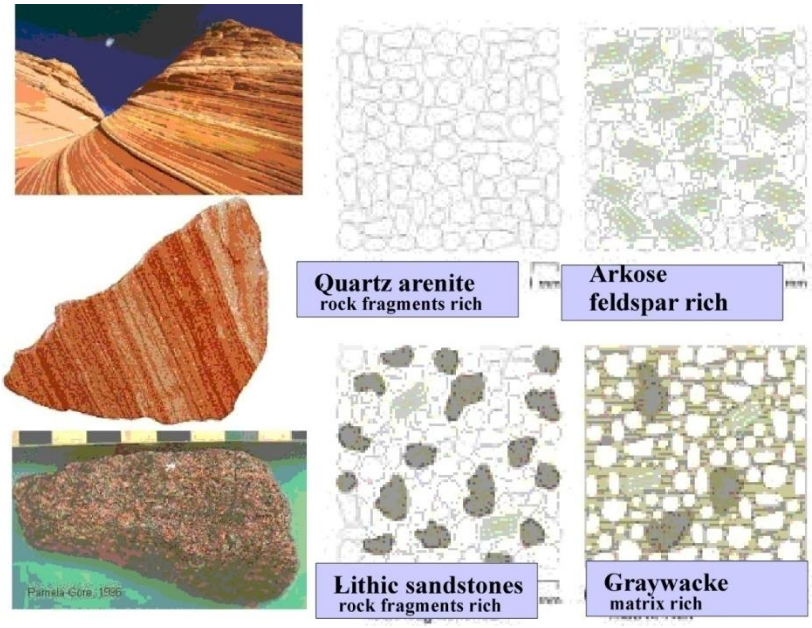


Flow diagram of the determination of the textural maturity of sandstones

From figure four stages of textural maturity. (Folk, J. Sed. Pet. 1951, 1956). This concept proposes that, as sediments suffer a greater input of mechanical energy thorugh the abrasive and sorting action of waves or currents, they pass sequentially through the following four stages:

I. Immature stage. Sediment contains over 5 percent terrigenous clay matrix;
sand grains usually poorly sorted and angular.

II. Submature stage. Sediment contains under 5 percent clay, but sand grains are still poorly sorted (a over 0.5~)) an1 clre not well rounded.



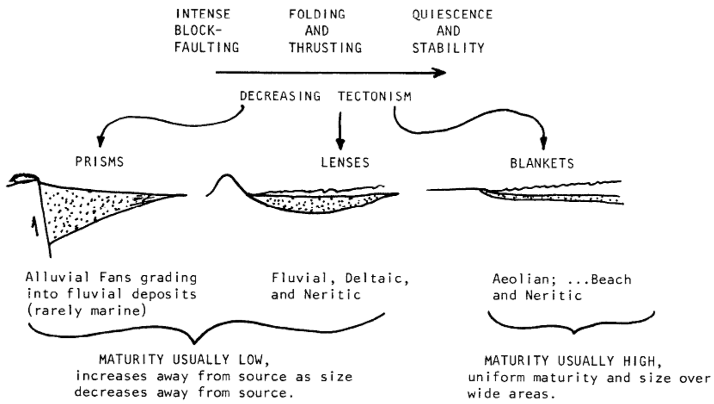
Ill. Mature stage. Sediment contains little or no clay, and sand grains are well sorted (a under &SC)), but still not rounded.

IV. Supermature stage. Sediment contains no clay, sand grains are well sorted and well rounded (Waddell roundness over .35; p over 3.0). This determination should be made, if possible, on quartz grains of medium and fine sand size

2-Mineralogical maturity

Compositional maturity is a measure of the proportion of resistant or stable minerals present in the sandstones. The proportion of highly resistant clasts such as quartz and siliceous lithic fragments in a sandstone, compared with the amount of less resistant, such as feldspars and other less resistant lithic clasts. A sandstone is compositionally mature if the proportion of quartz grains is very high and it is a quartz arenite according to the

Pettijohn classification scheme: if the ratio of quartz, feldspar and lithic fragments meant that the composition falls in the lower part of the triangle it is a mineralogically immature sediment.



Tectonism and Textural Maturity

B-provenance of a sediment

Provenance, where sediment originated. The provenance of a sediment is determined by aspects of composition that refelect the source rock and tectonic and climatic characteristics of the source area for the sediment.

1-tectonic setting

The source rock of a sediment and the tectonic setting are closely linked: the tectonic setting determines the relative abundance of different types of rock that is available for weathering and the production of clastic sediment, e.g. An arkosic sandstones (rich in feldspars) would have a source area that is rich in granite.

2-climate

Climate exerts a strong control on the type of weathering that takes place in the source area of a sediment; this, in turn, influences composition.

Cold and arid climate; predominantly physical weathering, producing abundant detrital grains (unaltered mineral grains and rock fragments). Sandstones produced in such settings will be relatively immature, depending on the source rocks.

Sandstone compositions and petrofacies (Model in Egypt)

The results of the petrographical analysis of the collected sandstone samples are represented in different ternary plots , following some authors: Folk (1980) (Q, F, L); and Dickinson (1985) (Qm/F/Lt). The sandstone framework composition of the Late Oligocene- Miocene siliciclastic allowed the identification of characteristic petrographic units (i.e. petrofacies), which coincide with the lithostratigraphic sedimentary units adopted in this study.

Four different petrofacies were distinguished on the bases of these plots and
compositional framework similarities.

Petrofacies 1

``

This petrofacies consists of quartzosedimentolithic ($Qm_{51\pm20}$ F_0 $Lt_{48\pm16}$)
sandstons with abundant quartz. They are sedarenites and quartz arenites that record erosion of the marine siliciclastics and carbonates pre-rift cover rocks. This sedimentolithic petrofacies plots in the transitional recyclic

field of the Qm F Lt diagram. This petrofacies shows a high lithic fragments content, mainly including micritic carbonate fragments, siltstone and sandstones fragments. The origin of this petrofacies is related to recycling from sedimentary source (mainly pre-rift sediments), and formed during depositions of Nakheil and Abu Ghusun formations.

3.4.2.

Petrofacies 2

``

Sandstones of this petrofacies are quartzofeldspathic Qm38±11 F17±7 Lt42±11 ``composition, and plots within the (Dissected to transitional arc) field of the Qm F Lt ternary plot. Granitic and sedimentary lithic fragments are the main constituents. Metamorphic lithic fragments are present with lesser amounts. Relatively high k-feldspars content is characteristic of this petrofacies, tending to increase towards the top of this petrofacies. Polycrystalline quartz and plagioclase are found with a lower percentages. Petrofacies 2 was formed during the deposition of Unit 1 of Lower Um Abas Member of Ranga Formation.

Petrofacies 3

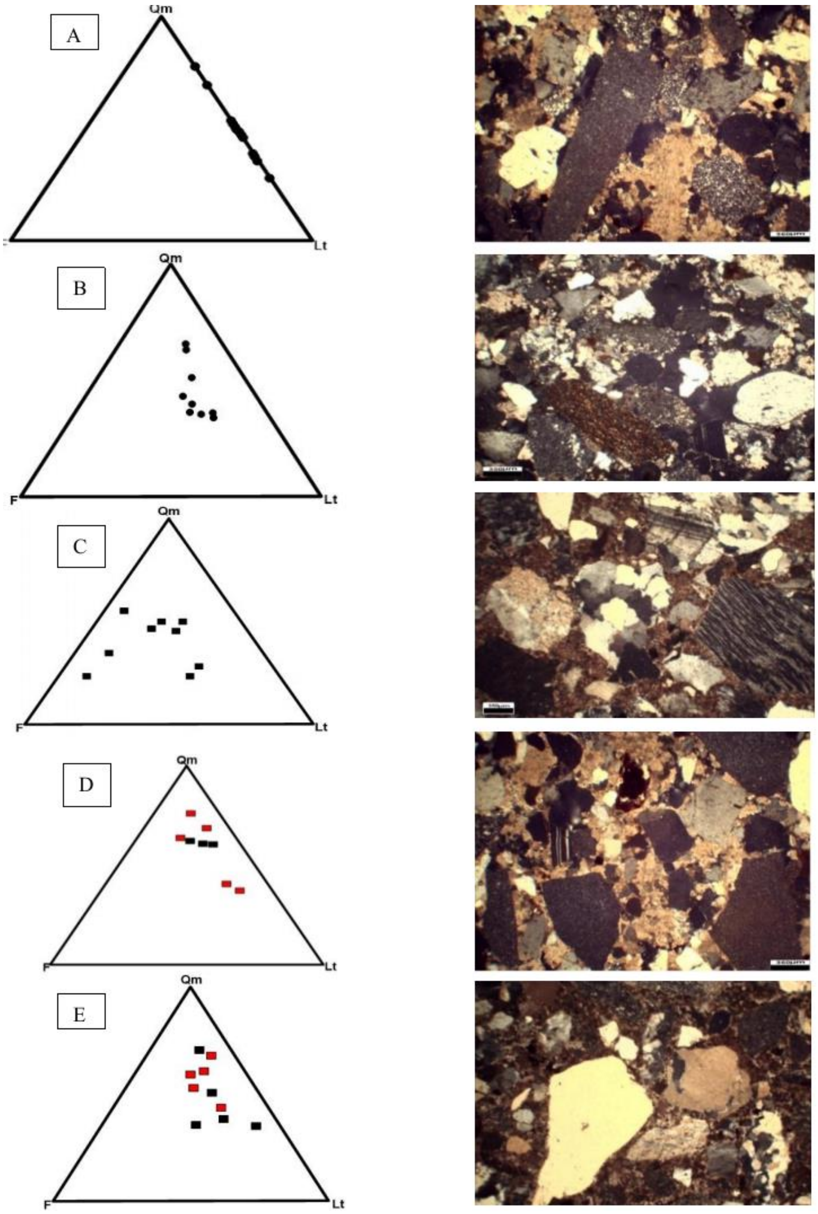
``

Sandstones of this petrofacies are feldspathic litharenite to slightly lithic arkose composition (Qm47±11 F10±12 Lt19±11), with a predominance of plagioclase over k-feldspar. This petrofacies plots within the basement uplift and dissected arc field of the Qm F Lt diagram. It shows a high lithic fragments content mainly including micritic carbonate fragments, chert, and granite grains. Granitic detritus is slightly more abundant than sedimentary. Minor amounts of metamorphic fragments are present. Petrofacies 3 accumulated during the deposition of Unit 2 of lower Um Abas Member of Ranga Formation. Petrofacies 3 is not compositionally homogeneousin the hanging wall dip-slope and faulted bounded margins of the Hamadat half graben basin. Those, It is possible to distinguish petrofacies 3 A (in hanging wall dip slope margin) from 3B (in the fault bounded margin). Petrofacies 3 A evolves from the (basement uplift to the mixed dissected to transitional arc) field of the Qm F Lt ternary plot, where as petrofacies 3B recordes an evolution from (Quartz recycled to the transitional recycled and mixed) field of the Qm F Lt plot ternary plot. Petrofacies 3 A shows higher plagioclase feldspar than 3B. K-feldspars attain an equal proportion in petrofacies 3A and 3.. petrofacies 3A shows higher granitic and sedimentary rock fragments than petrofacies 3B

Petrofacies 4

``

These sandstones are feldspathiclithicarenite with predominance of K-feldspar over plagioclase feldspars ($Qm50\pm8$ $F16\pm7$ $Lt26\pm7$). This petrofacies evolves from quartz recycl`` to the``Mixed and dissected arc``field of the Qm F Lt ternary plot . Petrofacies 4 was formed during the deposition of Upper Um Abas Member. This petrofacies shows lateral variation in composition. In the distal facies metamorphic lithic fragments are predominant, and absence in the upper part in this facies. The proximal sandstones shows relatively higher K-feldspar than distal sandstones. Sedimentary lithic fragments show equal proportions in the proximal and distal sandstones of this petrofacies. Relatively higher igneous fragments in the distal sandstones, tending to increase towards the top.



Triangular plots of sandstone compositions with Photomicrographs illustrating Late Oligocene and Miocene unroffing of syn-rift section. (A): Photomicrographs from the Late Oligocene NAkheil Formation showing- Quartzosedimentolithic Sandstons Petrofacies ``1``. (B): Photomicrograph from sandstones of unit 1 of Lower Um Abas Member showing quartzofeldspathic Petrofacies ``2``. (C): photomicrograph of sample from the Unit 2 of lower Um Abas Member showing lithicarkose of petrofacies ``3 A`` (in hanging wall dip slope margin). (D): Photomicrograph from Unit 2 of lower Um Abas Member showing feldspathiclithicarenite of petrofacies``3 B`` (in fault bounded margin). (E): Photomicrograph from Upper Miocene Upper Um Abas Member showing feldspathiclithicarenite of petrofacies

FINE -GRAINED TERRIGENOUS ROCKS
Mudrocks

Fine-grained terrigenous clastic sedimentary rocks tend to receive less attention than any other group of deposits despite the fact that they are volumetrically the most common of all sedimentary rocks types. They constitute some 45–55% of sedimentary rock successions. Mudrocks can be deposited in river floodplains and lakes, large deltas, the more distal areas of clastic shelves, basin slopes and deep-sea floors.

In terms of grain size , Clay is a textural term to define the finest grade of clastic sedimentary particles (less than 4 microns in diameter), whereas silt is between 4 and 62mm. Clay minerals are a group of phyllosilicate minerals that are the main constituents of clay-sized particles. The term **mud** (also **lutite**) loosely refers to a mixture of clay- and silt-grade material.

Mudstone, the indurated equivalent of mud, is a blocky, non-fissile rock. Shale is usually laminated and fissile (fissility is the property of split-ting into thin sheets), , which is a strong tendency to break in one direction, parallel to the bedding.

Argillite is used for a more indurated mudrock and hardened by incipient metamorphism but showing no slaty cleavage.

A sedimentary rock dominated by clay-grade material is called a claystone, and one that contains more silt-grade particles than clay is called a siltstone. Calcareous mudrocks are marls.

Clay minerals

Clay minerals commonly form as breakdown products of feldspars and other silicate minerals. They are phyllosilicates with a layered crystal structure similar to that of micas and compositionally they are aluminosilicates. The most common of these clay minerals are kaolinite, Montmorillonite, illite and chlorite.

Kaolinite is generally formed in soil profiles in warm, humid environments. Kaolinite forms under intensely leach bedrock lithologies such as granite: abundant rainfall, good drainage, acid waters; in marine basin tends to be concentrated nearshore

Montmorillonite is a product of more moderate temperature conditions in soils with neutral to alkaline pH. It also forms under alkaline conditions in arid climates.

Illite, is related to the mica group and is the most common clay mineral in sediments, derived mainly from preexisting shales, forming in temperate areas where leaching is limited

Chlorite forms most commonly in soils with moderate leaching under fairly acidic groundwater conditions and in soils in arid climates. Montmorillonite, illite and chlorite all form as a weathering product of volcanic rocks, particularly volcanic glass.

Petrographic analysis of clay minerals

There are two principal techniques for identification and interpretation of clay minerals : scanning electron microscopy (SEM) and X-ray diffraction (XRD).

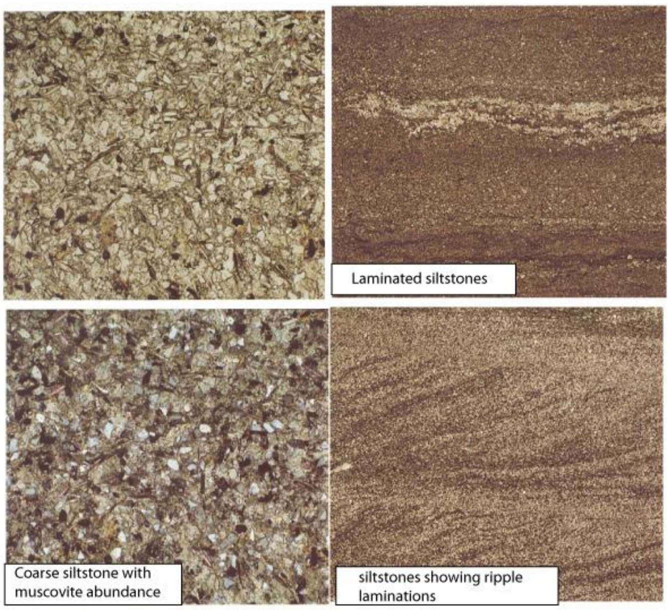
Textures and structures of mudrocks

A common texture is: 1- **preferred orientation** of clay minerals and micas parallel to the bedding. This texture is the result of deposition of clay flakes parallel to bedding and, perhaps to the effect of compaction and dewatering

<u>2- Fissility</u>, is the ability of mudrocks to split along smooth planes parallel to the stratification. The origin of fissility is due to the compaction-induced alignment of clay minerals. The degree of fissility shown by mudrocks may be related to weathering at outcrop.

<u>3-Lamination</u>, is the result mainly of variations in grain size and/or changes in composition. Size-graded laminae may be deposited from low-density turbidity and suspension currents or from decelerating storm currents, in relatively short periods of time.

4-Other sedimentary structures occur in mud rocks are cross-lamination, planar bedding, flaser and lenticular bedding, grooves and flutes, slump structures small-scale scour-and-fill structures. Some mud rocks possess massive sedimentary structures. Also desiccation cracks, rain-spot prints, sediment shrinkage formed through subaerial exposure of mud rocks.



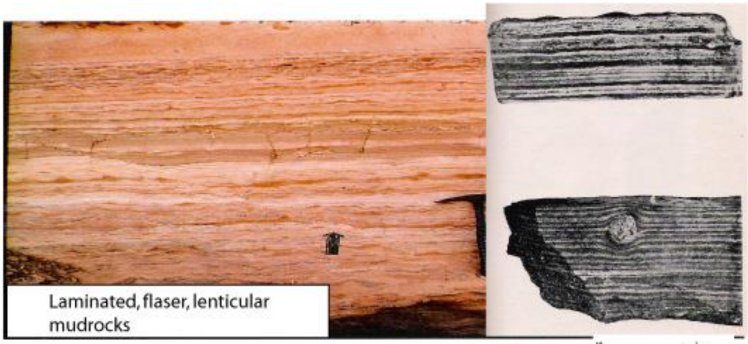
Nodules and concretions

Many mudrocks contain nodules (concretions). These are regular to irregular, spherical, ellipsoidal to flattened bodies, commonly composed of calcite, siderite , pyrite, chert or calcium phosphate. Nodules grow from pore waters within the sediment during diagenesis and may take place below the sediment–water interface. In some cases nodules formed around a nucleus composed of fossil, or as a result of the local chemical conditions. Elongate nodules may show a preferred orientation, reflecting the direction of pore-water movement. Some nodules may form before compaction during early diagenesis, and others may form after compaction of the host sediment during burial diagenesis.

The colour of mudrocks

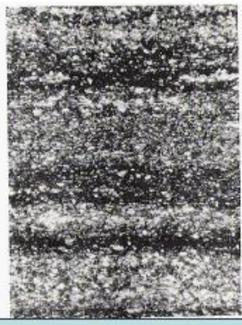
The colour of a mudrock, is a function of its mineralogy and geochemistry.

Colour is very useful in field mapping to distinguish between various mudrock units. The main controls on colour are the organic matter and pyrite content and the oxidation of the iron. The colors of muds that prevail at the time of deposition can be changed during burial diagenesis and uplift. Thus, the ferric iron (Fe_3) that characterizes red shales at the time of deposition may subsequently be reduced to ferrous iron (Fe_2) during diagenesis to yield green shales. Also, green shales can be changed to red shales after deposition if initial reducing conditions are subsequently followed by oxidizing conditions.

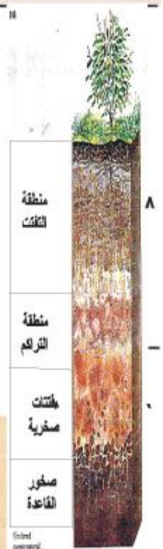


Laminated, flaser, lenticular mudrocks

Rhythmites consisting of graded Silt passing up into clay-grade material



Rhythmites consisting of alternations of silt-grade quartz and clay-organic matter.



قطاع مثالى فى التربة

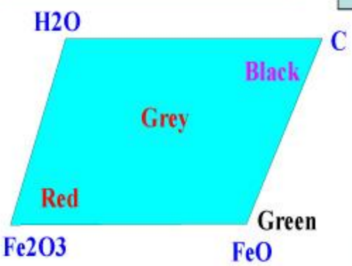
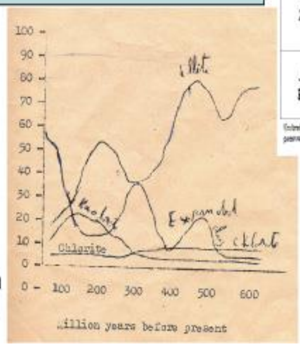


Diagram showing the relation between contaminations and colours of shales



Black colours, tends to increase in amount of organic matter and pyrite, mudrocks take on a darker grey colour and become black shales, and tend to form in relatively deep, restricted basins where reducing conditions.

Red and purple colours result from the presence of ferric oxide, hematite, occurring chiefly as grain coatings and intergrowths with clay particles. Many non-marine mudrock successions of arid-zone playa lakes and floodplains are red in colour, reflecting the dominantly oxidizing nature of the depositional and early diagenetic environment.

Green mudrocks result from the presence of ferrous iron within the lattices of illite and chlorite. The green colour may be original but in some cases it develops in mudrocks that originally were red, through the reduction of hematite by migrating
groundwaters.

Olive and yellow mudrocks, may owe their colour to a mixing of green minerals and organic matter.

Some mudrocks have **a colour mottling**. It takes in grey, may be the result of bioturbation, but in yellows/reds/browns, as the result of pedogenic processes, of water moving through a soil causing an irregular distribution of iron oxide/hydroxide and/or carbonate. Colour mottling is common in lacustrine and flood-plain muds and marls, especially those of palustrine facies.

Mudrocks and their depositional environments

Three major groups of mudrock in the geological record .These are:

1-Residual mudrocks and soils

Those formed in situ through contemporaneous processes of weathering and soil formation upon pre-existing rocks and sediments. The clays stay where they are formed, i.e. no transportation. Under favourable climatic conditions, the residual clays are enriched in hydroxide of aluminium and ferric iron. At the same time, lime, magnesium, and alkalines, and in some cases, silica are partly or completely removed, leading to the formation of the laterites, which are composed mainly of alumina and iron oxides.(e.g. **Ilian Soil** at the contact between Precambrian rocks and Nubia group, Idfo-Marsa Alam road, south Egypt , Philobbos and Abdel Rahman, 1990).

The **residual mudrocks and soils** are preserved at unconformities. Soils developed upon/within sediment**s** are characterized by the presence of calcrete (or caliche). Calcretes vary from scattered to densely packed nodules of $CaCO_3$ and are typical of semi-arid climatic areas where evaporation exceeds precipitation . They occur in many river floodplain sediments, and the clay minerals formed in these soils include palygorskite and smectite.

<u>Detrital,transported shales and mudrocks</u>

These are formed by deposition of silt and clay materials after transportation, and deposition taking place in quiet, low-energy environments. The source of the silt and clay materials are:
a-residual matters
b-products of abrasion
c-chemical and biochemical additions.

Such materials can be deposited associated with various materials such as organic matter , fossils, etc.

Classification of mudstones and shales

A- <u>Based on colour</u>

Organic-rich mudrocks and black shales

The black shales and carbonaceous and bituminous mudrocks are shales containing black organic matter, which typically contain 3–10% organic carbon. Black shales tend to form in relatively deep, restricted basins where reducing conditions prevail and abundant organic matter is preserved. They form in some shallow-water, tidal-flat and estuarine environments where organic matter is abundant. Associated with the black shales are pyrite. With an increasing organic content, organic-rich mudrocks pass into oil shales, which yield a significant amount of oil on heating. Thus, the importance of organic-rich mudrocks in the geo-logical record lies in their potential as source rocks for petroleum, if buried to suitable depths and subjected to appropriate temperatures.

The accumulation of organic matter is favoured when the circulation of water is restricted to some extent so that insufficient oxygen reaches the bottom sediments to decompose the organic material, such as in lakes , sediment-starved basins and deep-ocean trenches. As a result of the poor circulation and restriction, the water body becomes stratified and the sea or lake floor may become oxygen deficient and totally anoxic. Mudrocks deposited in an anoxic environment would contain only pelagic fossils. Example of Cretaceous black shales and organic-rich mudrocks can be found in southern Egypt.

Usually the black shales contain high concentrations of certain trace elements,
in particular Cu, Pb, Zn, Mo, V, U and As. The trace elements are adsorbed onto the organic matter and also onto the clay minerals.

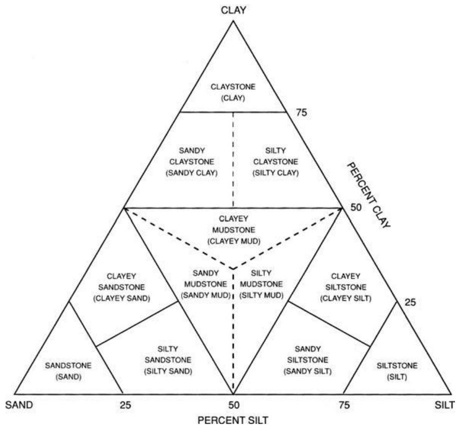
Red shales,

The red shales are shales containing more than about 15 percent iron oxides are called ferruginous shales , are characteristic of oxidizing environments. Such environments occur particularly in continental settings. Red muds accumulate also in deep-sea basins where sedimentation rates are low, and the muds are thus in contact with oxidizing bottom waters for long periods of time.

B- Based on cementing materials or chemical constituents

Calcareous shales,ferriferous shales,high-alumina shales,phosphatic shales, siliceous shales

C- Based on texture (grainsize)



Textural classification of fine-grained rocks and sediments

D - Based on structure(lamination).

Percentage clay-size constituents			0–32	33–65	66–100
Field adjective			Gritty	Loamy	Fat or slick
Nonindurated	Beds	Greater than 10 mm	Bedded silt	Bedded mud	Bedded claymud
	Laminae	Less than 10 mm	Laminated silt	Laminated mud	Laminated claymud
Indurated	Beds	Greater than 10 mm	Bedded siltstone	Mudstone	Claystone
	Laminae	Less than 10 mm	Laminated siltstone	Mudshale	Clayshale
Metamorphosed		Degree of metamorphism Low	Quartz argillite	Argillite	
		↓	Quartz slate	Slate	
		High	Phyllite and/or mica schist		

E - Based on environment of deposition.

Non-marine mudrocks

Non-marine mudrocks are represented by river floodplains , and identified by their association with fluvial-channel sandstones. In many cases the mudrocks are overbank deposits and constitute the upper part of fining-upward of the river channels. Flood plain silts and clays deposited under a semi-arid climate, including calcareous nodules (calcretes) of pedogenic origin.also, mudrocks deposited in lakes depending on the chemistry of the lake waters, organic productivity and climate. In the majority of cases the clay minerals are detrital, The accumulation of the organic matter can lead to the formation of bituminous mudrocks, black shales and oil shales .

Marine mudrocks

In the marine environment, mudrocks are deposited in five main locations: muddy coastlines; nearshore and mid-shelf mud belts; open-shelf mud blankets; basinal slopes; and basin floors.

Loess sediments

Loess is regarded primarily as an aeolian deposit. Loess derived from hot, arid, desert weathering and deflation It is a yellow- to buff-coloured clastic deposit com-posed principally of silt-sized quartz grains, generally in the size range of 20–50mm. Loess is usually well-sorted , un-stratified and unconsolidated, but it may contain shells of land snails and concretions formed around roots.

CARBONATE ROCKS

Carbonate rocks occur throughout the world in every geological period from the Cambrian to the Quaternary, and make up about one-fifth to one-quarter of all sedimentary rocks in the stratigraphic record. Both limestone and dolomite are well represented in the stratigraphic record. Dolomite is the dominant carbonate rock in Precambrian and Paleozoic sequences, whereas limestone is dominant in carbonate units of Mesozoic and Cenozoic age

Importance of carbonate rocks

1-They contain much of the fossil record of past life forms, and they provide invaluable insight into environmental conditions of the past (acted as indicators of Earth history)
2-They are used for a variety of agricultural and industrial purposes, they make good building stone, they serve as reservoir rocks for more than one-third of the world's petroleum reserves, and they are hosts to certain kinds of ore deposits such as epigenetic lead and zinc deposits.

Mineralogy of carbonate rocks

1-Carbonate components

The common carbonate minerals fall into three main groups: the calcite group, the dolomite group, and the aragonite group. Two types of calcite are recognized depending on the magnesium content: low-magnesium calcite and high-magnesium calcite. A limestone also may be dolomitized, whereby dolomite, $CaMg(CO3)_2$, replaces the $CaCO_3$ minerals and is precipitated as a cement (dolomitization).

Dolomite-group minerals differ from calcite-group minerals.They contain Mg^{2+} and/or Fe^{2+} in addition to Ca^{2+}

Aragonite is unstable at surface temperatures and pressures and in time high-Mg calcite loses its Mg. High-magnesian calcite is metastable with respect to calcite and may lose its Mg in time and alter to calcite. Thus all carbonate sediments with their original mixed mineralogy are converted to low-Mg calcite during diagenesis

2- <u>Noncarbonate components</u>

Non-carbonate minerals in limestones include silicate minerals such as quartz, chalcedony or microquartz, feldspars, micas, clay minerals, organic matter, and heavy minerals., and pyrite, hematite, chert and phosphate of diagenetic origin. Evaporite minerals, in particular gypsum–anhydrite, may be closely associated with limestones. Noncarbonate minerals are commonly separated from carbonate constituents by acid treatmen, and they are referred to as insoluble residues

Identification of carbonate minerals

It is often difficult to distinguish among these minerals in hand specimens and thin sections. Identification can be greatly aided by staining techniques. For example, aragonite is stained black with Fiegl's solution (Ag2SO4+MnSO4), whereas calcite remains unstained. Calcite is stained red in a solution of Alizarin red S and dilute HCl, whereas dolomite remains unstained. The carbonate minerals can be differentiated also by X-ray diffraction methods.

CLASSIFICATION OF LIMESTONE

GENETIC CLASSIFICATION

Genetically limestones are classified into two main groups

A) -Autochthonous B)-Allochthonous

Autochthonous limestones are formed by direct precipitation of calcium carbonate from sea water either by organic or inorganic processes. Allochthonous limestones are transported and redeposited limestones.

Autochthonous limestones

These limestones can be subdivided genetically into

1- Biochemical 2- chemical

The biochemical limestones are deposited by the action of organisms, which extract the carbonates from the water to build up their skeletons, e.g. corals, foraminifera and algae. **Bioherms or biohermal limestones** belong to the biochemical limestones. They are dome like, lens like or mound like masses built of sedimentary organisms. The most common type belonging to the bioherms are the coral reefs.

Biostromes or biostromal limestones are bedded fossil accumulations , e.g. shell beds, crinoidal limestones and algal beds.

Pelagic limestones are also formed by accumulation of organism but in contrast to the bioherms and bistromes, they are formed of pelagic organisms. The pelagic limestones are relatively rare and believed to be restricted to geosynclinal belts, where they occur as commonly thin siliceous beds. Chalks belong also to this group. They are formed of planktonic microorganisms embedded in a structureless fine grained carbonate matrix.

<u>Various chemical limestones</u>

Tufa and travertine are limestones formed by evaporation of spring and river waters. Tufa is spongy porous rocks. Travertine is more dense banded and especially common in limestone caves forming stalactites and stalagmites.

B- Detrital (allochthonous) limestones

On the basis of grain size, the limestone is divided into: Calcarenites, calcrudites and calclutites

These are mechanically deposited carbonates. Calcarenites have grain size 2-1/16 mm, whereas calcrudites are of >2mm, grain size. The calclutites have a grain size less than 1/16mm.

The detrital carbonate grains consists mainly of fossil materials, both entire and broken, pebbles and gravels of limestones and oolites.

Detrital limestones consisting of wholly or nearly wholly sorted coarse debris are coquinas. If the shell debris are fine the term microquina is applied.

The calcarenites in which the oolites are chief ingredient are termed oolitic calcarenites or oolitic limesone. Some calcarenites contain small rounded fine grained.such grains may be or pseudo-oolites or may beformed penecontemporaneously be aggregation from calcareous silts.

Calcarenites may be found mixed in any proportion, with quartz sand grains. If quartz grains are more than 90% the rock is termed **calcareneoceous orthoquartizites**. If the quartz is less than 50% the rock is termed **arenaceous calcarenite.**

Calcarenites may contain grains of glauconite and in this case, they are described as glauconitic calcarenites.

The calcilutites are allochthonous limestones of grain size less than 1/16mm.

PETROGRAPHIC CLASSIFICATION

This type of classification of limestones is now widely used among the sedimentary petrologists. It is based mainly upon Folk (1959) classification. It is based upon the nature and amount of the coarser components (allochems) in contrast with the finer grain matrix.

Components of limestones

Limestones are very varied in composition but broadly the components can be divided into four groups: (i) non-skeletal grains, (ii) skeletal grains, (iii) micrite and (iv) cement.

Non-skeletal grains

Ooids and pisoids

Ooids are spherical–subspherical grains, consisting of one or more regular concentric lamellae around a nucleus, usually a carbonate particle or quartz grain. The term ooid has been restricted to grains less than 2 mm in diameter and the term pisoid is used for similar grains of a larger diameter. Ooids that consists of numerous laminae are called normal ooids. If only one lamella is developed around a nucleus, is called superficial ooid. Composite ooids consist of several small ooids enveloped by concentric lamellae. The ooids form in agitated waters where they are frequently moved as sand waves, dunes and ripples by tidal and storm currents, and wave action. and high water agitation. Ooids can form in high-energy locations in lakes. Ooids and pisoids can also form in quieter-water marine locations, such as in lagoons and on tidal flats. Quiet-water ooids are more likely to have a radial structure, less-spherical form.

Pisoids are coated grains that resemble ooids; but are commonly larger (several millimeters to centimeters). Most are nonmarine in origin, and formed in the groundwater vadose zone(are called vadoids).

Intraclasts

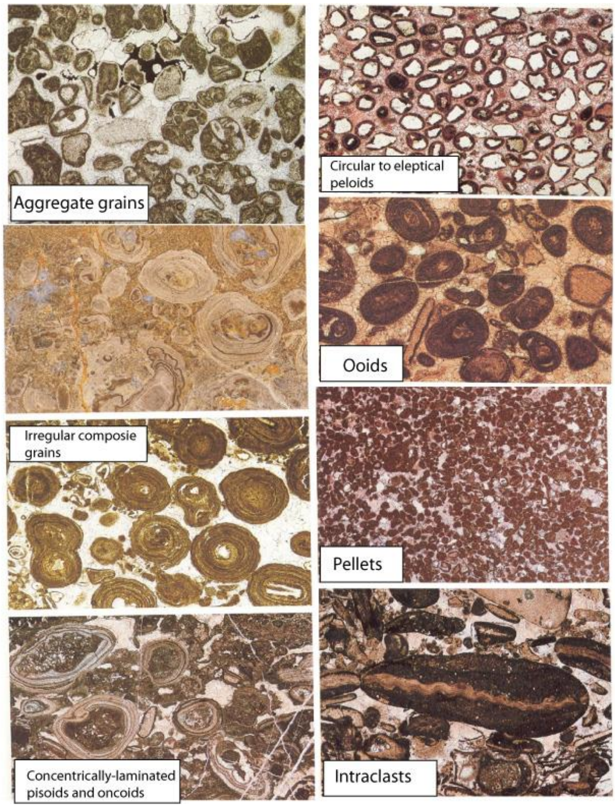
Intraclasts originate within a depositional basin by disruption of desiccated and cracked lithified calcareous muds, and redeposited to produce lime-mud clasts fragmentation.Also they produced from erosion of lithified beachrock within the intertidal and supratidal zones.

Aggregate grains and lumps

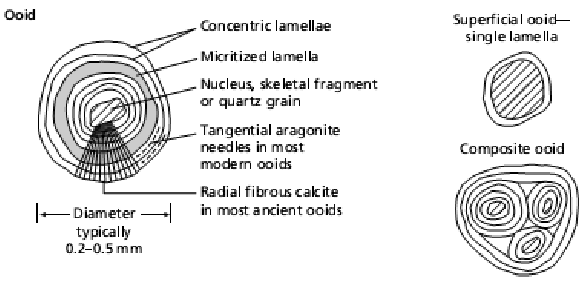
Aggregate grains and lumps are irregularly shaped carbonate grains that consist of two or more carbonate fragments joined together by a micritic (lime-mud) matrix. or that form in situ by agglutination of adjacent carbonate grains such as peloids, ooids, and skeletal fragments. They have produced from protected shallow subtidal areas, brackishwater of tidal zones and sea-marginal hypersaline pools.

Peloids

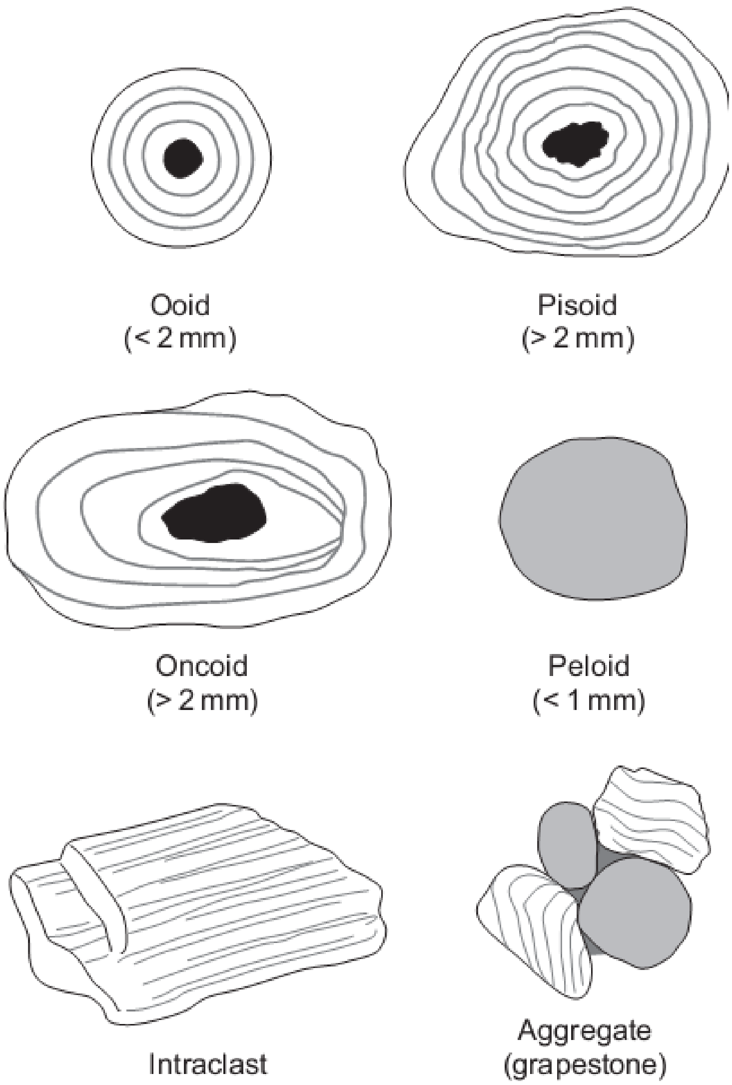
Peloids are spherical, ellipsoidal or angular grains, composed of microcrystalline carbonate, but with no internal structure. The size of peloids ranges from about 0.05 to 0.20mm. Many peloids, are to be of fecal origin, and called pellets. They are most common in the sediments of protected environments such as lagoons and tidal flats. Most pellets appear to be produced by organisms living in quiet, marine water with muddy bottoms. Thus, pellets in ancient limestones occur most commonly in muddy limestones (micrites), and their presence suggests deposition in low-energy environments. Not all marine peloids are fecal pellets. Some are believed to form by carbonate encrustation around filaments of cyanobacteria, endolithic algae, or fragments of other algae.



Aggregate grains
Circular to eleptical peloids
Ooids
Irregular composie grains
Pellets
Concentrically-laminated pisoids and oncoids
Intraclasts



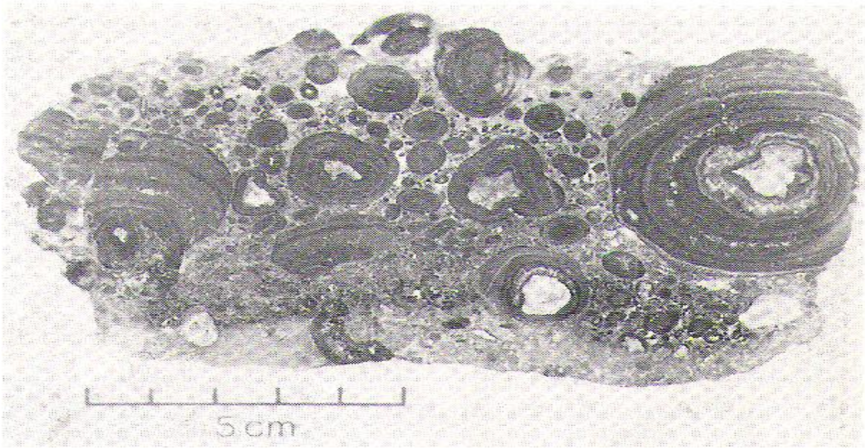
Ooid
Concentric lamellae
Micritized lamella
Nucleus, skeletal fragment or quartz grain
Tangential aragonite needles in most modern ooids
Radial fibrous calcite in most ancient ooids
Diameter typically 0.2–0.5 mm
Superficial ooid—single lamella
Composite ooid

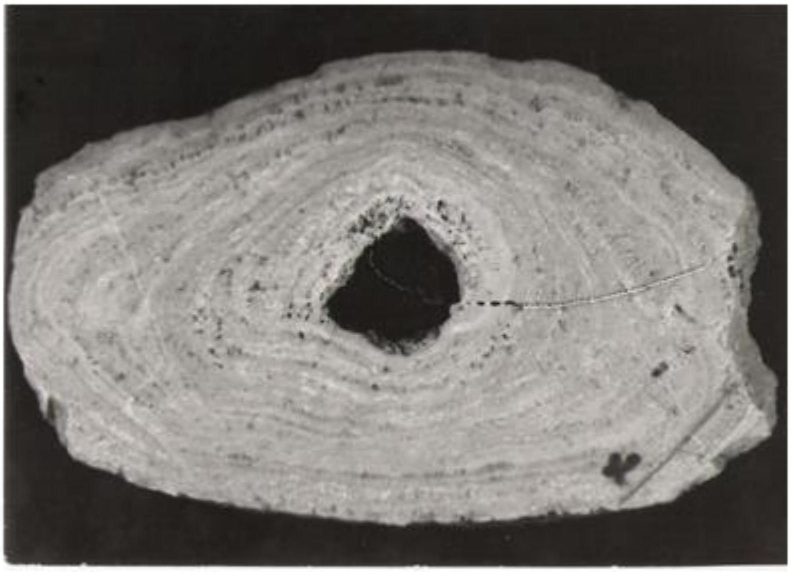


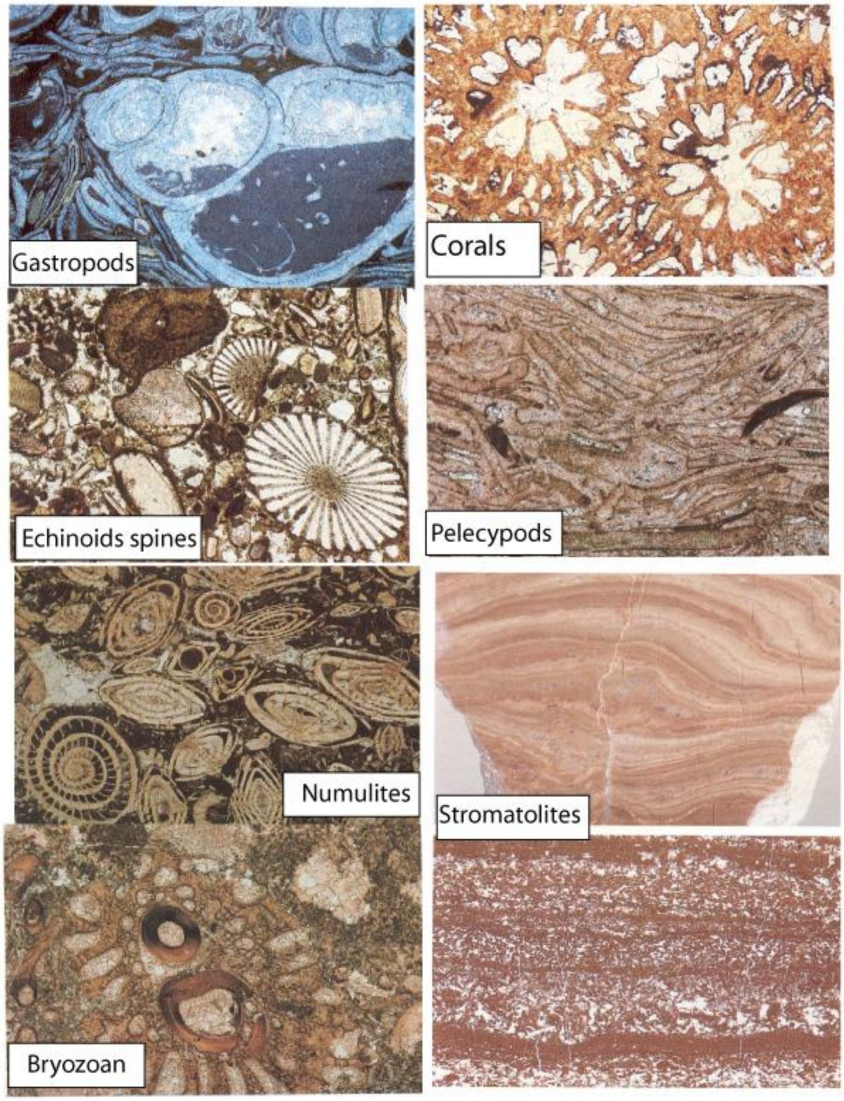
The principal non-skeletal grains in limestones.

Oncoids

Oncoids are more irregular in shape than ooids, and with more irregular laminae. Oncoids are commonly ranging from <2 mm to >10 mm. They form in both nonmarine and marine environments. Most oncoids that form through the activities of encrusting organisms are a type of stromatolite .Oncoids are surrounded by layers composed mainly of fine micrite. These layers form through the activities of organisms such as cyanobacteria that cause the formation of successive layers around the nucleus. The layers may be wavy or crinkly.







<u>Skeletal grains (bioclasts</u>)

Skeletal grains constitute an important part of the coarser components in carbonate rocks. The main skeletal grains are as follows:

Bivalves

The bivalves are a very large group with species occupying most marine, brackish and freshwater environments. Certain bivalves, such as oysters, may form reef-like structures. The majority of bivalve shells are dissolved out completely to leave a mould, which subsequently may be filled by calcite. Bivalve shells consist of several layers of specific internal microstructure, composed of micron-sized crystallites. Bivalve fragments in thin-section will be seen as elongate, rectangular to curved grains.

Gastropods

Gastropods are dominant in shallow-marine environments. The majority of gastropods have shells of aragonite with similar internal microstructures to bivalves.

Brachiopods

Brachiopods are particularly common in limestones of shallow-marine origin. The common structure is a very thin, outer layer of calcite fibres orientated normal to the shell surface, and a much thicker, inner layer of oblique fibres.

Corals

Corals require shallow, warm and clear seawater. Identification of coral is based on such internal features as septa.

Foraminifera

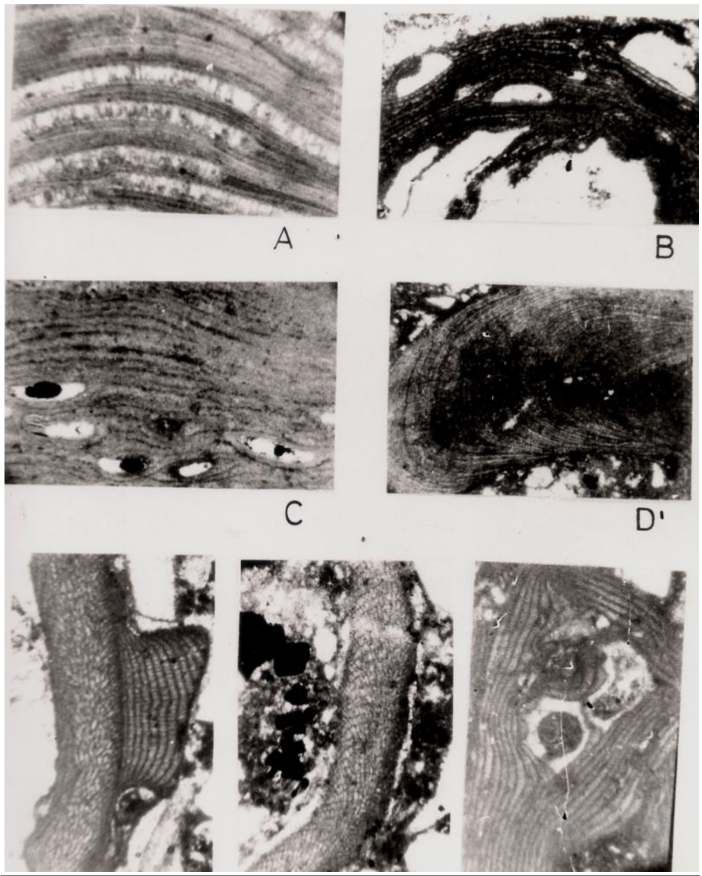
Planktonic foraminifers dominate some pelagic deposits, such as the Globigerina oozes. Benthic foraminifers are common in warm, shallow seas, living within and on the sediment, and encrusting hard substrates.Foraminifera are very diverse in shape but in section many common forms are circular to subcircular with chambers. The test wall is dark and micrgranular in many thin-walled foraminifers such as the miliolids, and fibrous in larger, thicker species, such as the rotali-ids, nummulitids and orbitolinids.

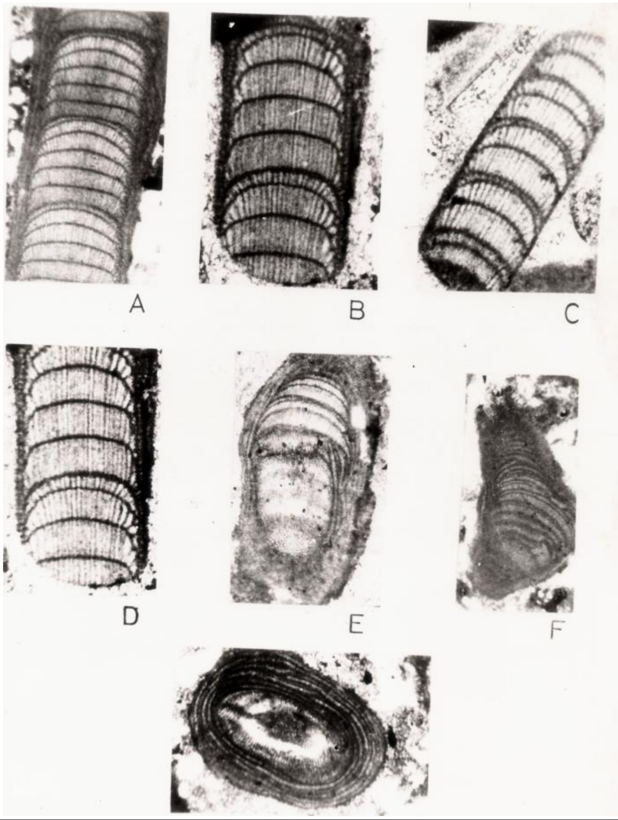
Algae

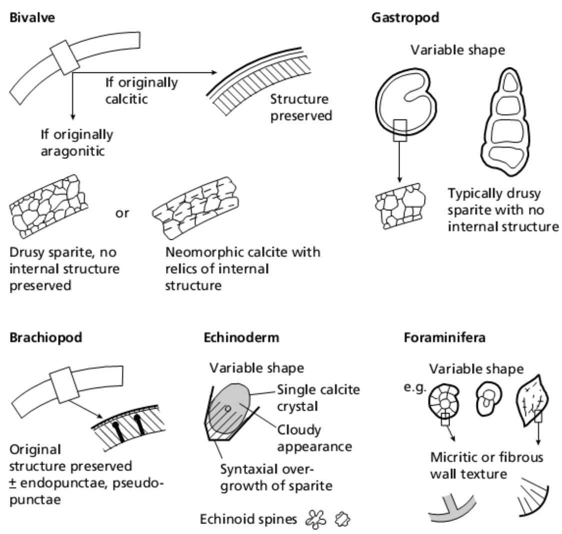
Four groups of algae are important: red algae (Rhodophyta), green algae (Chlorophyta), yellow–green algae (Chrysophyta) and cyanobacteria (formerly blue–green algae).

Rhodophyta (red algae)

Many of the coralline algae encrust substrates and if this is a pebble or shell then nodules, referred to as rhodoliths, develop. Encrustations may be massive and rounded, or delicately branched, depending on ecological factors. One of the most important roles of these red algae is in coating, binding and cementing the substrate, particularly in modern reefs.







Typical thin-section appearance of bivalve, gastropod, brachiopod, echinoderm and foraminiferal skeletal grains in limestone

Micrite

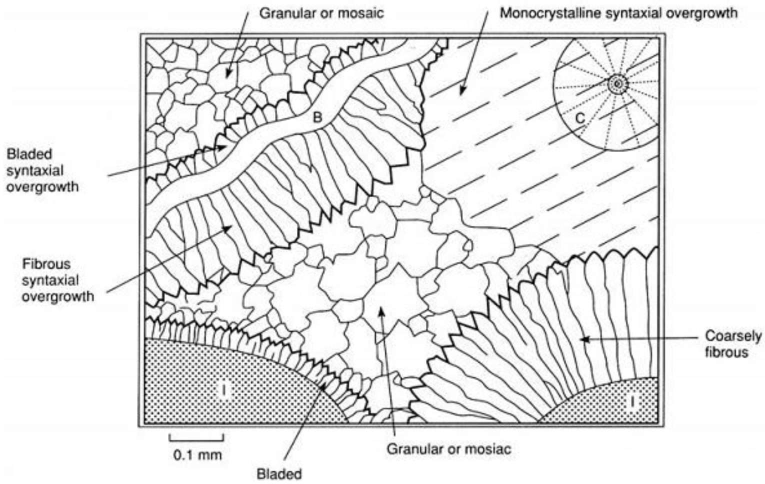
The term micrite is a microcrystalline calcite; and very fine-grained carbonate sediment. Micrite has a grayish to brownish, subtranslucent appearance under themicroscope . It is generally easily distinguished fromcarbonate grains by its finer size and from sparry calcite, which is coarser grained and more translucent. The presence of substantial micrite in a limestone is commonly interpreted to indi-cate deposition under fairly low-energy conditions.

Sparry calcite

Crystals of sparry calcite are large (0.02–0.1 mm) compared to micrite crystals and appear clear or white when viewed in plane light under a polarizing microscope. They are distinguished frommicrite by their larger size and clarity and fromcarbonate grains
by their crystalline shapes and lack of internal microstructures.

Sparry calcite cement is particularly common in grain-rich limestones, such as oolites, that were deposited in agitated water that prevented micrite from filling pore spaces. Therefore, as mentioned, the presence of significant amounts of sparry calcite cement in a limestone is commonly interpreted to indicate deposition of the limestone in agitated water.

Sparry calcite can form a variety of cementation fabrics, and several distinctive types of cement are recognized. The most common types aregranularormosaic cement, which is composed of nearly equant crystals;fibrous cement, either coarsely or finely fibrous; bladed cement; and syntaxial cement(overgrowths). Euhedral to anhedral, bladed or coarsely fibrous crystals that are oriented perpendic-ular to carbonate grain surfaces may display an increase in grain size toward the center of the pore or cavity and a concomitant change to more equant, granular crystals. This distinctive pore-filling fabric is commonly referred to **as drusy cement**.



Some important types of sparry calcite cement fabrics in limestones. The change from bladed crystals to larger, granular crystals in the lower left corner of the figure illustrates“drusy’fabric.

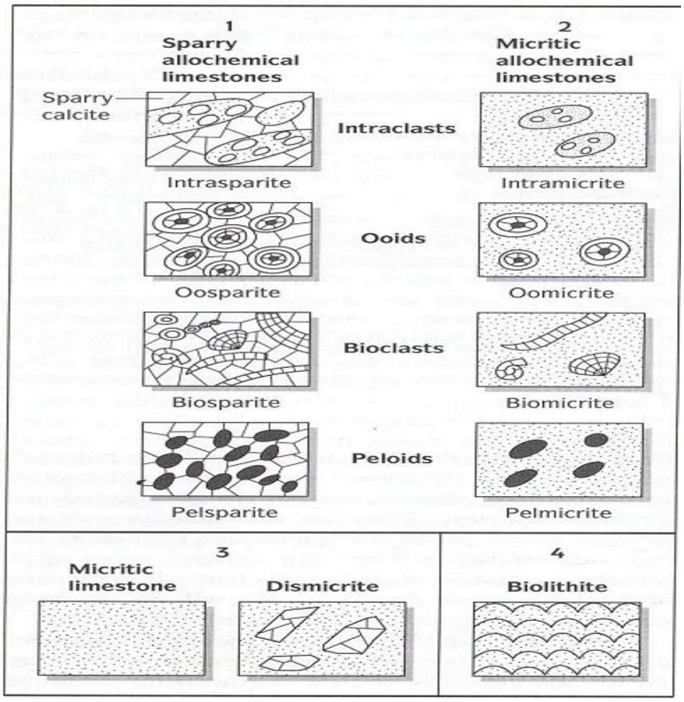
According to the different frequencies of the above mentioned components, two classification systems are currently used, each with a different emphasis.

1- The classification scheme of R.L. Folk, based mainly on composition, distinguishes three components: (a) the grains (allochems), (b) matrix,
chiefly micrite and (c) cement, usually drusy sparite. An abbreviation for the grains (bio—skeletal grains, oo—ooids, pel—peloids, intra—intraclasts) is used as a prefix to micrite or sparite, whichever is dominant. Terms can be combined if two types of grain dominate, as in biopelsparite or bio-oosparite. Terms can be modified to give an indication of coarse grain size, as in biosparrudite or intramicrudite. Other categories of Folk are biolithite, referring to a limestone formed in situ, such as a stromatolite or reef-rock; and dismicrite, referring to a

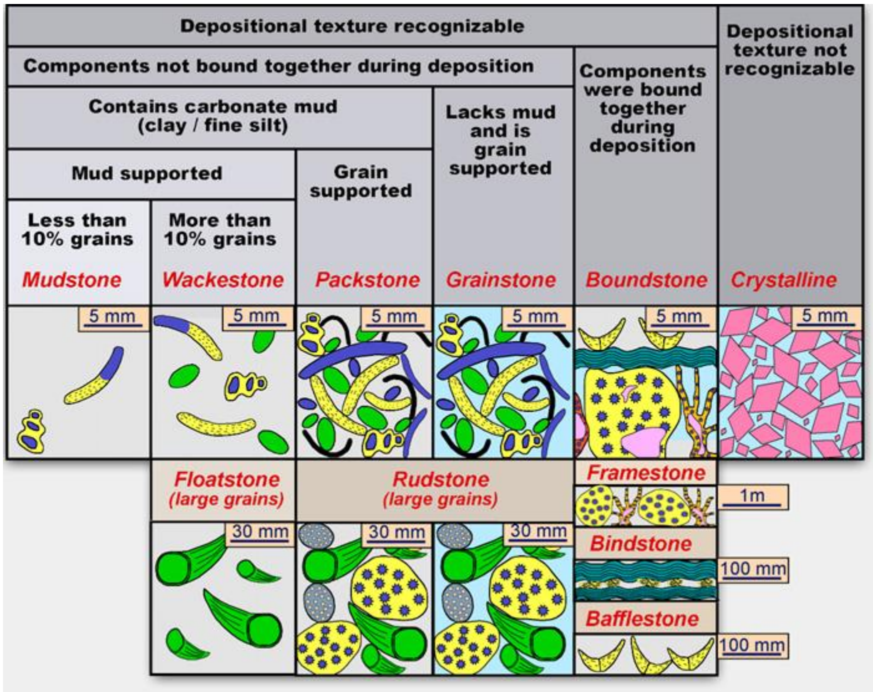
micrite with cavities (usually spar-filled), such as a birdseye limestone.

Interstitial Material	**Fosil or ossil fragments**	**oolites& Pisolites**	**Pellets**	**Intraclasts**	**Lumps**
	Biogenic L.S.	Oolitic L.S.	Pellet L.S.	Intraclast L.S.	Lump L.S.
Micrite	Crinoid **Biomicrite**	Oolitic **micrite**	Pellet **micrite**	Intraclast **micrite**	Lump **micrite**
Sparite	Coral (etc.) **Biosparite**	Oolitic **sparite**	Pellet **sparite**	Intraclast **sparite**	Lump **sparite**

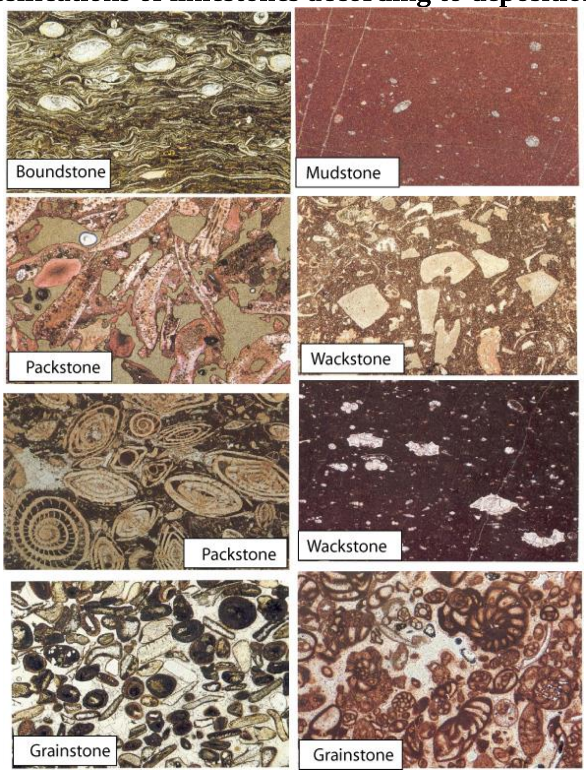
Folk's classification (1962)



2- The classification of R.J. Dunham divides limestones on the basis of texture into: grainstone, grains without matrix (such as a bio- or oosparite); packstone, grains in contact, with matrix; wackestone, coarse grains floating in a matrix; and a mudstone, micrite with few grains. Additional terms of A.F. Embry & J.E. Klovan give an indication ofcoarse grain size (floatstone and rudstone), and of the type of organic binding in boundstone during deposition (bafflestone, bindstone and framestone).



Dunham (1962) and Embry and Klovan (1972) classifications of limestones according to depositional texture



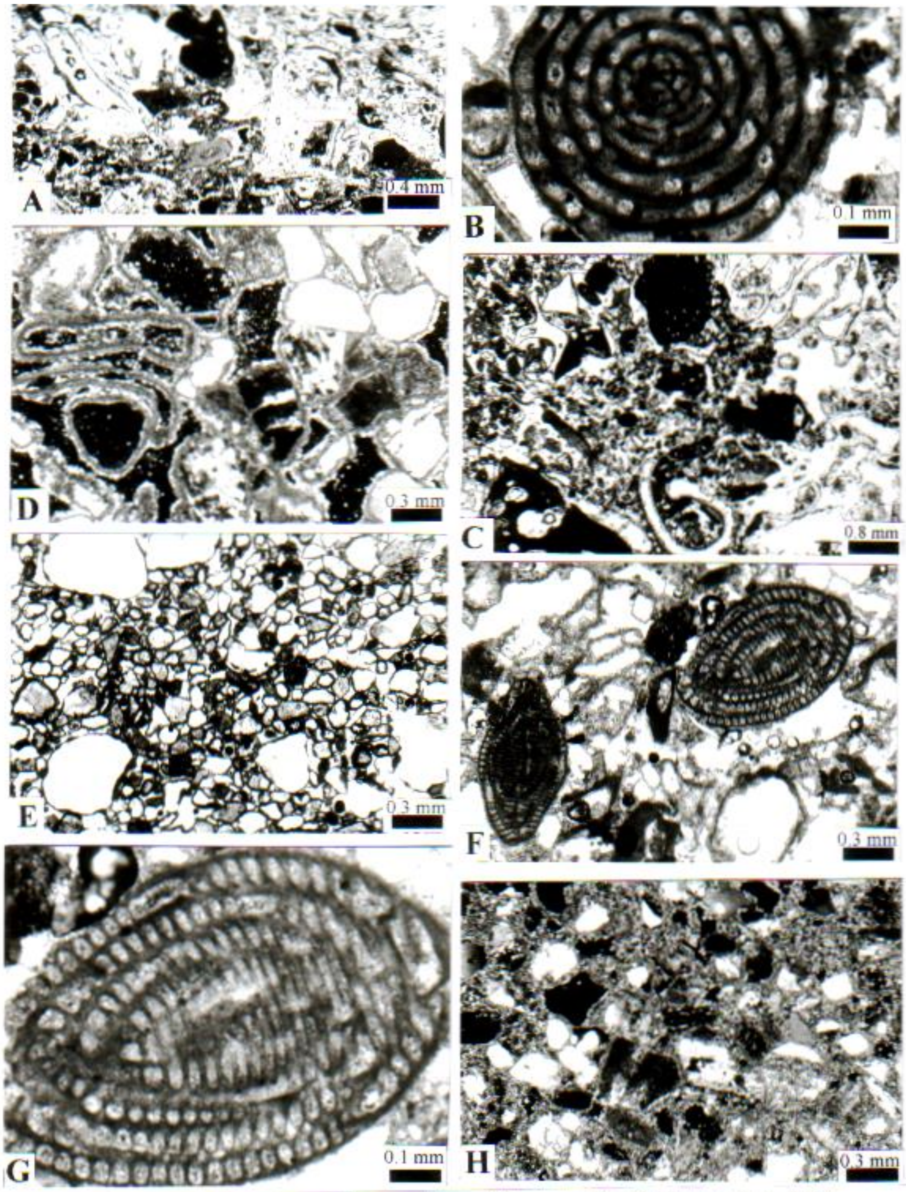
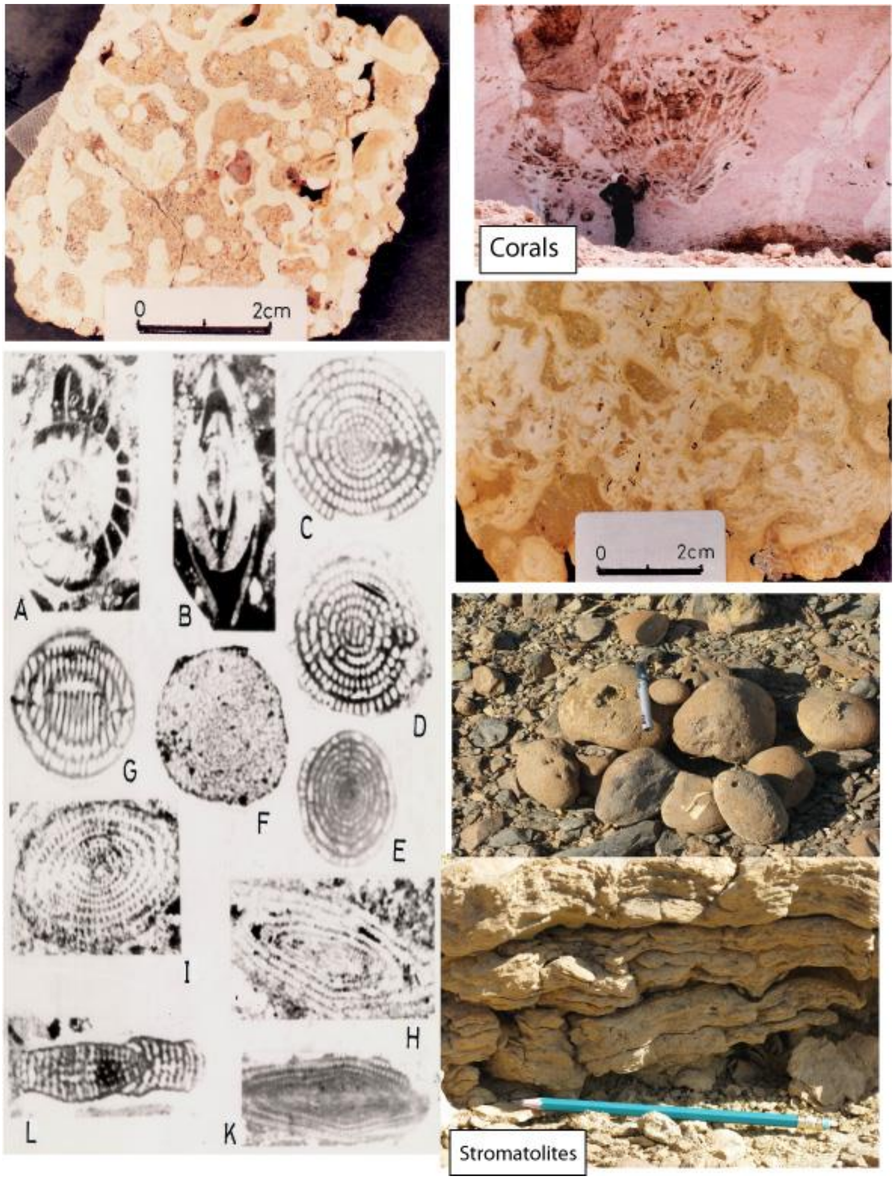


Fig. (8): Photomicrographs of microfacies associations of Shagra and Samadai formations. A- Algal bioclastic bondstone, Dashet El Dabaa Member, Wadi Wizr, B- Enlarged *Borelis sp. Cf. B. costulatus*, Dashet El Dabaa Member, Gabal Gazerat El Hamra, C- Coralgal bioclastic packstone, Dashet El Dabaa Member, Wadi Abu Dabbab, D- Molluscan bioclastic grainstone, Sharm El Arab Member, Gabal Gazerat El Hamra, E- Quartz arenite, Sharm El Arab Member, Gabal Gazerat El Hamra Foramineferal bioclastic grainstone (*Borelis Schlumbergeri*), Sharm El Arab Member, G- Enlarged *Borelis Schlumbergeri* of Fig. 8F, Sharm El Arab Member, H- Feldspathic arenite, Um Geradiat Member, Wadi Marsa Shuni



Algal laminites and stromatolites

These are formed by cyanobacteria . These organic mats occur on sediment surfaces from marine and nonmarine environments, from moderate depth subtidal through to supratidal marine areas, and fresh to hypersaline lakes and marshes. They form planar sheets, columns and domes. The cyanobacteria has filamentous nature, results in the trapping and binding of sedimentary particles to produce a laminated sediment, a microbialite or stromatolite. The lamination in many modern intertidal mats consists of couplets of dark organic-rich layers alternating with light, sediment-rich laminae. Microbial filaments may be to distinguish them from laminae deposited purely by physical processes. There may be evidence of desiccation—broken laminae, intraclasts and laminoid fenes-trae.

Microbial mats give rise to a range of laminated structures (microbialites). The simplest are planar stromatolites (or microbial laminites). They typically develop on protected tidal flats and so may show desic-cation polygons and contain laminoid fen-estrae and evaporite minerals or their pseudomorphs. Domal stromatolites, where the laminae are continuous from one dome to the next, occur on the scale of centimetres to metres. Columnar stromatolites are individual structures, which may be several metres high

Non-marine carbonate sediments

1- Lacustrine limestones

Two types of Lacustrine carbonates: (a) inorganic, (b) algal/microbial precipitates. Inorganic precipitation, producing lime muds, mostly takes place through evaporation, but CO 2 loss, as a result of plant photosynthesis or pressure–temperature changes, and mixing of fresh stream or spring water with saline lake water, also causes carbonate precipitation (e.g. Tufa mounds)

Also the activities of algae, bacteria and microbes produced Lime muds. and the formation of stromatolites. Oncoids also occur, as well as bivalves and gastropods.

Lacustrine carbonates are arranged in a similar facies pattern to their marine counterparts. Stromatolite ‘reefs’ and ooid shoals occur in more agitated, shallow waters, with lime muds occurring shoreward on littoral flats and in protected bays, and in the central deeper parts of lakes.

Shallow-water, muddy lacustrine facies are commonly modified by pedogenesis and the term **palustrine** is used for these deposits, which usually are nodular and mottled. They are widely developed in the Oligocene, Red Sea , and in the Pleistocene lakes in the Nile Valley (Mahran, 1999; Mahran and Hassan, 2012).

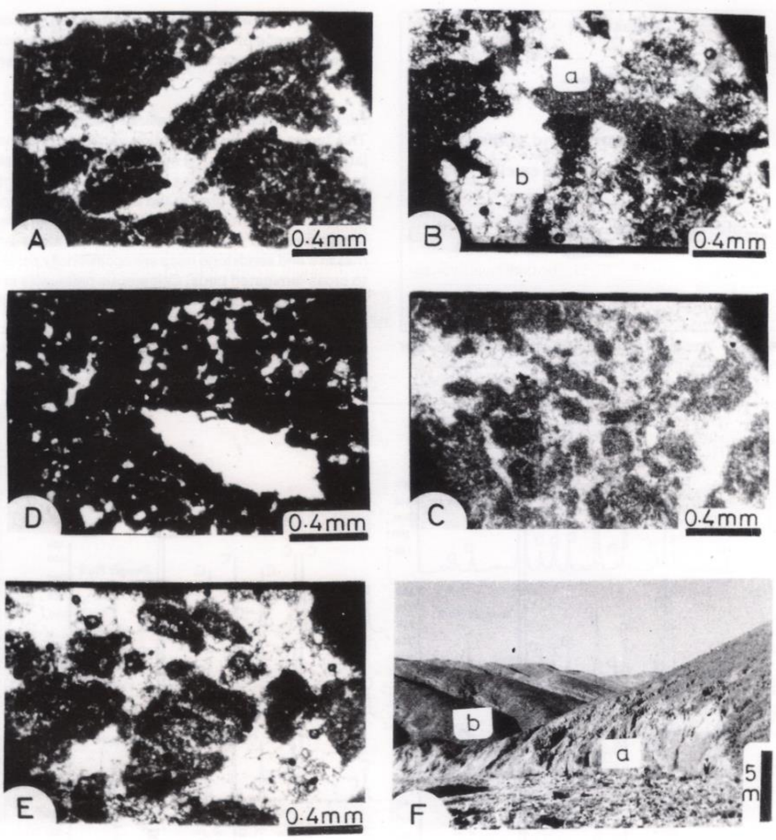


Figure 7. Marginal palustrine carbonate facies associations. (A) Photomicrograph of limestone grainstone with elongate desiccation cracks. (B) Large cavities filled with subangular clasts (a) and sparry calcite (b). (C) Photomicrograph of rounded pellets in filled cavities. (D) Photomicrograph of pseudo-microkarst cavities and fenestral birds-eye structures. (E) Photomicrograph of intraclastic grainstone cemented by sparry calcite. (F) Field view showing fine siliciclastics of distal alluvial facies (a) at the foot of marginal palustrine carbonate facies (b).

2- Calcrete or caliche

In many parts of the world where rainfall is between 200 and 600 mm year and evaporation exceeds this precipitation, calcareous soils are formed (**Calcrete or caliche)**. They occur in river floodplain and in aeolian, lacustrine and colluvial deposits, and in marine sediments too, should they become subaerially exposed.

Calcrete occurs in several forms, from nodules to continuous layers, with massive, laminated and pisolitic textures. Many calcretes form in the upper vadose zone through a perdescensum process of dissolution of carbonate particles in the upper A horizon of the soil profile and reprecipitation in the lower B horizon.

Model in Egypt

(Lacustrine carbonate)

Within Hamadat half graben basine the Lower Um Abas Member is subdivided stratigraphically into two units: Unit 1 and Unit 2. In Unit 1 the lacustrine facies is dominated by carbonate-fill channels deposits. They comprise irregularly bedded, carbonate bodies that displayclear channel geometries. This facies repeated vertically, and terminates each cycle . Thickness of the carbonate channels average 0.5 m and the width reaches up to 50 m. Lower surfaces of the channels are erosive on underlying conglomerates and/or sands, while the upper surfaces are commonly planar. Channel fill consists mainly of oncoids, bioclasts, and occasionally encrusted by stromatolites. Locally the channel components include large amounts of gastropods sp. Onclites are composed almost entirely of variable sizes ranging from a few millimeters to several centimeters in maximum dimension up to 30 cm in diameter. Most oncolites have an obvious nucleus; the most common nucleus is the silicified carbonate clasts. The carbonate-fill channels is mainly distributed along the eastern margin of the basin forming a NW-SE stretching fringe which extends parallel to the basin margin.

Interpretation

The laminations of the internal structure of these oncoids confirms their algal origin. The occurrence of stromatolites in highly siliciclastic lakes seems to be restricted to shoreline and nearshore environments, and can be used to locate ancient lake margins. The localized of carbonate-fill channels in the eastern marginal part of the basin suggests that these areas were relatively more subsiding zone allowing preferential marginal drainage flows within the lake system. This evidenced by the incision of the channels and may be also related to pulses of subsidence.

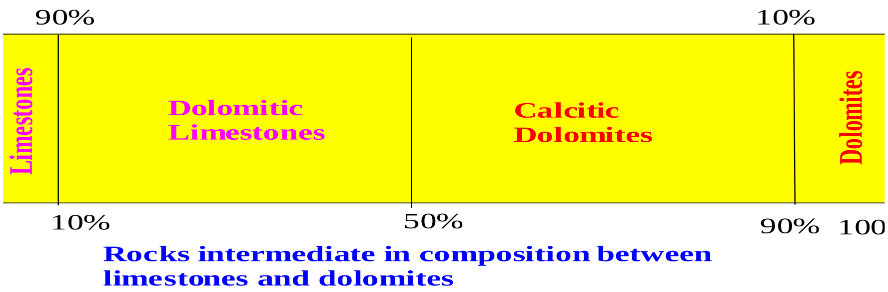
Dolomite

Dolomite is a calcium magnesium carbonate mineral . Dolomite forms when Mg in pore water replaces some of the calcium present in limestone. Dolomites abundance increase with age.

Theoretically, dolomites can form in three different ways: (1) by dolomitization, which is the replacement of CaCO3 by CaMg(CO3)2, (2) by dolomite cementation, which is precipitation of dolomite from aqueous solution in primary or secondary pore spaces, and (3) by precipitation from aqueous solution to form sedimentary deposits ("primary dolomite"). The primary is restricted to some evaporitic lagoonal and/or lacustrine settings. The great bulk of dolomite in the geologic record apparently formed by dolomitization (replacement).

Mineralogy of dolomites

Three minerals belong to the dolomite group: dolomite [CaMg $(CO_3)_2$], and ankerite [Ca(Mg, Fe,)$(CO_3)_2$] and Magnesite Mg CO_3



Classifications of dolomites

1- **Based on Ca ratio**

a- **Stoichiometric dolomite:(ideal dolomite**) in which calcium and magnesium have equal molar proportions. Ideal dolomite is rare in the geologic record.

b- **B- Protodolomite** is particularly modern or Holocene dolomites , poorly ordered with an excess of calcium

2- **Based on texture**

A- Planar dolomite (or idiotopic dolomite) subdivided into four subcategories

i-Planar–euhedral dolomite is made up of packed but crystal-supported,well-formed rhombs. The intercrystalline spaces among crystals may be filled with another mineral such as calcite, or the spaces may be empty (porous). The texture of porous dolomites of this type is sometimes referred to as sucrosic (sugary).

ii-**Planar–subhedral dolomite** has subhedral to anhedral crystals and very low porosity. Dolomites of this type have straight boundaries.

iii- Planar void-filling dolomite consists of euhedral dolomite crystals with terminations that project into open spaces. Such dolomite may be a cement, but it may form also by replacement of the margins of a carbonate grain.

iv**- porphyrotopic dolomites** (matrix-supported dolomites) form by replacement of a precursor cement. Some dolomite rhombs may appear to “float” in a limestone (micrite) matrix.

B- Nonplanar–anhedral dolomite (xenotopic dolomite). It is divided into three subcategories

i- Nonplanar-anhedral dolomite consists of anhedral grains that lack well developed crystal faces. These anhedral grains are closely packed, with curved, lobate, or irregular crystalline boundaries.
ii- Nonplanar void-filling dolomite is irregular-shaped or saddle-shaped dolomite that fills open space.
Saddle dolomite crystals of cement origin have long, curved edges leading to pointed terminations.
iii- Nonplanar–porphyrotopic dolomite is similar to planar-porphyrotopic dolomite except that the crystals are mainly anhedral. The crystals probably form by replacing micrite.

C-Zoned dolomite

Many rhombic dolomite crystals have a cloudy, rhombic central zone surrounded by a clear rim. These crystals are often referred to as **zoned dolomite**. The "zoned crystals" may form by replacement of a CaCO3 precursor such as a micritic limestone, or they may grow into open pore space where they form within a precursor limestone, the cloudy centers represent replacement of the precursor CaCO3. The clear rims must have formed in empty pore space around the margins of the cloudy rhombs.

3- **Based on the timing and nature of dolomite formation**:

i-Syndepositional (Penecontemporaneous) dolomites

Dolomites can form penecontemporaneously, while the host sediments are still in their original depositional setting; that is, they form under the geochemical conditions of the depositional environment.

They form primarily in shallow-marine to supratidal environments and mainly by direct precipitation from normal or evaporated seawater. They occur as thin layers and lenses in sabkhas, salinas, and evaporative lagoons/lakes, and as supratidal crusts and fine-crystalline cements and replacements in peritida sediments

The penecontemporaneous dolomite is best recognized by its close association with intertidal-supratidal features, such as fenestrae (birdseyes), stromatolites, desiccation cracks and flakes, evaporates, and perhaps a restricted range of fossils indicating hypersalinities. Petrographic features are a fine grain size and some preservation of fossils and structures

ii- Postdepositional dolomites form after deposition has ceased and the host carbonates have been removed from the zone of active sedimentation by sediment progradation, burial, uplift, eustatic sea-level change or any combination of these factors. Postdepositional dolomites form at various burial depths, ranging from a few meters to thousands of meters

Dolomitization

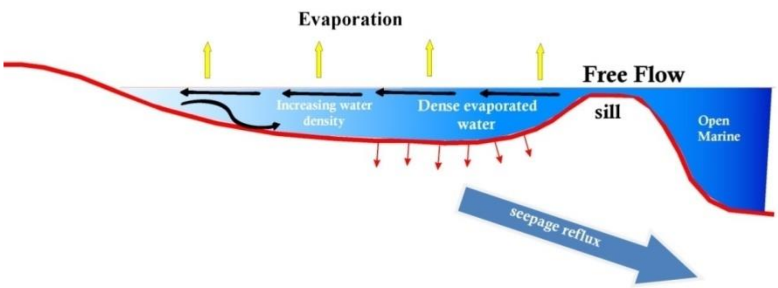
Dolomite is thought to form when the calcite ($CaCO_2$) in carbonate mud or limestone is modified by magnesium-rich groundwater. The available magnesium facilitates the conversion of calcite into dolomite($CaMg(CO_2)_3$). This chemical change is known as “dolomitization.” Dolomitization can completely alter a limestone into a dolomite or partially alter the rock to form a "dolomitic limestone."

Models of Dolomite Formation

The following three hypotheses and well established hydrologic models and settings were given to describe the origin of the common types of diagenetic dolomites.

1- Brine- Reflux model

Reflux dolomitization, which occurs in hypersaline environments, In this instance , seawater in a restricted lagoon,lake and shallow water settings evaporates to form a hypersaline dense brine (concentrated by evaporation as dense brines) behind barrier reef would sink downward to the lagoon floor through underlying lime calcium carbonate sediment and thus displace less dense seawater in the pores of the sediment (or refluxes back to the sea) . Flushing of large volumes of this magnesium-rich brine through the underlying sediment would bring dolomitization, a process they referred to as seepage refluxion. During this process, Mg^{2+} replaces Ca^{2+} in $CaCO_3$ minerals, releasing Ca^{2+} in solution.



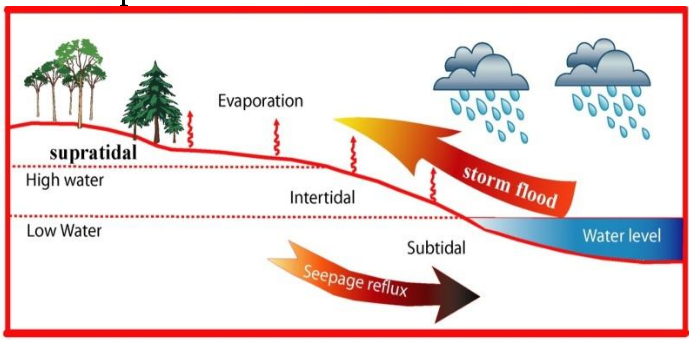
Brine reflux in an evaporitic setting. a sill to seaward restricts circulation of waters. Some of the seawater evaporates, causing water density to increase. the dense brines sink below the sediment, reflux through the basin or lagoon floor and dolomitize any carbonate sediments that they pass through

2- Sabkha reflux model

Sabkhas are coastal supratidal mudflats that are common in arid regions. Evaporite minerals precipitate displacively within sabkha sediments, which consist of carbonates and possibly siliciclastics, in a capillary zone above a saline water table. Upward flow of water from the saturated groundwater zone replaces the water lost by capillary evaporation, a process referred to as evaporative pumping. Water lost from sabkha sediments by evaporation is replaced by storm-driven seawater (storm surges or high tides push seawater landwared, over the preitidal sediments of a sabkha flat).

Magnesium for dolomitization is supplied by this seawater, onto the lower supratidal zone and along remnant tidal channels. The salinity of seawater is elevated within the supratidal flats. The resulting dense brines continues to percolate downward into the underlying lime sediments (reflux through the sabkha sediment) similar to downward flow in the reflux model.

Sabkha dolomite is commonly associated with supratidal sediments and features , such as algal stromatolites, nodular anhydrites and wind-driven interbedded deposits.



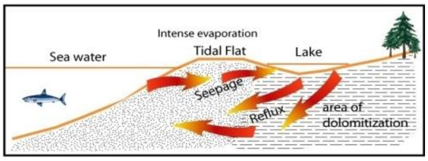
Sabkha reflux environment. this schematic of peritidal sediments in sabkha showing another variation on the reflux. Seawater is pushed onshore during storm floods, becomes concentrated through evaporation, then seeps into the underlying sediment to reflux to its source

3-Mixing-zone model

Dolomites were produced through mixing of seawater derived brines and freshwater. Such low-salinity groundwaters could be saturated with respect to dolomite at Mg/Ca ratios as low as 1:1. The mixing-zone model, or variations thereof, has been referred to also as the Dorag model. Mixing of meteoric waters with seawater causes under saturation with respect to calcite, whereas dolomite saturation increases, resulting in replacement of $CaCO_3$ by dolomite .When seawater or evaporated brine with high Mg/Ca ratios is diluted by mixing with freshwater , the mixture will retain the high Mg/Ca ratio and become special waters capable of forming ordered dolomite.

Mixing of seawater and freshwater takes place within sediments in coastal zones where saline formation waters are mixed through a transition zone with overlying, seaward flowing meteoric waters that form a lens above the saline formation waters. These mixed waters were initially considered capable of bringing extensive dolomitization.

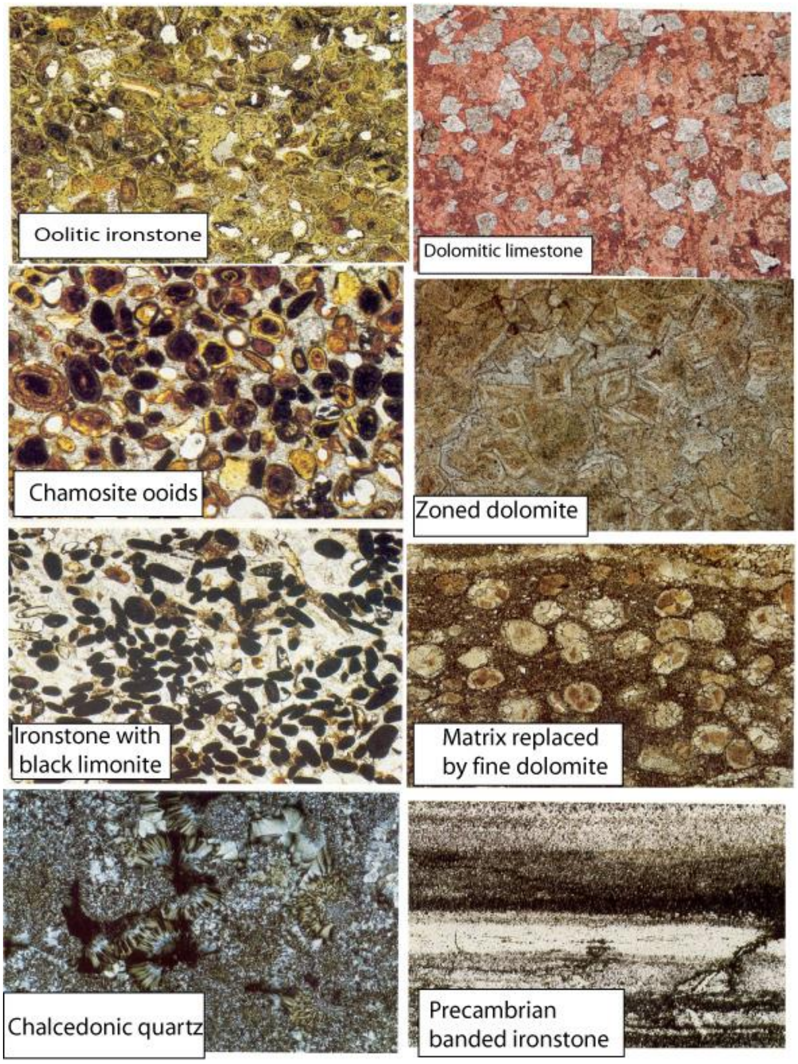
One characteristics of dorag dolomitization is the presence of cavities lined with dolomite crystals and are easily recognized with a hand lens,



The Dorag Model for dolomitization. Seawater seeps through sediments in intertidal-supratidal zone and into supratidal lakes if present. As a result of intense evaporation and precipitation of gypsum (which raises the Mg/Ca ratio), dense porewaters are produced which sink downwards and flow seawards, supposedly dolomitization the calcareous sediments through which they pass.

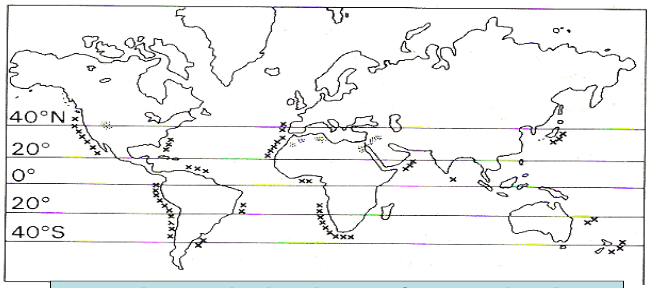
Dedolimitization

Although replacement of limestone by dolomite is common . thew reverse process, that is , replacement of dolomite by calcite is rare. The latter process is called dedolimitzation (calcitization of dolomite). The process of dedolimitization requires solutions with high Ca-Mg ratio, rapid flow of such waters, and tempetre under 50c



Phosphorites

Sedimentary phosphates occur in rocks of all ages from Precambrian to Holocene. Phosphorites are sedimentary deposits containing more than about 15–20 percent P_2O_5. Shales, sandstones, or limestones that contain less than 20 percent P_2O_5 but which are enriched in phosphorus over that found in average sediments are referred to as phosphatic, e.g. phosphatic shale. Phosphorus-rich sedimentary rocks are called by a variety of names –phosphate rock, rock phosphate, phosphates, phosphatites, and phosphorites.



A map showing the distribution of marine phosphorites in the world

Mineralogy

The most common sedimentary phosphate minerals are varieties of apatite. The apatite of igneous rocks is chiefly fluorapatite, $Ca_5(PO_4)3F$. Sedimentary phosphates are composed of phosphate minerals, all of which are varieties of apatite. The principal varieties are fluorapatite [$Ca_5(PO_4)3F$], chlorapatite [$Ca_5(PO_4)3CL$], and hydroxyapatite $Ca_5(PO_4)$ OH]. Most sedimentary phosphates are carbonate hydroxyl fluora-patites in which up to 10 percent carbonate ions can be substituted for phosphate ions to yield the general formulaCa10(PO4,CO3)6F2–3. These carbonate hydroxyl fluorapatites are commonly called francolite

The term cellophane is often loosely applied to sedimentary apatite of cryptocrystalline form, and it is isotropic

Phosphorites commonly also contain some detrital quartz and authigenic cher,. and organic-rich shales. Both calcite and dolomite may occur in phosphorites, and dolomite may be particularly abundant. Glauconite, illite, montmorillonite, and zeolites, may also be present.

Types of phosphorites

1- Nodular and bedded phosphorites

The marine phosphorite generally occurs in areas of slow sedimentation, on outer continental shelves and slopes, particularly on the tops and sides of local ridges and banks, on fault scarps and the flanks of submarine canyons. Phosphate nodules and crusts generally occur at depths from 60 to 300 m.

The nodules usually are several centimetres in diameter but may reach a metre or more. They range in shape from flat slabs to spherical nodules and ir-regular masses. The internal structure of the nodules varies from homogeneous to concentrically laminated and conglomeratic, and many contain pellets and coated grains (‘ooids’)

2- Guano and ocean-island phosphorites

The excrement of birds, and to a lesser extent, bats, may form thick phosphate deposits of economic significance. Leaching of the fresh guano leaves an insoluble residue composed mainly of calcium phosphate. Thick accumulations of bird guano are found on some small oceanic islands in the eastern Pacific coast of South America, and in the West Indies. Geologically, guano itself is not significant. However, downward percolation of solutions derived from guano may cause phosphatization of underlying carbonate sediments.

Petrographic classification of phosphates

Phosphorites have textures that resemble those in limestones. Thus, they may contain ooids, peloids, fossils (bioclasts), and clasts. Some phosphorites lack distinctive granular textures and are composed instead of fine, micrite-like, textureless cellophane,this called sphospho-lutite. In polarized light they appear as brownish isotropic material; however, when crystal size increases above a few micrometers their birefringence appears. Peloidal or pelletal phosphorites are particularly common. Peloidal phosphates may also include ooids that contain a nucleus and display well-developed concentric layering

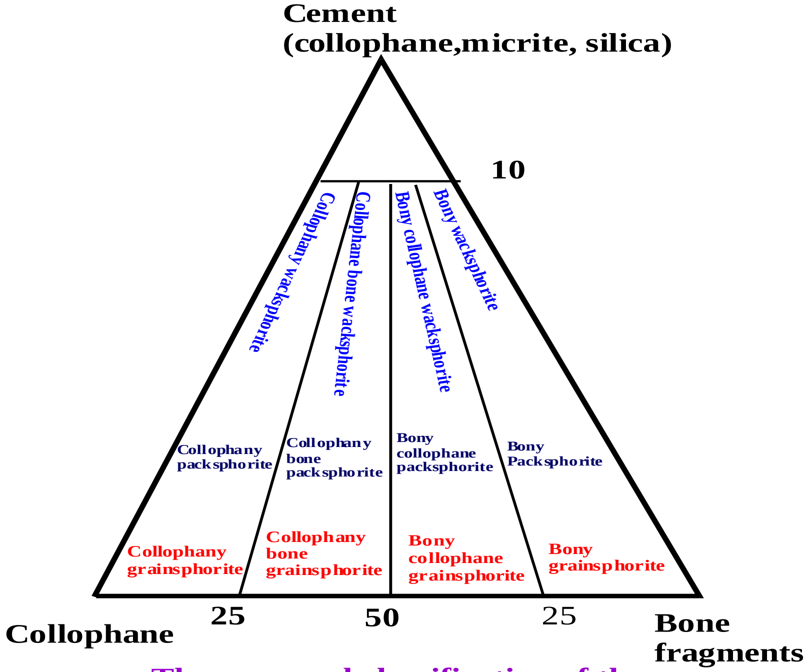
Bedded Phosphorites								Nodular Phosphorites	
Clastic Phosphorites							Metasomatic	Clastic	Metasom
Grains cement	Sand Size						Rara	Rara	Rare
	Grain Size More 2mm	. V.C. sand	C. Sand	M. Sand	F. Sand	V.F. sand			
Calcite	Calcitic phospho arenite								
Dolomite		Dolomitic V.C Phosphoarenite.							
Collophane			Collopho C.Phosphoarenite						
Gypsum				G.M.Phosphoarenite					
Silica									

Philobbos (1969)

Contain Mud			Lake Mud
Mud- Supported	Grain Supported		
Less than 10% Phosphate grains	More than 10% phosphate grains		
Mud-sphorite	Wack-sphorite	Pack-sphorite	Grain-sphorite
Not recorded	Collophany (collophane>75%, bone<25%	Collophany	Collophany
	Collophany bone (50-75%, Bone 25-50%	Collophany bone	Collophany bone
	Bony collophane, Collophna 25-50, bone,50-75%	Bony Collophane	Bony collophane
	Bony (bone grains>75%, Collophane <25%	Bony	Bony
Cement	Calcareous; siliceous; Collophane	Calcareous; Siliceous; Collophane	Calcareous; Siliceous; Collophane
	Primary micrite; Diagenetic sfter micrite	Primary micrite; Diagenetic sfter micrite	Sparite; Diagenetic after sparite

Classifications of phosphorites according to their depositional textures, mineralogical components and type of cement

Philobbos and Abdel Rahman (1990)



The proposed classification of the Egyptian Phosphorites

Origin of sedimentary phosphorites

1- As accumulations of bones, teeth , fish skeletons and Brachiopods

2- As accumulations of death animals and birds on islands and submarine. (called submarine Gauna)

3- By leaching of carbonate rock enriched by phosphate ions and concentration of phosphatic materials

4- As metasomatic replacement (by replacment of limestones by phosphates)

5-By direct precipitation from sea waters under physico-chemical conditions (by changes in Eh,Ph,temperatures and concentrations).

Origin of Egyptian Phosphorites

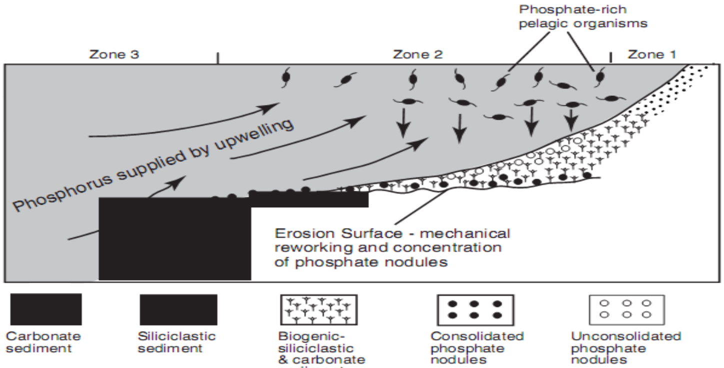
1- Ball (1913): The egyptian phosphorites are formed by replacement of calcium carbonates (limestones)by a solutuion enriched with phosphates

2- Hussein (1954): By mixing cold waters enriched by CO3 with heat shallow waters enriched with phosphate ions due to alterations of live organisms

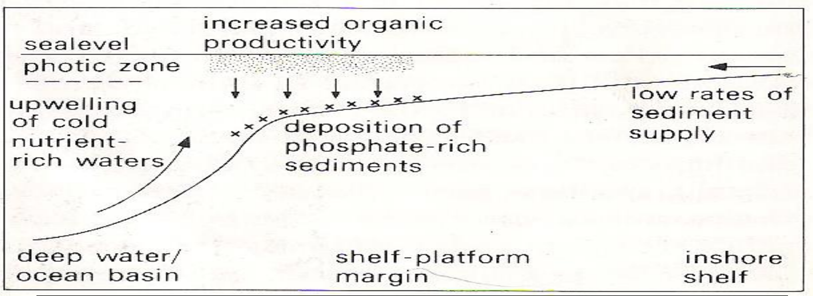
3- El Tarabelli (1964): Used the Upwelling current which lead to high organic productivites and phytoplankton growth in the surface waters which in turn results in organic-rich for formation of phosphorite deposits . Upwelling currents rise cold, nutrient-rich waters from depths towards the surface (photozone areas). Phosphate becomes concentrated in the pore waters of sediment during slow decay of the phosphate-bearing, soft-bodied organisms and other organic detritus, and deposited as nodular with accumulating sediment . Carbonate-apatite precipitates diagenetically from these phosphate-enriched pore waters. Subsequently, these diagenetic deposits are reworked mechanically owing to lowered sea levels, allowing concentration and deposition of phosphatic sediments by waves and currents in high energy environments. **Such** diagenetically formed phosphorite sediments are mechanically reworked under shallow-water conditions, is supported by the clastic textures and primary sedimentary structures found in many ancient phosphorites, as evidenced by Glenn and Arthur (1990) in their study of major phosphorite deposits of Egypt. They report that phosphorite grains that were precipitated initially in reducing shales have been mechanically reworked and concentrated. Reworking occurred as a result of sea-level fall and delta progradation. These processes are summarized diagrammatically below.

Opinion of Philobbos (1969)

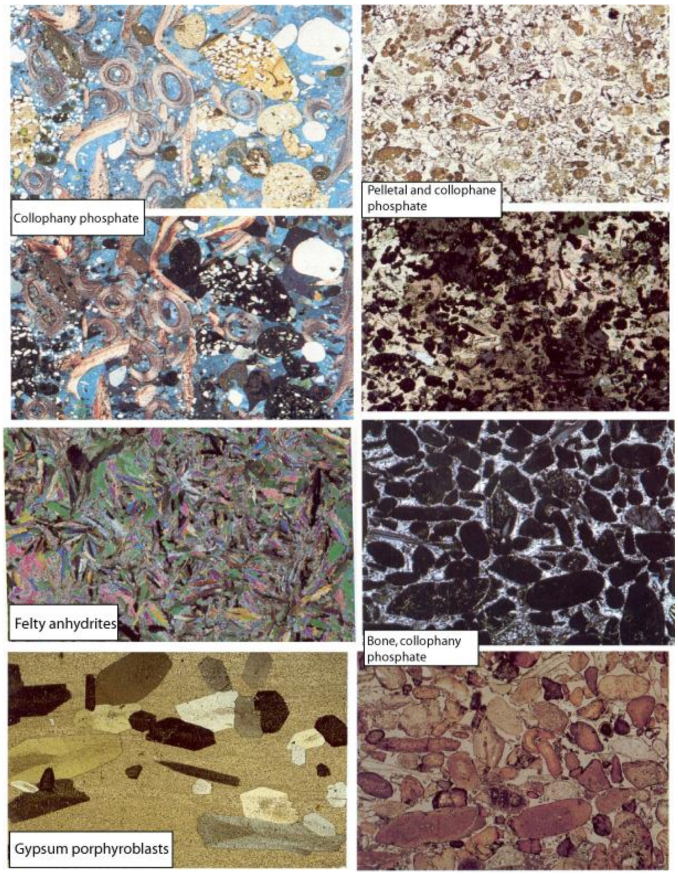
After death the organisms and alterations the phophate ione realized. Due to physico-chemical conditions a Collophane precpitates. By marine currents the collophane fragmented and changed into concentric bodied in sand size (oolites) or less than as pellets. Later on these materials are cemented by different materials as calcitic, ferrigenous, etc.



Schematic diagram illustrating formation of phosphorites in areas of upwellingon ocean shelves. A- supply of phosphors to the shelf by upwelling waters; B- consumption of phosphorus by organisms; C- deposition of phosphorus on the bottom and buried by accumulating sediment. D- formation of phosphate concretions in the biogenic sediment by diaganesis. E- mechanical reworking and concentration of phosphorus when sea level is low. Zone I : zone of shallow water clastic deposits. Zone II: is the zone of high phosphate accumulations, Zone III: reworking of Phosphate –rich sediments during low sea level, Zone IV: deep water zone. The arrows refer to paths of movement of phosphorus in the ocean and interstial waters.

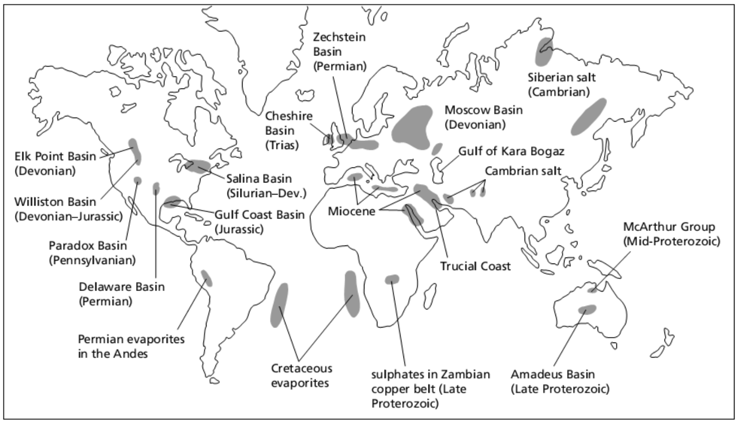


Model for formation of marine phosphorites



Evaporites

The term evaporites to include all sedimentary rocks formed by evaporation of saline waters. Evaporites make up less than about 2 percent of the sediments deposited on the world's platforms during Phanerozoic time. One of the thickest known deposits of evaporites is the Mediterranean Messinian evaporite sequence, which may exceed 2km in thickness. Evaporite deposits are commonly cyclic. Some consist of numerous thin evaporite beds, a few to tens of metres thick, typically of gypsum–anhydrite with little or no halite, alternating with limestone and marl.



Map showing location (and age) of the major evaporate deposits of the world

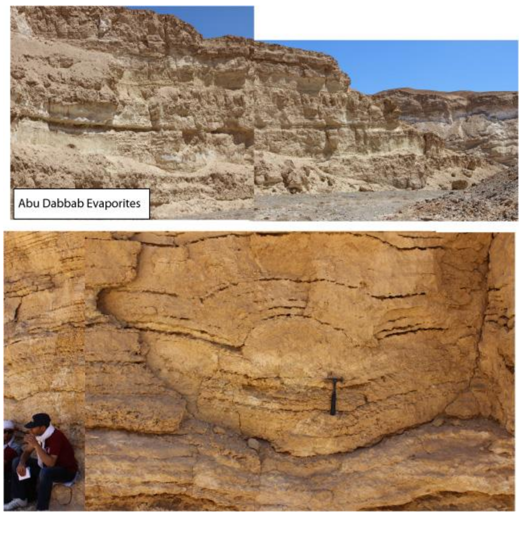
Importance of evaporites

Halite, gypsum, trona, and other salts are currently used for a variety of industrial and agricultural purposes. Evaporites are associated with carbonate rocks in many major oil fields of the world. Squeezing and remobilization of salt deposits creates petroleum traps in association with salt domes. Subsurface solution of evaporite cements in carbonate and siliciclastic rocks can create important amounts of secondary porosity in these rocks; and evaporite caprocks form seals over many petroleum traps that prevent the petroleum from escaping. Evaporites are useful in the studies of palaeoclimatology (arid conditions).

Mineral composition

Evaporite deposits are composed dominantly of varying proportions of halite (NaCl), anhydrite ($CaSO_4$) and gypsum ($CaSO_4 \cdot 2H_2O$.

Gypsum is more abundant than anhydrite in modern evaporite deposits; however, anhydrite is more abundant than gypsum in deposits buried to depths exceeding about 600m owing to dewatering of gypsum and its conversion to anhydrite**. Gypsum** is readily distinguished in the field because it is softer (hardness 2, easily scratched with a finger) and does not react with dilute HCl: it can be distinguished from halite by the fact that it does not taste salty. Gypsum has a low relief, cleavage is usually well developed and the birefringence colours are low-order greys. **Anhydrite** is a harder (hardness 3.5), denser mineral than gypsum: it is commonly white in hand specimen, and is not easily scratched by a finger. In thin section the high density means crystals have a relatively high relief; birefringence colours are moderate, higher-order colours than gypsum.



Classifications of evaporites

I- based on mineral composition

a- <u>Marine evaporite</u>

Marine evaporites are precipitated from seawater, the mineral composition of marine evaporites in various deposits tends to be relatively constant.

Mineral class	Mineral name	Chemical composition	Rock name
Chlorides	Halite	$NaCl$	Halite; rock salt
	Sylvite Carnallite	KCl $KMgCl_3 \cdot 6H_2O$	Potash salts
Sulfates	Langbeinite Polyhalite Kainite	$K_2Mg_2(SO_4)_3$ $K_2Ca_2Mg(SO_4)_4 \cdot H_2O$ $4MgSO_4 \cdot 4KCl \cdot 11H_2O$	
	Anhydrite Gypsum Kieserite	$CaSO_4$ $CaSO_4 \cdot 2H_2O$ $MgSO_4 \cdot H_2O$	Anhydrite; anhydrock Gypsum; gyprock –
Carbonates	Calcite Magnesite Dolomite	$CaCO_3$ $MgCO_3$ $CaMg(CO_3)_2$	Limestone – Dolomite; dolostone

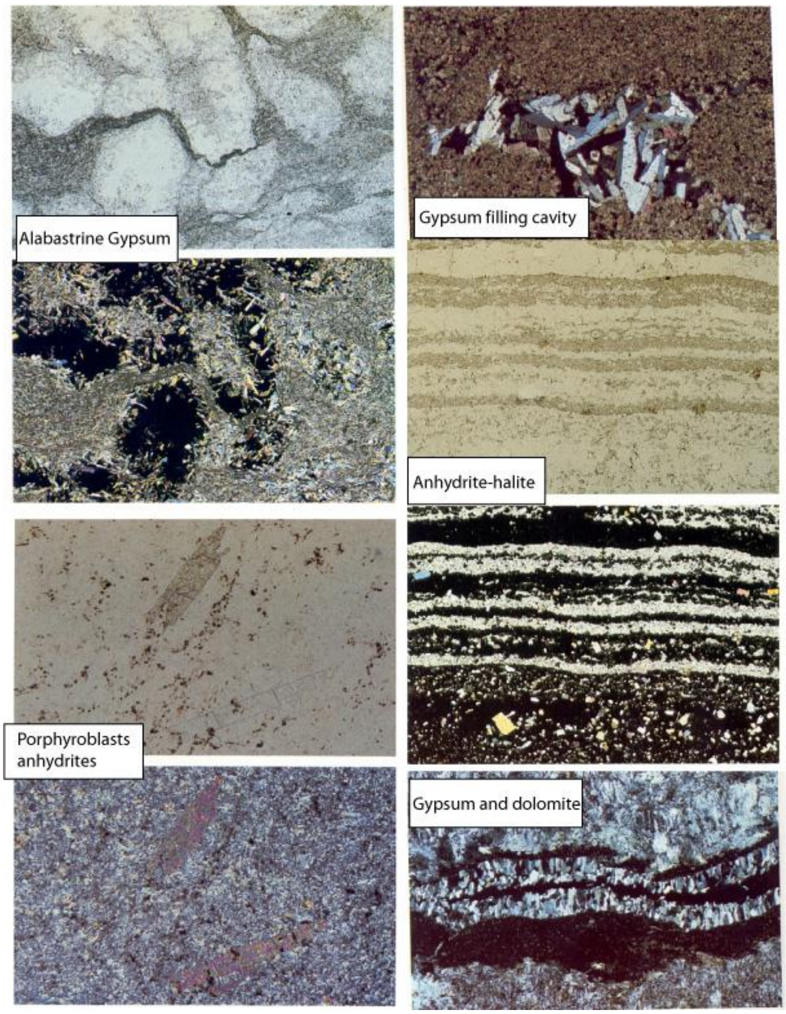
Classification of marine evaporites on the basis of mineral composition

2- <u>Nonmarine evaporites</u>

Non marine evaporites form river water or groundwater. The chemistries of these waters can be highly variable, depending upon the lithology of the rocks with which they interact. For example, rivers that flow across limestones are commonly enriched in Ca 2+ and HCO3, whereas those that flow across igneous and metamorphic rocks tend to be enriched in silica, Ca 2+ , and Na + . Thus, many nonmarine deposits contain evaporite minerals that are not common in marine evaporites.

Lacustrine evaporites

Evaporites deposited subaqueously in lakes will have similar sedimentary features to marine sub-aqueous evaporites: fine lamination in lake centres, slumped and turbidite beds near slopes, and reworked evaporites and evaporitic stromatolites in shallow water. Salt lakes commonly are surrounded by inland sabkhas and saline mudflats. Saline pans are common in deserts. Ancient lacustrine evaporites are well developed in the Eocene Green River Formation of Wyoming and Utah.



II-based on fabric and bedding.

. This classification divides anhydrites into three fundamental structural groups: nodular anhydrites, laminated anhydrites, and massive anhydrites.

1- **Nodular anhydrite** consists of irregularly shaped lumps of anhydrite that are partly or completely separated from each other by a salt or carbonate matrix.

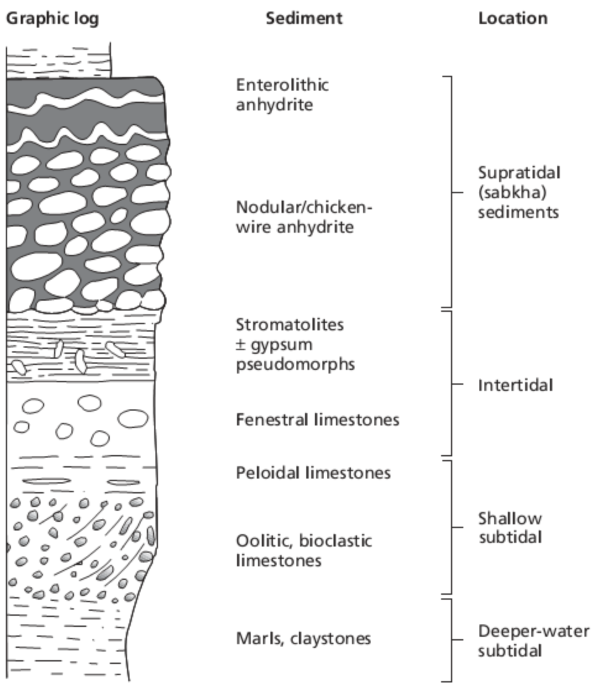
A-Mosaic anhydrite is a type of nodular anhydrite in which the anhydrite masses or lumps are approximately equidimensional and are separated by very thin stringers of dark carbonate mud or clay. The formation of nodular anhydrite begins by displacive growth of gypsum in carbonate or clayey sediments. Subsequently, gypsum crystals alter to anhydrite pseudomorphs, which continue to enlarge by addition of Ca and SO4. This displacive growth ultimately results in segregation of the anhydrite into nodular masses

B-Chickenwire structure is a term used for a particular type of nodular anhydrite that consists of slightly elongated, irregular polygonal masses of anhydrite separated by thin dark stringers of other minerals such as carbonate or clay minerals. This structure apparently forms when growing nodules ultimately coalesce and interfere. Most of the enclosing sediment is pushed aside and remains as thin stringers between the nodules

Sabkha Evaporite sequence

Sabkhas and supratidal flats as important environments of evaporite (particu-larly sulphate) precipitation. The typical facies are nodula (chicken-wire) and enterolithic anhydrite, , these textures can develop in subaqueous anhydrites by replacement.

The identification of sabkha evaporites are the shallow-water and intertidal sedimentary structures contained within associated carbonates .Occasionally the sabkha gradually progrades seawards over the intertidal sediments. A sabkha cycle of supratidal evaporites overlying intertidal and subtidal carbonates is produced, which may be repeated many times in an evaporite formation.

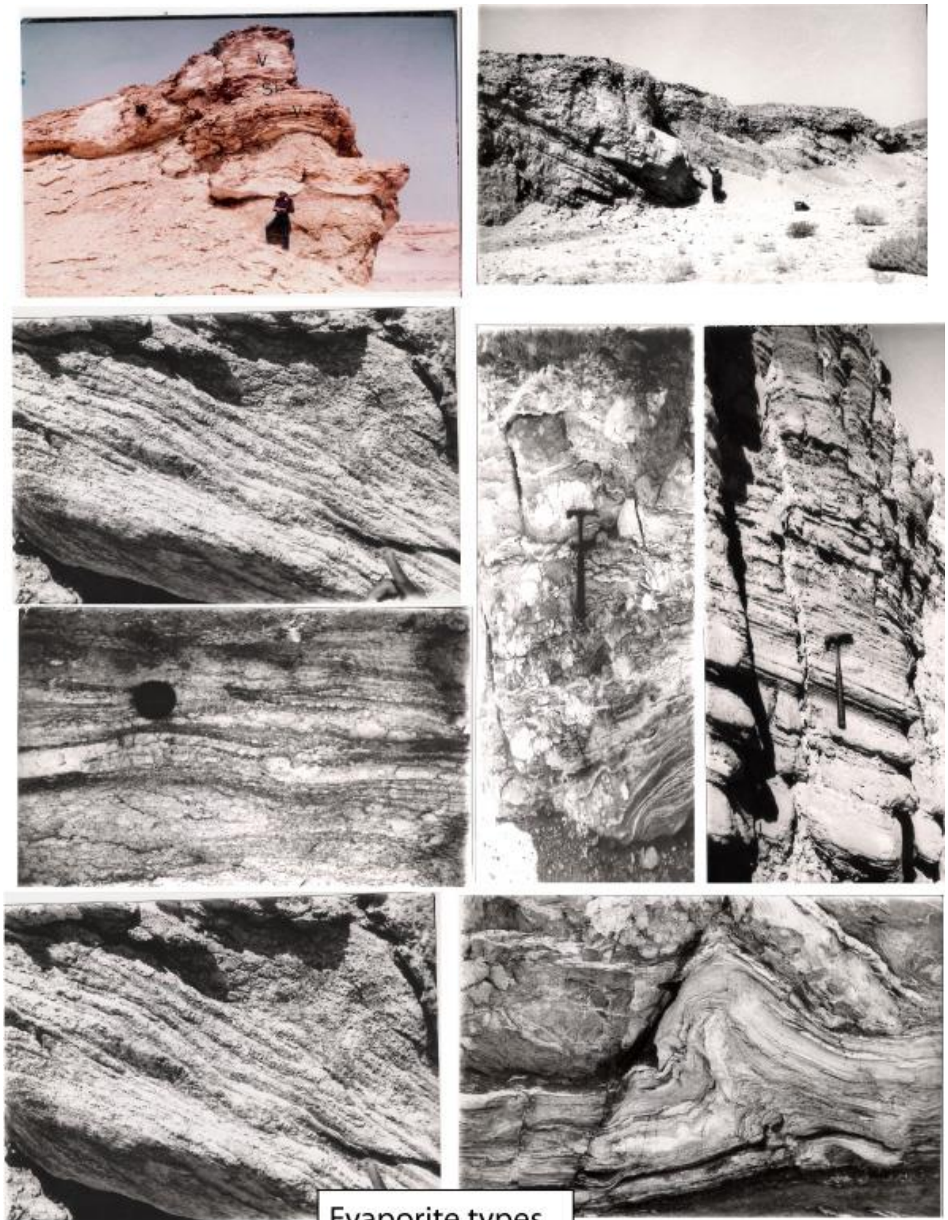


A sabkha cycle. Such cycles typically range from several to several tens of metres in thickness.

2-Laminated anhydrites, sometimes called laminites, consist of thin, nearly white anhydrite or gypsum laminations that alternate with dark-gray or black laminae

rich in dolomite or organic matter. The laminae are commonly only a few millimeters thick and rarely attain a thickness of one centimeter. Many thin laminae are remarkably uniform with sharp planar contacts that can be traced laterally for long distances. Laterally persistent laminae are believed to form by precipitation of evaporites in quiet water below wave base. They form either in a shallow-water area protected from strong bottom currents and wave agitation or in a deeper-water environment. Alternating light and dark pairs of bands in laminated evaporites may represent annual varves resulting from seasonal changes in water chemistry and temperature.
Some laminated anhydrite may form by coalescence of growing anhydrite nodules, which expand laterally until they merge into a continuous layer. Layers formed by this mechanism are thicker, less distinct, and less continuous than laminae formed by precipitation. A special type of contorted layering that has resulted from coalescing nodules has been observed in some modern sabkha deposits where continued growth of nodules, forming ropy bedding or enterolithic structures.

Primary textures exhibited by anhydrite in both modern and ancient nodular deposits include fine equant mosaics (aphanitic), and felted and parallel– subparallel arrangements of laths. Recrystallization of equant and lath anhydrite may take place to produce coarse granular mosaics, large fibrous crystals and fibro-radiating aggregates.



Evaporite types

3-Massive anhydrite is anhydrite that lacks internal structures, forms by evaporation at brine salinities, under relatively dry climatic conditions.

4- Bottom-growth gypsum

Gypsum crystals can be precipitated on the floor of lagoons, lakes and shallow shelves around evaporate basins in a variety of crystal forms, some spectacular and very large (up to 15cm high). **This selenitic gypsum** commonly grows vertically, almost as grass, as single (prismatic), twinned and split crystals. Curved crystal faces are common, too, induced by the presence of organic matter. Beds of these gypsum crys-tals commonly have a distinctive palmate (palm-frond shape) texture.

5-Secondary and fibrous gypsum

Uplift of anhydrite deposits, perhaps a long time after their formation and burial, results in the generation of secondary gypsum, as the anhydrite comes into con-tact with fresh near-surface ground water. Secondary gypsum consists of two varieties**, porphyrotopes and alabastrine gypsum**. Gypsum porphyrotopes are large crystals, typically several millimetres across or larger, which occur scattered through the anhydrite. **Alabastrine gypsum** consists of small to large, commonly poorly defined interlocking crystals, many with irregular extinction. In spite of gypsifi-cation, the original nodular or laminated texture of anhydrite normally is preserved.

Veins of fibrous gypsum are commonly parallel or subparallel to the bedding, having a displacive (intrusive) relationship. Usually they are a few millimetres to centimetres in thickness and consist of vertically arranged fibres. It is thought that the fibrous gypsum grew under pressure in brine-filled veins induced by hydraulic fracture or unloading and exhumation. The gypsum probably has been derived from dissolution of near-surface gypsum or rehydra-tion of more deeply buried anhydrite.

Depositional models for evaporites

Many thick sequences of ancient marine evaporites appear to have formed in laterally extensive, subaerial to shallow- to deep-water basins

1-Subaerial evaporites

Many modern subaerial evaporites accumulate in sabkhas or salt flats.

The sabkhas as “extensive, barren, salt-encrusted, and periodically flooded, coastal and inland mudflats. Sabkhas can occur in both continental settings (nonmarine sediments and continental groundwaters) and marginal-marine settings.

In continental environments, they occur in inland areas in fluvial-lacustrine (playa)-dominated settings and in interdune settings. In the marginal-marine environment, they occur as coastal sabkhas in the intertidal and supratidal zones. Sabkhas are composed of a combination of evaporite minerals and carbonate and/or terrigenous clastic sediments. Evaporite minerals may form both at the surface of sabkhas and displacively within sabkha sediments, creating distorted fabrics.

The subaerial precipitation of evaporites has taken place on the floor of deep-marine basins following evaporative drawdown and extreme desiccation. This occurred during the late Miocene when the Mediter-ranean dried up after being cut off from the Atlantic. Evaporites may precipitate in soils of desert areas to give crusts and indurated horizons (e,g. Gypcrete or gypsite).

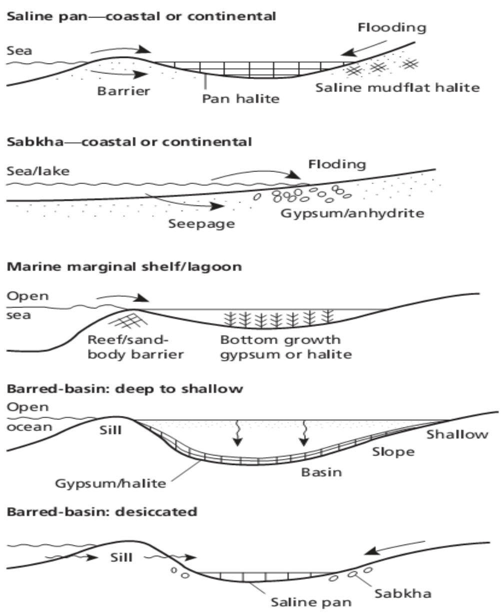
2-Shallow subaqueous evaporites

Shallow subaqueous evaporites accumulate in the coastal-marine environment **(Salinas)**.Salinas typically occur in carbonate environments, commonly in depressions within coastal carbonate dunes or in association with carbonate reefs. Gypsum appears to be the most common evaporite mineral in most salinas; The stratigraphic record suggests that many ancient shallow subaqueous evaporites were deposited on broad platforms. Subaqueous evaporites occur also in lacustrine settings in continental basins.

3-Deep-water evaporites

They are characterized by thin bedding and lamination and lateral continuity
of beds and laminae. Strata are composed predominantly of laminar evaporitic carbonate, sulfate, and halite. They occur in sections tens to hundreds of meters thick that can be correlated over tens to hundreds of kilometers

Although most deep-water evaporites probably formed mainly by in situ precipitation, some appear to be turbidites. These evaporite turbidites are formed by rapid precipitation in the shallow areas of evaporite basins, resulting in unstable marginal accumulations that were subsequently redeposited downslope by turbidity currents.

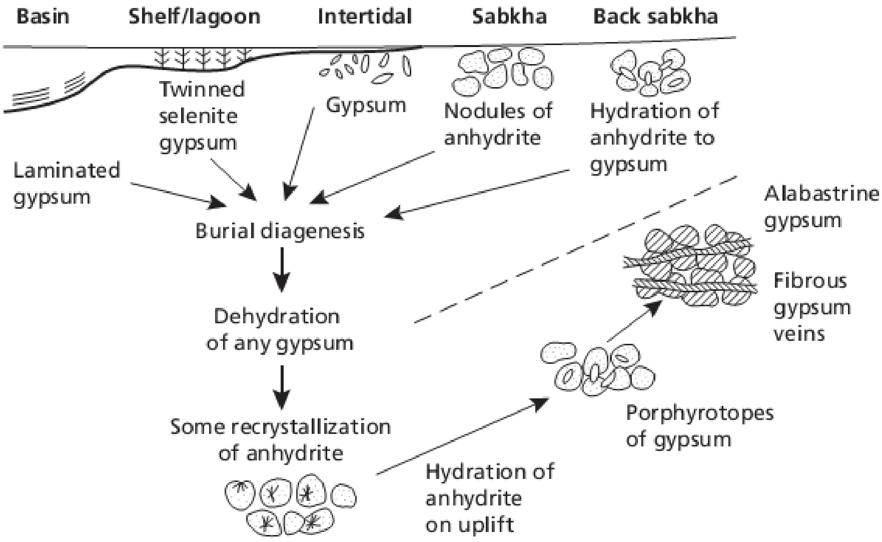


Principal depositional environments of evaporates (Tucker, 1990)

The ‘gypsum–anhydrite cycle’

The main site of marine sulphate precipitation today, where the early part of the gypsum–anhydrite cycle can be observed, is in the high intertidal and supratidal Zones. Gypsum is being precipitated displacively within the sediments as discoidal, rosette, selenite and twinned crystals from less than 1 mm to more than 25 cm in size. Sediment pore waters are largely derived from surface flooding of sea water, a process referred to as flood recharge. Dolomitization of carbonate particles is commonly associated with gypsum precipitation, as a result of the increased Mg/Ca ratio (Section 4.8.2), and this releases calcium ions for further gypsum precipitation. Gypsum is the most common precipitate within the sediments of inland sabkhas, where it forms the familiar desert roses.

Continued displacive precipita-tion of anhydrite results in closely packed nodules with host sediment restricted to thin stringers. The nodular texture produced is referred to as chicken-wire anhy-driteand this is the typical texture of many ancient sul-phate deposits. Anhydrite also is precipitated as thin beds or layers of coalesced nodules in the more landward parts of the sabkhas. These beds are com-monly irregularly contorted and buckled, forming the so-called enterolithic texture, also common in ancient sulphate formations. In the most landward part of the sabkha, some rehydration of anhydrite to gypsum may occur from contact with fresh continental ground waters.



The ‘gypsum–anhydrite cycle’ showing mineral and textural changes, from the surface, into the subsurface and on uplift (Tucker, 1990)

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Siliceous sedimentary rocks (cherts)

Siliceous sedimentary rocks are fine-grained, dense, commonly very hard rocks composed dominantly of the **SiO2** minerals. **Chert** is the general group name used for siliceous sedimentary rocks. **Cherts** make up less than l percent of all sedimentary rocks, but they are represented in stratigraphic sequences ranging in age from Precambrian to Quaternary.

Chert is a dense, very hard rock, which splinters with a conchoidal fracture, fine grained siliceous sediment, composed mainly of chalcedony and cryptocrystalline silica.

Physico-chemical conditions govering the precipitation and dissolution of silica

The silica depends on Ph value, and it dissolves in alkaline media and deposits in acidic media.

Source of Silica

1- (Source of biogenic silica) Siliceous organisms (radiolaria, diatoms, sponges **and silicoflagellates**) can segregate the silica to build their skeletal framework),

2- quartz grains, silicates(clay minerals) and volcanic materials can be dissolved under certain physico-chemical conditions.

Mineralogy and texture

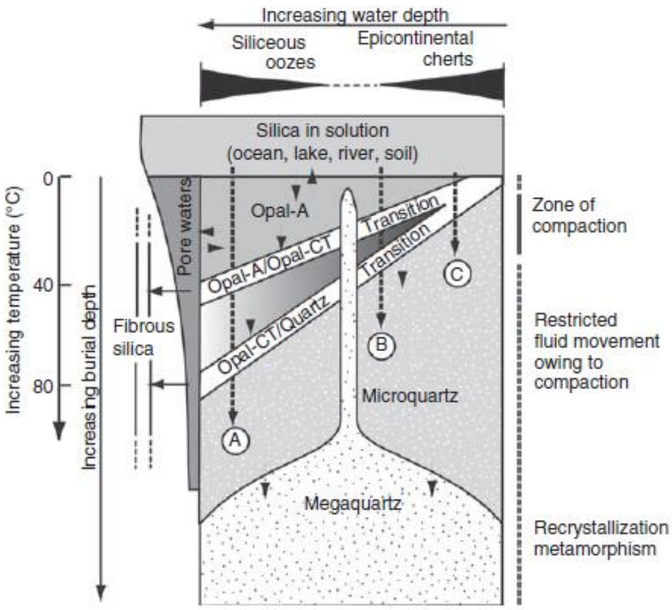
Quartz is the primary mineral of siliceous deposits; however, other SiO_2 minerals in these deposits can include chalcedonic quartz, amorphous silica **(opal-A),** and low-temperature disordered cristobalite **(opal-CT**). We use the group name **chert** to cover all rocks formed of these minerals.

Opal-A is primarily of biogenic origin and forms the tests of siliceous plankton and the spines of some sponges. Skeletal opal-A is metastable and converts in time to opal-CT and finally quartz

Opal-A organic remains are present in some cherts, particularly those of Cenozoic age, suggesting a biogenic origin for these cherts.

All gradations may be present in chert deposits – from pure opal to pure quartz chert, depending upon the age of the deposits and the conditions of burial.

Texturally, the quartz that forms cherts can be divided into three main types: (l) microquartz, consisting of nearly equidimensional grains of quartz less than about 20 microns in size, (2) megaquartz, composed of equant to elongated grains greater than 20 microns, and (3) chalcedonic quartz, forming sheaflike bundles of radiating, thin crystals about 0.1mm long. The chalcedony can occur as pseudomorphs of gypsum and anhydrite or in nodules that closely resemble the chickenwire structure of nodular evaporates.



Schematic diagram showing major silica phases and their possible diagenetic transformations

Principal kinds of cherts

Flint: is used frequenrtly, for chert nodules occurring in in Cretaceous chalks.and Eocene limestones

Jasper is a variety of chert colored red by impurities of disseminated hematite. Jasper that is interbedded with hematite in Precambrian iron formations is called jaspilite.

Porcellanite is a term used for fine-grained siliceous rocks with a texture and fracture resembling that of porcelain. The term is often used by chert workers for cherts having this character that are composed mainly of opal-CT

Chert petrology

Nodular and bedded cherts consist of three main types: microquartz, megaquartz and chalcedonic quartz.
Microquartz consists of equant quartz crystals only a few microns
Megaquartz; are larger reaching 500 um
Chalcedonic quartz is a fibrous , larger in size, radiating arrangement, forming wedge-shaped and spherulitic growth structures.

Classifications of cherts

I-Based on Gross morphology:

1-Bedded Chert

Bedded chert consists of layers, ranging to several centimeters in thickness, that are commonly interbedded with millimeter-thick laminae of siliceous shale.

Bedded cherts can be subdivided on the basis of type and abundance of siliceous organic constituents into four principal types:

i-Diatomaceous deposits include both diatomites and diatomaceous cherts. Diatomites are light-colored, soft, friable siliceous rocks

ii-Radiolarian chert is well-bedded, microcrystalline radiolarite that has a well-developed siliceous cement or groundmass.

iii-Siliceous spicule deposits include spicularite, a siliceous rock composed principally of the siliceous spicules of invertebrate organisms, particularly sponges. Spicularite is loosely cemented .

iv-Bedded cherts containing few or no siliceous skeletal remains.

Cherts in this group include most cherts associated with the Precambrian iron formations, formed by silicification (replacement) of volcanic rocks,

2- Nodular cherts

Chert can also occur in the form of small nodules, lenses, or thin, discontinuous beds. Nodular cherts are especially common in carbonate rocks but occur also in evaporites and siliciclastic rocks. Nodular cherts range in size from a few centimeters to several tens of centimeters. They commonly lack internal structures, but some nodular cherts contain silicified fossils or relict structures such as bedding, and the preservation of burrows and other sedimentary structures suggest that most are formed by diagenetic replacement

Nodular cherts are often concentrated along bedding planes, they coalesce to form near-continuous layers.

Origin of nodular chert

Nodular cherts originate mainly by diagenetic replacement of carbonate rocks. Diagenetic origin is clearly demonstrated by the presence of partly or wholly silicified remains of calcareous fossils or ooids, burrow fillings, algal structures, etc.

Nodular replacement cherts occur in shallow, platform carbonate rocks suggests that the mixing zone where meteoric ground waters mix with seawater in a coastal area**.** Nodular cherts have also from silica replacement of anhydrite

II-Based on Age relations (relative to the associating sediments):-

1-Syngenetic origin

The chert is deposited directly, at the same time of deposition of the associating sediments .The absence of replacement indications in some chert support this theory.

This theory fails to explain the cherts showing replacement indications.such as:

1- The presence of irregualr patches of limestones within some charts.
2- The very irregular shape of some chert nodules
3- The presence of some silicified fossils.
4- The preservation of some textures and structures of the host rocks within the chert
5-The occurrence of silicified nummulites and ooliths in some cherts

2-Epigenetic origin

It is formed after deposition and consilidation of the host rocks namely by replacement of the latter. This theory cannot be applied for all cherts. Chert along joints and faults is best explained by this mode of formation.

3-Penecontemporaneous origin

The cherts are formed by replacement after deposition of the surrounding sediments and before its consilidation This theory explains the bedded, lenticular and nodular cherts .

Non--marine(Nonbiogenic) siliceous sediments and cherts

Inorganic precipitation of silica **(no siliceous organic** remains) was formed from dissolved detrital quartz grains and clay minerals. Also another inorganic source of silica, is volcanic rocks and rock fragments. In the lake waters the great fluctuations in Ph values (greater than 10), so that the lake waters become supersaturated with respect to amorphous silica. Evaporation of lake water and a decrease in Ph causes silica to be precipitated.

Direct, inorganic precipitation of amorphous silica has been reported in some ephemeral Australian lakes, Pleistocene cherts from alkaline Lake Magadi, Kenya,

.Also Late Oligocene Sodmin Formation west of Quseir , Egypt include imorganic nodular and lenticular cherts (Mahran, 1999). The scarcity of radiolarians and sponge spicules in these preclude the possibility that these cherts were formed by organisms.

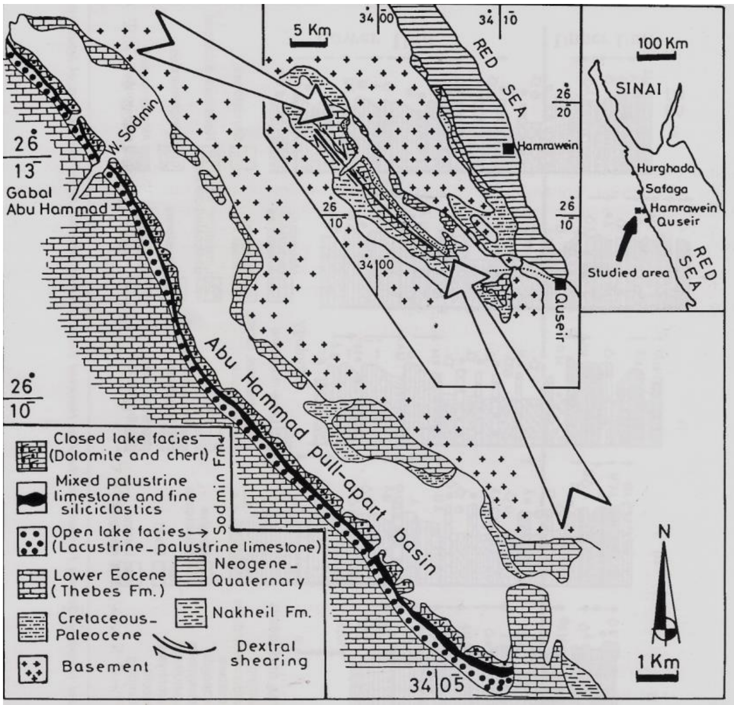
. Presumably, silica was precipitated by chemical (inorganic) processes. They could have been derived from siliceous rocks that were subsequently almost completely dissolved and recrystallized. Also some of the inorganic nodular cherts apparently formed by replacement of precursor sediments, were formed by silicification of volcaniclastic and pyroclastic deposits, terrigeneous sandstones and shales, biogenic sediments (algal mats), and evaporites.

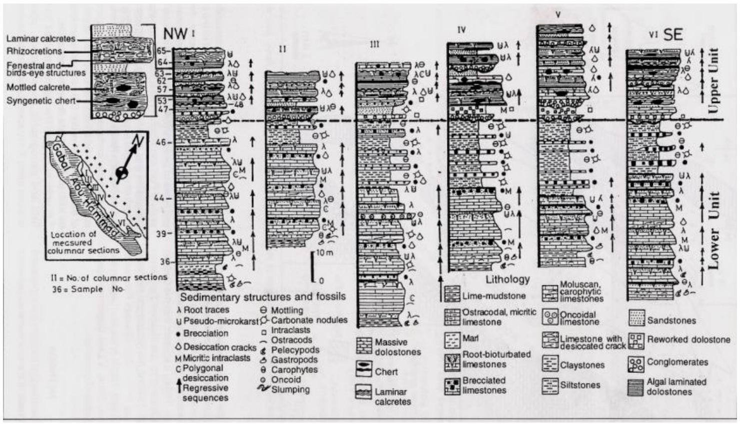
Model in Egypt

Example, of inorganic silica, Abu hammad lake basin, Red Sea (Mahran, 1999)

<u>Source of silica</u>

Silica sourced from dissolution of Eocene cherty limestones surrounding the lakes, also detrital quartz and chert clasts as well as the alluvial siliciclastic (siltstones and claystones) of the lower unit. Silica probably entered the lake in groundwater (having high Ph) in solution or as colloids, rather than as detrital grains.





Precipitation of silica

Fluctuations of Ph are considered an essential mechanism of the chemistry of Abu Hammad lake water for precipitation of cherts. Two scenarios may explain fluctuations of the Abu Hammad lake water:

1- The first scenario, algal materials can raise Ph in lake water. On the other hand increased CO_2 (due to decay of organic matter) will lower the Ph value. This hypothesis is compatible with the presence of nodular cherts within algal laminited dolostones (unit B).

2- In the second scenario fluctuations took place due to mixing of siliceous bicarbonate groundwater and rain water. The mixing resulted in dilution and the lowering of Ph and increasing precipitation of silica.

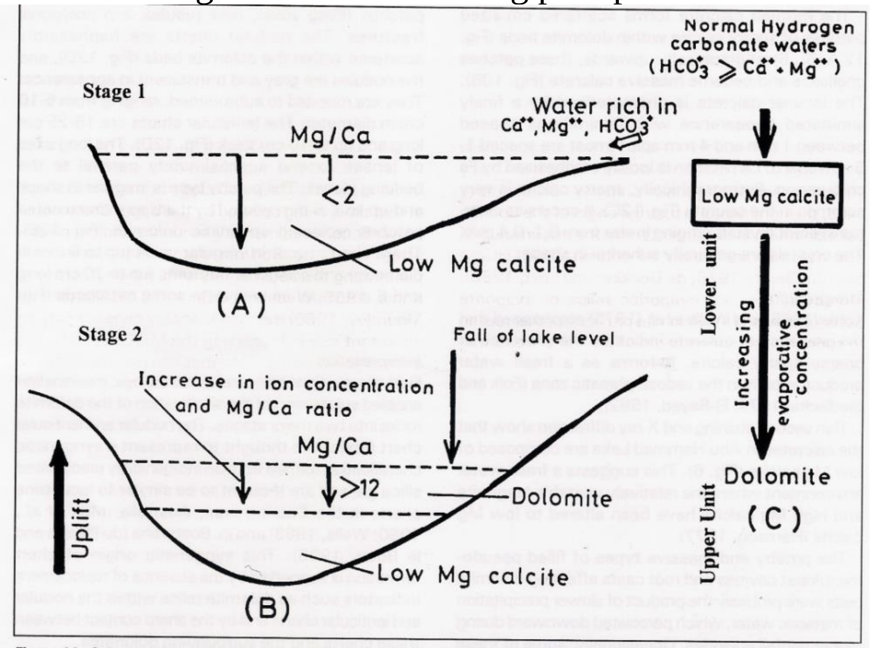


Figure 11. Schematic models of a carbonate-dominated, closed lacustrine depositional system illustrating different stages of lake level and of ion concentrations. (A) Stage 1. Precipitation of low Mg calcite. (B) Stage 2. Precipitation of dolomite. (C) Simplified scheme of carbonate precipitation due to increasing evaporative concentration.

Importance of cherts

1- The origin and the common occurrence of bedded cherts in ancient subduction complexes is of great interest to geologists concerned with aspects of Earth history such as paleogeography, paleoceanographic circulation patterns, and plate tectonics.

2- Siliceous rocks also have some economic significance. Silicon is used in the semiconductor and computer industries, and it is used also for making glass and related products such as **fire bricks**.

3- Siliceous deposits occur in association with important economic deposits of other minerals, including

Precambrian iron ores, uranium deposits, manganese deposits, and phosphorite deposits.

4-Petroleum deposits also occur in association with siliceous, which may be source rocks and possibly even reservoir rocks for petroleum.

Iron-bearing sediments

(Ironstones)

The term iron-bearing sediments is denoted for rocks containing high percentages of iron. Or sedimentary rocks with more than 15% iron,

On the basis of the iron minerals present, they are classified.

1- The sulphides, 2- The carbonates , 3- The oxides 4- The silicates.

1-The sulphides

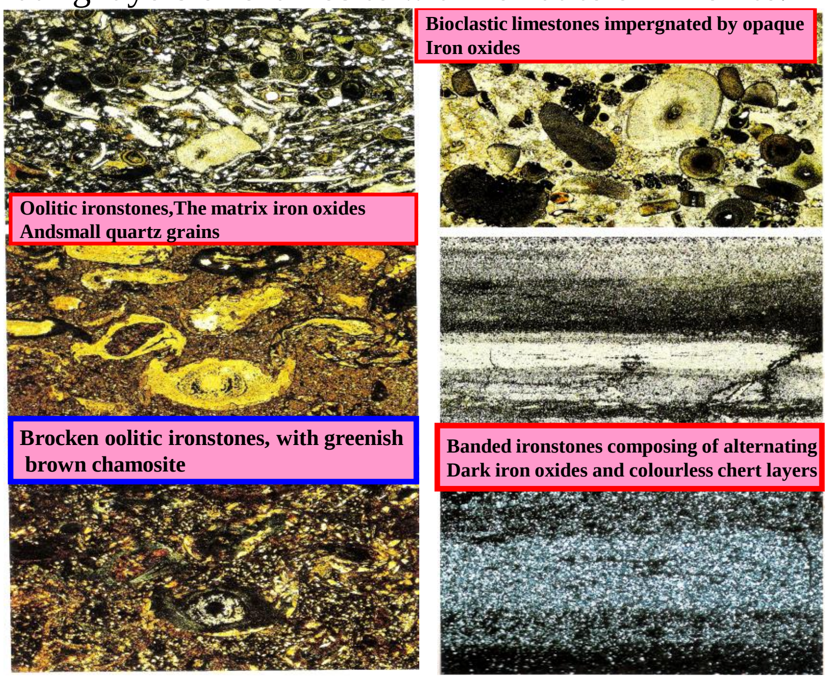
Marcasite and pyrite are very common minor constituents in some sedimentary rocks such as black shales and cherts. They are found in geodes or may be present as spherulites, nodules or as fine dissaminations. In some cases pyrite is found as beds containing primarly minute spherulites which may be cemented by siliceous groundmass.

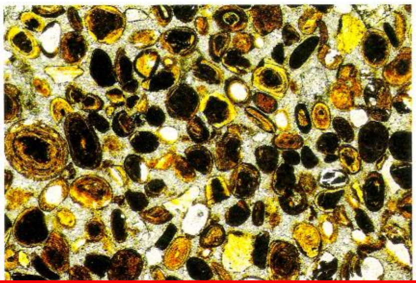
2--Carbonates (bedded siderites)

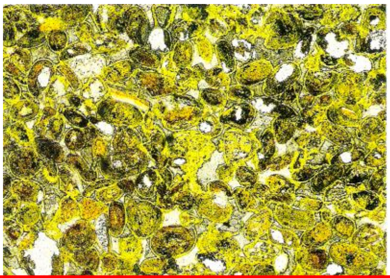
Siderite is found as beds usually intercalated with chert or mixed with clay forming clay ironstones. The siderite can be deposited directly from sea water or may be formed by penecontemporaneous replacement of calcite beds

3-Bedded iron oxides

A) fossil ores and b) oolitic ores. Fossil ores consist of two types of bedded iron oxides are distinguished by brocken shell fragments, which are formed by action of waves and currents and subsequently replaced and cemented together by hematite. The oolitic ore consists of an aggregates of hematite or limonite oolites. Some oolites contain alternating layers of chamosite with hematite or limonite.

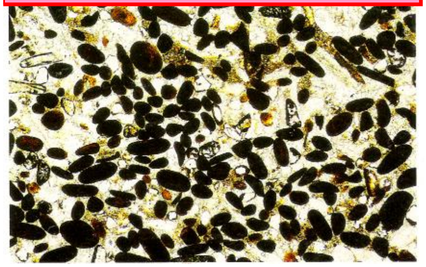


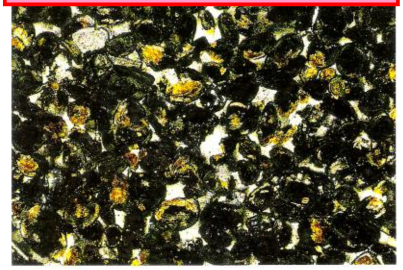




Oolitic ironstones (chamosite, iron silicate) The brown areas within grains are siderite (iron carbonates), cement is calcite

Oolitic ironstones (chamosite, iron silicate) The brown areas within grains are siderite (iron carbonates), cement is calcite





2-Bedded iron silicates

The most important iron silicates are chamosite, greenalite and glauconite. Chamosite is a ferrous silicate, and it forms under reducing conditions. It occurs as ooids, without nucleous in most cases. The ooids are commonly flattened and distorted

Origin of Iron

1- Iron is attributed to be volcanic activity
2- It formed due to weathering processes
3- Due to karstification processes
4- pH plays a major role in the deposition of iron minerals.(ex. Pyrite requires pH 3, and siderite requires for its precipitation pH 6.5.
5- Certain iron minerals, such as pyrite can be deposited by action of sulphate reducing bacteria

Miocene Iron mineralization in the Red Sea area

Iron mineralization is recorded north of Abu Ghusun in the form of fine crystalline masses filling cavities, fractures, joints and intergranular spaces of some siliciclastic and carbonate sediments. Mineralogically, the iron minerals include geothite, hematite. Texturally, the iron minerals exhibit colloform,intergranular, radiated, dendritic and zoned textures

Localization of the iron

The localization of iron is controlled by stratigraphic,lithologic and structutral factors. Generally the iron is concentrated in the uppermost parts of the carbonates and clastic rocks. The mineralized beds are usually underlain by fine terrigenous sediments (shales).

Origin of the Miocene iron at Abu Ghusun area, Red Sea

It isbelived that the iron mineralization have resulted from surface solutions derived from the weathering of nearby basement rocks.

Volcaniclastic sediments

Volcaniclastic sediments are those composed chiefly of grains of volcanic origin, derived from contemporaneous volcanicity. The type of magma— acid, intermediate or basic, determined largely by plate-tectonic setting—is an important factor in the type of volcaniclastic deposit generated. Thus, volcaniclastic refers to any deposit that is mainly composed of volcanic detritus.

During eruption volcanoes produce a range of materials that include molten lava flowing from fissures in the volcano and particulate material that is ejected from the vent to form volcaniclastic deposits. The location of volcanoes is related to the plate tectonic setting, mainly in the vicinity of plate margins and other areas of high heat flow in the crust. The pre-sence of beds formed by volcanic processes can be an important indicator of the tectonic setting in which the sedimentary succession formed. Lavas are found close to the site of the eruption, but ash may be spread tens, hundreds or even thousands of kilometres away. Volcaniclastic material may therefore occur in any depositional environment and hence may be found associated with a wide variety of other sedimentary rocks . Volcanic rocks are also of consid-erable value in stratigraphy as they may often be dated radiometrically, providing an absolute time constraint on the sedimentary succession.

Classification of volcaniclastic sediments

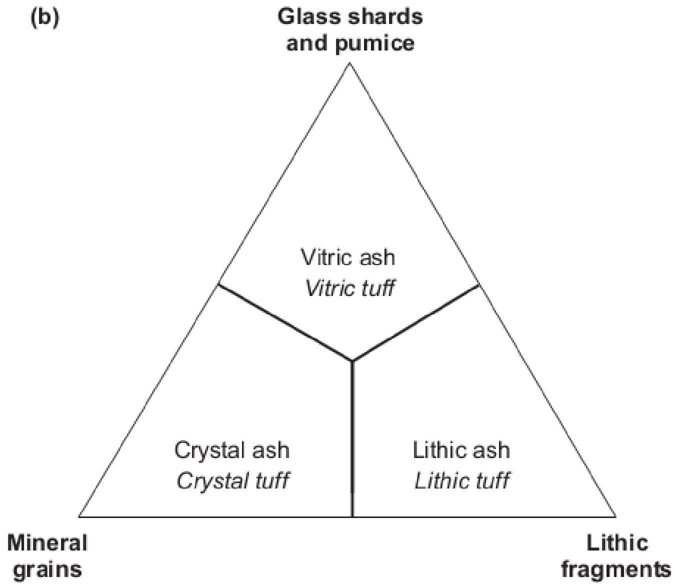
1- based on grain size (textural classification)

Coarse material (over 64mm) is divided into volcanic blocks, which were solid when erupted, and volcanic bombs, which were partly molten and have cooled in the air; consolidated into a rock these are referred to as volcanic breccias and agglomerate respectively. Granule to pebble-sized particles (2–64mm) are called lapilli and form a lapillistone. Accretionary lapilli are spherical aggregates of fine ash formed during air fall. Sand-, silt- and clay-grade tephra is as hwhen unconsolidated and tuff upon lithification. Coarse ash/tuff is sand-sized and fine ash/tuff is silt-and clay-grade material.

(a)

Clast size	Unconsolidated	Consolidated
>64 mm	Bombs	Agglomerate
	Blocks	Volcanic breccia
2-64 mm	Lapilli	Lapillistone
0.063-2 mm	Coarse ash	Coarse tuff (volcanic sandstone)
<0.063 mm	Fine ash	Fine tuff (volcanic mudstone)

(b)



(a) The classification of volcaniclastic sediments and sedimentary rocks based on the grain size of the material. (b) Nomenclature used for loose ash and consolidated tuff with different proportions of lithic, vitric and crystal components.

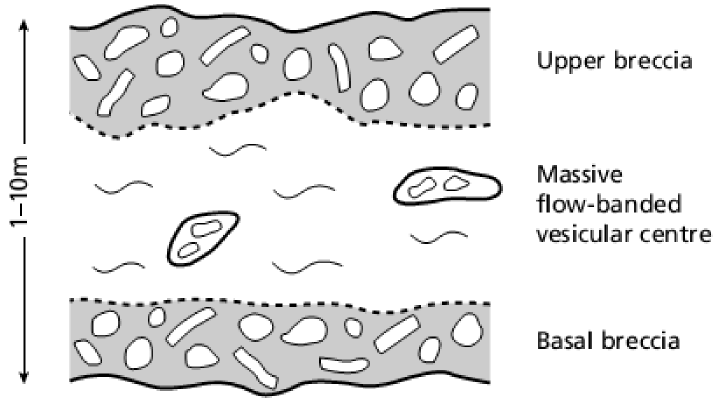
2- Based on their modes of formation

Five principal categories can be distinguished: autoclastic, pyroclastic-fall, volcaniclastic-flow, hydroclastic and epiclastic deposits.

a- Autoclastic deposits

These are volcanogenic rocks produced by autobrecciation of lavas. As a lava flows along, it cools and the upper surface may develop a brittle crust, which fractures and brecciates on top of the moving lava. As the lava advances, the brecciat-ed material slides off the front of the flow and is then overriden to give a basal breccias. The textures can vary from clast-supported to matrix- (lava-) supported.

Generally , In the more distal parts of lava flows, the whole of the deposit may be flow breccia, whereas in more proximal areas, the breccias just occur at the bottom and tops of the more massive lava beds.



Typical succession through flow-brecciated lava. After Suthren (1985)

b-Pyroclastic-fall deposits

These sediments are simply formed through the fallout of volcanic fragments ejected from a vent or fissure as a result of a magmatic explosion. The characteristic features of air-fall deposits are a gradual decrease in both bed thickness and grain size away from the site of eruption, and a good to moderate sorting. Blocks and bombs are deposited relatively close to the vent, whereas ash may be carried many tens of kilometres and dust thousands of kilometers away from the vent. Indi-vidual beds of air-fall material typically show normal grading of particles, although in some cases, invers grading of pumice and lithic clasts has occurred.

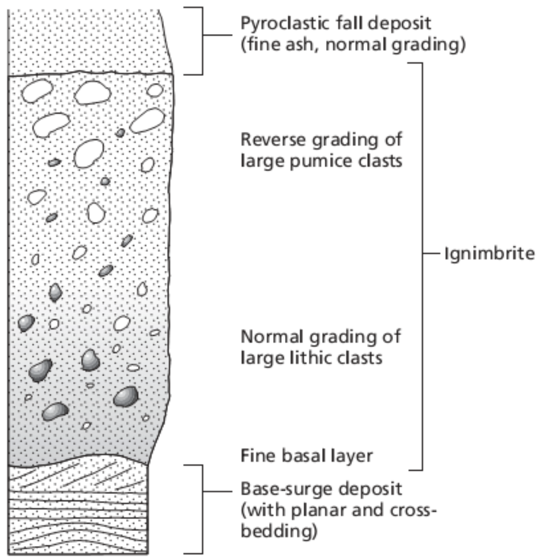
Where deposition takes place in water quite large frag-ments of low-density pumice may occur towards the top of an air-fall bed, as a result of the pumice floating on the water surface before deposition.

c- Pyroclastic-flow deposits

Pyroclastic-flow de-positsare the product of hot gaseous particulate den-sity currents. They generally form through fluidization by magmatic gas and give rise to deposits known as **ignimbrites.**

d-Ignimbrites

These pyroclastic ash-flows are generated by the col-lapse of eruption columns and they are hot, dense, laminar flows of volcanic debris. The deposits of these flows, ignimbrites, are characterized by a homogeneous appearance with little sorting of the finer ash particles, but if coarse lithic clasts are present they may show normal size grading, and large pumice fragments are commonly reversely graded. There generally is a lack of internal stratification, although bedding may develop in more distal regions. Pumice clasts may be concentrated on the top surface of an ignimbrite. Ignimbrites typically are derived from acid magmas. Ignimbrites also may be intrusive into previously deposited pyroclastic-flow deposits.



Diagrammatic section through the deposits of an ignimbrite eruption

E- Pyroclastic-surge deposits

Pyroclastic surges are dilute, subaerial, fast-flowing turbulent mixtures of volcanic particles and gas. Pyroclastic-surge deposits are generally thin and fine grained, but with a thicker accumulation in depressions. Grain size and bed thickness decrease away from the volcano and erosive bases and channel struc-tures are common. The deposits show evidence of downflow decrease in turbulence. The distinguishing feature of base-surge deposits is the presence of stratification, and planar and cross-bedding.

F= Lahar deposits

Lahars or volcanic mudflows occur on the slopes of some subaerial volcanoes. Cold lahars are mostly pro-duced by heavy rain falling on unconsolidated ash. Hot lahars are formed when a pyroclastic flow enters a lake or river or when air-fall ash is dumped into a crater lake.

g- Epiclastic volcanogenic deposits

Volcaniclastic sediments can be reworked in the sedimentary envi-ronment in the same way as any other sediment. In continental settings, volcanic ash is carried into river systems and lakes by surface runoff. In the shallow-marine environment, ash will be reworked by waves, tidal and storm currents and mixed with non-vol canogenic material. Redeposition of hyaloclastites in the deep sea by sediment gravity flows has been noted already. Thus all the various depositional sedimentary structures can be found in reworked volcaniclastic deposits. Being relatively soft and friable, volcanic debris is easily broken down into finer grades and rounded by abrasion in moderate- to high-energy environments.