CHAPTER FIVE

DIAGENESIS

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In the studied sandstones, a variety of diagenetic alterations which strongly affected the original composition of the present sediments have been considered. Four diagenetic stages are recorded: 1) eodiagenesis, 2) immature mesodiagenesis (mechanical compaction), 3) mature mesodiagenesis and 4) telodiagenesis.

5.1 Diagenetic features

The detailed and comprehensive petrographic studies conducted on more than 150 thin sections and 100 samples examined by the scanning electron microscope showed that the studied sandstones are affected by several diagenetic processes which include:

1. Mechanical infiltration of clay minerals.
2. Dissolution and leaching of unstable detrital silicate mineral grains such as feldspars and ferromagnesian minerals and, consequently, the development of secondary porosity.
3. Formation of an authigenic mineral assemblage that includes quartz, calcite, iron minerals (mainly hematite) and clay minerals (mainly kaolinite).
4. Development of compaction features such as closer packing, mechanical fracturing, deformation and bending of ductile and cleavable grains.
5. Dissolution of early-formed carbonate cements, followed by fracturing and secondary porosity initialization.
6. Development of late diagenetic cements consisting of quartz, K-feldspars, kaolinite, iron minerals, calcite, gypsum and halite.

5.1.1 Mechanical infiltration of clay minerals

Mechanically infiltrated clays represent the earliest diagenetic product in the studied sandstones. They were introduced to the sediments shortly after deposition as a result of infiltration of water containing suspended clay particles (cf. Wilson and Pittman, 1977). They are represented mainly by clay materials that partially or completely fill pores or interstitial voids (PL. 4.34A, B). In samples with intense clay infiltration, the framework grains exhibit normal packing and lack quartz overgrowths (PL. 4.34A), the infiltrated clays probably inhibited the nucleation of quartz cement (cf. Molenaar, 1986).

The process of mechanical infiltration of clays decreased the textural maturity and changed the original bulk composition of the rocks by adding clay minerals which were not present at the time of deposition (cf. Walker et al., 1978; Molenaar, 1986). It is suggested that these infiltrated clays, which are distributed on grain surfaces and in pores in a manner similar to authigenic clays, most probably exerted a significant influence on subsequent diagenetic processes by promoting pressure solution, inhibiting quartz cementation, and acting as nucleation substrates for authigenic clays (cf. Matlack et al., 1989). Several authors suggested that early diagenetic clays in certain ancient sandstones may have originated by infiltration (Kessler, 1978; Turner, 1980; Loucks et al., 1984; Molenaar, 1986). Moreover, the results of Crone (1975) and Walker et al. (1978) suggested that infiltrated clays may be common constituents in ancient sandstones of various non-marine depositional facies.

5.1.2 Dissolution and Leaching

Dissolution is the diagenetic process by which a solid component in the host sediment is dissolved by an aqueous pore solution leaving behind a space or cavity within the host sediment (cf. Schmidt and McDonald, 1979a, b; Burley and Kantorowicz, 1986). In the continental sandstones, dissolution of unstable ferromagnesian minerals such as pyroxenes and amphiboles is a common diagenetic process. Although the dissolution process releases many ions that are considered as source for the formation of different authigenic minerals, it affects the cement as well as the detrital grains at different burial depths. The alteration process includes the dissolution and the *in situ* replacement of clay. The degree of dissolution depends on the mineral stability (Keller, 1969), pore water chemistry, temperature and meteoric water composition.

These processes resulted in the development of about 8.36, 5.80 and 0.96 % oversized pores (both empty and filled with authigenic cement) in the Naqus, Araba and subsurface sandstones, respectively (PLs. 4.16C, D; 4.35; 4.41A, B; 4.44A, B; 4.46B; 4.49C; 5.1A, B). These pores resulted from partial or complete dissolution of detrital grains, most probably ferromagnesian minerals, feldspars and rock fragments (cf. Schmidt and McDonald, 1979). Based on data of McBride (1985) and Milliken et al. (1989), 80 % of these pores were feldspars and the other part was rock fragments. However, a proportion of these oversized pores could be due to dissolution of early-formed carbonate cements.

Ferromagnesian minerals may have been originally scarce in the studied sandstones, as the source rocks were sedimentary, metamorphic and plutonic igneous in a decreasing order. Complete dissolution of these grains leaves no clue about their nature. However, the relics or complete dissolution of grains could be seen where the original grains were completely leached leaving iron minerals or clay rim (PLs. 4.16C; 4.47D). Such selective dissolution is due to the chemical stability of this post depositional film in the interstitial pore water (Walker et al., 1978). On the other hand, dissolution of cleavable grains takes place by hydrolysis. Na, Ca and Mg are released from the silicate framework rather easily, whereas Fe, Al and Ti tend to remain behind (Turner, 1980).

K-feldspar grains in the studied rocks may have been originally more abundant than ferromagnesian minerals and rock fragments. Extensive intrastratal dissolution leads to complete dissolution of these grains, however relics of the dissolved feldspars are recorded (PLs. 4.16A, B; 5.1C). In general, the dissolution takes place preferentially along the cleavage planes of the grains as well as their peripheries. The released K, Al and Si produced during the breakdown of feldspars may be responsible for the formation of the authigenic quartz overgrowths and clay minerals.

Micas are rare in the studied sandstones. They do not show obvious dissolution features. The complex behavior of micas in soils may yield diverse alteration products depending on the precursor mineral and the interstitial environmental conditions (Turner, 1980). Dissolution of rock fragments, particularly those made up of chert, was locally recorded (PLs. 4.17D; 5.1A-B).

The intrastratal solution has also affected quartz grains (PLs. 4.46D; 4.50A; 5.1D; 5.2A). According to McBride (1985), quartz grains undergo dissolution along strained zones and microcrystalline boundaries to develop dissolution channels in intensely-weathered sandstone. The dissolution may be continued after compaction as indicated by dissolution of quartz grains along microfractures and outer rims resulting in an increase in porosity (PL. 5.1D). The effect of dissolution on quartz grains may extend to their authigenic overgrowths as well (PLs. 4.25B; 4.46D; 5.2B, C).

Dissolution of authigenic minerals, particularly calcite, is a common feature in the studied sandstones. This is expressed by the development of loose packing (PL. 5.2D), presence of calcite relics (patches) within pores and/or corrosion of quartz overgrowths (PLs. 4.25B; 4.46D; 5.1D; 5.2A-C). The irregular oversized pores, larger than the surrounding corroded quartz grains, may have been originally filled by carbonate (Pittman, 1979; Abdel-Wahab, 1988; 1999) that were dissolved later (PLs. 4.46B; 4.49C). Dissolution of halite cement and, rarely, clay clasts is a common feature in the studied sandstones (PLs. 4.48A-B, D; 5.3A-C).

The heavy minerals were also drastically affected by dissolution. Almost all the unstable and most of the metastable heavy minerals were dissolved and, consequently, the heavy mineral suite in the studied sandstones is mainly represented by the ultrastable zircon-tourmaline-rutile group. Moreover, minerals of this group sometimes display dissolution features (PL. 5.3D) which indicates the strong effect of intrastratal solution.

5.1.3 Authigenesis

Authigenesis, literally means "generation *in situ*", is usually applied to describe all diagenetic mineral formation in sediments. Authigenic minerals are thus distinct from detrital (transported) minerals and formed *in situ* within the host sediment in which they now occur (Worden and Burley, 2003). In the studied sandstones, authigenic minerals belong to two groups: non-clayey minerals (such as quartz, feldspars, iron minerals, calcite, gypsum and halite) and clay minerals (mainly kaolinite). Of the various authigenic phases present, only micritic calcite and evaporate minerals are incompatible with deposition from meteoric water.

5.1.3.1 Quartz

Regarding silica sources, timing of cementation, controls on cement distribution, and the chemical and hydrologic conditions under which quartz cement precipitates, the questions have not yet resolved in a conclusive manner (cf. McBride, 1989). These questions will be briefly addressed in the following sections.

5.1.3.1.1 Cementation conditions

The mechanism of quartz precipitation from a cooling, saturated solution is the one most often invoked to explain the occurrence of quartz cement in a sandstone (cf. Land and Dutton, 1978; Dutton and Land, 1988). Presumably, pore water travels from more deeply buried (warmer) sediments and moves upward in a regional flow system, precipitating quartz as it cools. However, meteoric water that precipitated the first episode of quartz overgrowths in the studied sandstones was probably introduced into the rocks as a lateral flow system. The pore fluid that precipitated the second episode of quartz is believed to be descending meteoric water (cf. Salem, 1995).

The nature of silica phase reflects the chemistry of the solution from which it has been precipitated. The spectrum of silica polymorphs (opal, chalcedony, megaquartz and quartz overgrowth) is controlled largely by the degree of supersaturation of silica and the purity of the ground water (Millot, 1960; Williams and Crerar, 1985; Thiry and Millot, 1987; Thiry and Milnes, 1991). Highly concentrated silica solutions undergoing high rates of precipitation in the presence of impurities promote the formation of abundant nuclei that have low rates of growth. These conditions favor the precipitation of opal and microcrystalline quartz (chalcedony). On the other hand, solutions with comparatively low concentrations of silica and low levels of impurity ions can precipitate large quartz crystals and quartz overgrowths. Therefore, the abundance of megaquartz and quartz overgrowths relative to cryptocrystalline and microcrystalline quartz in the studied pedogenic silcrete suggests formation from diluted solutions.

5.1.3.1.2 Variation in quartz cementation

Two episodes of quartz cementation have been recognized in the studied sandstones. The first is the ordinary quartz overgrowth associated with shallow burial. The second is a late stage quartz overprint.

Petrographic and SEM examination distinguished two generations of quartz. The first is characterized by the presence of dust lines separating successive quartz overgrowths (PL. 4.4C). The second involved the development of prismatic bipyramidal crystals which grew into secondary pores indicating a late stage quartz cementation which followed feldspar dissolution (PL. 4.30D).

5.1.3.1.3 Silica sources

Many different sources of silica have been proposed as the source of quartz cement in sandstones. McBride (1989) summarized 23 proposed sources since Sorby (1880) till Land et al. (1987). Most authors recognize the likelihood that silica in quartz cements was derived from more than one source, but many interpret a particular source for a given formation to have been dominant. Pressure solution of quartz grains in sand and smectite-illite conversion in shale are the most often recently cited sources of silica for moderately to deeply buried sandstones. Pittman (1979) listed the replacement of silicates by carbonate minerals and decomposition of feldspars as two additional important sources. In addition, many workers advocate the meteoric water as a major cementing agent. The source of silica in meteoric water is from weathering or from dissolution of silicates in the shallow subsurface.

Blatt (1979) has argued that only meteoric water possesses sufficient dissolved silica and flux to produce abundant quartz cement that is present in many quartz arenites. However, the literature suggests that the amount of quartz cement precipitated by meteoric water in most ancient formations is quite small in non-silcrete forming conditions, and that only in silcrete sandstones is much quartz cement present (cf. Walker, 1967; Kantorowicz, 1985; Molenaar, 1986; McBride, 1989). This is compatible with this study where the amount of quartz cement in non-silcrete sandstones is very little compared with that in silcrete.

Sorby (1880) inferred that quartz cement was derived from the decomposition of feldspars, and Fothergill (1955) suggested specifically their kaolinization. In the studied sandstones, much K-feldspars were lost by dissolution rather than through replacement. Sufficient silica was released during this dissolution and provided the amount of quartz cement recorded in the highly-cemented sandstones. However, much of the silica released by feldspar dissolution certainly ends up in authigenic kaolinite, and more important the time of major feldspars dissolution and kaolinite generation in the studied sandstones was much later than the time of early stage quartz cementation, a feature which was recognized in other rocks (cf. Land et al., 1987).

Like many other sandstones, quartz cement in the studied sandstones was probably precipitated by formation waters with a strong meteoric component (cf. Fisher, 1982; Dutton and Land, 1988). Herein, the sources of silica for the late stage quartz cement were variable. Silica may have been released from diagenetic reactions involving silicate minerals such as feldspars dissolution and replacement, and dissolution of detrital quartz grains. Pressure solution and clay mineral diagenesis are also proposed.

Kaolinization may contribute, in part, to the population of silica. A possible intraformational source of silica was that released when feldspars were altered to kaolinite. SEM observations show some kaolinite booklets embedded in quartz overgrowths (PL. 4.36C). Thus, quartz cement locally preceded and followed the precipitation of kaolinite. McBride (1989) concluded that silica released by feldspars alteration may form the common, but small amount, of late stage quartz cement that grows in secondary pores as tiny crystals that apparently lack overgrowth fabric, a common texture in the studied silcretes.

Replacement of quartz and other silicate grains by carbonate is another possible process which provided silica for the late stage quartz cement. The importance of silica-carbonate replacement and their reversibility was reported by Walker (1960), Burley and Kantorowicz (1986) and McBride (1987). The replacement of quartz by calcite is a common feature in the studied sandstones. The margins of detrital quartz grains and their overgrowths show irregular contacts where calcite cement borders them (PLs. 4.25B; 4.46D; 5.1D). These embayed surfaces are considered to be evidence of corrosion of quartz by calcite. This finding is in agreement with that reached by Burley and Kantorowicz (1986) who demonstrated that the surfaces of detrital quartz grains were embayed and corroded by calcite cement to a significant degree.

Many authors have cited pressure-solution of detrital quartz and other silicates at grain contacts in sandstone as an important source of silica for quartz cement. Moreover, other workers (c.f., Waldschmidt, 1941; Pittman, 1972; Füchtbauer, 1974a), referred it to be the most important source, especially in sands that have undergone significant burial. Stylolite seams represent another pressure solution source that was suggested for sandstones (c.f., Heald, 1955; Dutton, 1986). However, Dutton (1986) and Houseknecht (1988) observed that most quartz cementation in sandstones had taken place before the commence of pressure solution in the same formation. Thus, silica released by pressure solution in these sandstones could not have been a major local source of silica for cement. In the studied surface sandstones, which are characterized by the rarity of sutured contacts and stylolites, silica released by pressure solution is believed to have minor contributions in the formation of late stage quartz cement. On the other hand, silica released by pressure solution is believed to be the main source for the formation of late stage quartz cement in the subsurface sandstones, which are characterized by the abundance of sutured contacts and stylolites (PLs. 4.40A, B; 5.5; 5.6) and, the rarity of clays and ductile grains which could have minimized the effect of compaction if it were present.

The sources of silica for silcretes remain largely conjectural, but the consensus is that silica is derived from weathering reactions of silicate minerals, including clays, and dissolution of quartz in bed rock and alluvium (Summerfield, 1983; Thiry and Millot, 1987). The role of pH and organic acids in promoting silica dissolution also remains debatable (McBride, 1989). Following Millot (1960), Summerfield (1983) and Williams and Crerar (1985), quartz overgrowths and megaquartz in the studied silcrete are attributed to precipitation from ground water low in silica and low in impurity ions in contrast with microquartz that precipitate from water enriched in dissolved silica and impurity ions (Salem, et al., 1998).

5.1.3.2 K-feldspars

Authigenic K-feldspars constitute a negligible component in the studied sandstones. Commonly they are found as coarse crystals cementing quartz grains (PLs. 4.32; 4.33). These crystals are not connected to detrital feldspar cores, but occur as isolated crystals partially filling pore spaces. Such feature is known to develop when the pore water becomes supersaturated with respect to K, Al and Si. Dissolution and alteration of authigenic K-feldspars in the studied sandstones have been locally recorded (PL. 4.33B-C).

5.1.3.3 Clay minerals

The recorded authigenic clays in the studied sandstones are represented by kaolinite together with traces to minor proportions of illite, smectite, illite-smectite mixed layer and chlorite as revealed by XRD and SEM examinations.

5.1.3.3.1 Kaolinite

Authigenic kaolinite is present in all the studied samples. It appears to have been precipitated after the formation of quartz cements (PL. 4.26). However, a few kaolinite particles are engulfed in quartz overgrowths which suggests synchronous development (PL. 4.36C), however, most kaolinite crystals rest on them and, thus, are younger. Diagenetic kaolinite displays morphological differences throughout the studied sandstones. It infills both primary and secondary pores and replaces feldspars and micas (PLs. 4.35-4.38). Kaolinite, which formed as a result of feldspars and/or micas alteration, differs in shape from the neoformed type. The latter generally preserves the precursor grain habit (PL. 4.38).

The most likely source of Al and Si for kaolinite formation was the dissolution of detrital K-feldspars. Detrital K-feldspars in the studied sandstones were commonly observed to have suffered significant alteration and dissolution during late diagenesis. Preservation of extremely delicate K-feldspars in the studied sandstones (PLs. 4.15B, C, 4.16A, B; 5.1C) shows feldspars dissolution to be relatively a late event, resulting in kaolinite precipitation. Dilute acidic water seems to be the favorable environment for neoformation of kaolinite (Krauskopf, 1979). A typical reaction could be (Hayes and Boles, 1992):

$$2\ KAlSi_3O_8 + 2H^+ + H_2O \rightarrow Al_2Si_2O_5(OH)_4 + 4\ SiO_2 + 2\ K^+$$

Orthoclase Kaolinite Quartz

If the reaction components are not removed from the sandstone, leached feldspars will precipitate 60 % kaolinite and 40 % quartz (Bjørlykke, 1983). The sharp, strong peaks of kaolinite and the presence of stacked kaolinite "books" may indicate coarse grained clays and a higher degree of crystallinity (Grim, 1953) that may favor a diagenetic origin of these kaolinites.

5.1.3.3.2 Illite, Smectite and Illite-Smectite mixed-layer

Detailed investigations by using XRD and SEM techniques revealed that authigenic illite, smectite and illite-smectite mixed-layer are present locally in the studied sandstones.

5.1.3.3.3 Chlorite

Traces to minor quantities of authigenic chlorite were detected in the subsurface sandstone. Authigenic chlorite was observed by SEM examinations, but was not detected by XRD diffractometry, this is attributed to its insignificant content. However, one sample from RB-A3 well (depth 12432 feet) was found to have exceptionally large proportion of Fe-rich chlorite (up to 7 % of the total volume).

Herein, chlorite formation is attributed to chloritization of kaolinite. Boles and Franks (1979) recorded that chloritization of kaolinite occurs at burial depths between about 3500 and 4500 m (165 - 200°C). The reaction was envisaged to be:

$$3Al_2Si_2O_5(OH)_4 + 3.5Fe^{2} + 3.5Mg^{2+} + 9H_2O \rightarrow Fe_{3.5}Mg_{3.5}Al_{6.0}Si_{6.0}O_{20}(OH)_{16} + 14H^{+}$$

Kaolinite Chlorite

Chloritization of kaolinite in Triassic red-bed sandstone of the Lunde Formation, offshore Norway occurs close to the oil-water contact at depths of about 2.5 km (100°C). The Fe^{2+} thus might be derived from the reduction of iron oxides (Curtis et al., 1985), as envisaged below:

$$2Al_2Si_2O_5(OH)_4 + 5Fe_2O_3 + 2SiO_2 + 4H_2O + 2.5C \rightarrow Fe_{10.0}Al_{4.0}Si_{6.0}O_{20}(OH)_{16} + 2.5CO_2$$

Kaolinite Hematite Quartz Organic Matter Chlorite

This reaction, therefore, may be controlled by CO_2 fugacity. The possibility of having the same reaction in the studied subsurface sandstone cannot be excluded.

5.1.3.4 Iron minerals and red beds

Many workers have discussed the origin of Fe-oxide pigments in sediments. There is a wide appreciation of the importance of pedogenesis in the formation of red hematitic pigment within soils (Birkeland, 1984; Schwertmann and Taylor, 1987; Schwertmann, 1988; Retallack, 1990). However, an equally important set of processes relates to the early, shallow burial alteration of sediments by ground waters (Wright et al., 1992), as in the case of the studied ground water ferricrete. Most workers agree on the post-depositional alteration of detrital iron-bearing grains in the formation of hematitic pigment in red beds (Walker, 1967, 1974, 1976; Walker et al., 1978, 1981).

The iron minerals in the studied red sandstone beds are predominantly of diagenetic origin and their occurrence reflects the prevalence of special conditions of intrastratal alteration of iron-bearing detrital grains after deposition.

Evidences of a diagenetic origin for the iron minerals in the studied sandstones include (cf. Walker 1967): 1) iron minerals coatings on quartz grains are absent at the grain contacts (PL. 4.27A), 2) some iron minerals are present as coatings of earlier authigenic minerals (PLs. 4.29C; 4.42; 4.44D), 3) the euhedral habit displayed by hematite crystals (PL. 4.43D), 4) the association of iron minerals with different types of cements (halite, calcite, kaolinite) reflecting different ground water chemistry, 5) the presence of iron minerals as a product of diagenetic alteration of unstable ferromagnesian minerals such as hornblende and pyroxene, 6) the presence of partially dissolved and replaced feldspars and rock fragments (PLs. 4.15; 4.16; 5.1A-C) which indicates that grain alteration has progressed to the point that minerals more stable than ferromagnesian silicates were altered, 7) the presence of iron minerals filling secondary pores which originated by dissolution of detrital feldspars and chert fragments, and 8) the occurrence of iron minerals filling fractures in detrital quartz grains which resulted by burial compaction (PL. 4.2C, D).

The above mentioned evidences support the hypothesis that hematite pigment in the studied red sandstones has formed as a product of the alteration of unstable detrital Fe-rich grains. The most commonly altered sand grains in red sandstones in general are ferromagnesian silicates (e.g., hornblende, augite and biotite), volcanic rock fragments, magnetite and ilmenite (Walker, 1967; Walker et al., 1978; Pye, 1983). Iron released during grain alteration enters solution as ferrous ion under reducing conditions and is precipitated under oxidizing conditions as iron hydroxides or possibly as hematite in the initial stage of formation. Iron hydroxides subsequently convert to hematite.

No distinctive colors for the identified textures of hematite have been established. However, the study reveals that the degree of "purpleness" in the red sandstone beds seems to depend mainly on the clustering of hematite crystals, that is, evenly distributed hematite produces red color, whereas clustered hematite produces purple color. This hypothesis is supported by the observation that grinding readily causes a change in color (from purplish red to yellowish red) of the red beds.

5.1.3.5 Calcite

Two types of calcite cement were recorded in the studied sandstones; namely: the micritic and the more common poikilotopic coarse spar. Both types locally fill intergranular and oversized pores. (PL. 4.44). Poikilotopic coarse spar is by far the most abundant.

Calcite cement was introduced into the studied sandstones shortly after quartz cement. The loose packing of the framework grains in some samples (PL. 5.2D) and the presence of the oversized patches of poikilotopic calcite indicate that cementation occurred before significant burial and mechanical compaction. A similar texture was recognized by Dickinson (1988) and Abdel-Wahab and Turner (1991). On the other hand, late-stage calcite cement was deposited slightly after recognizable compaction, particularly in the subsurface sandstone (PL. 4.46A).

Calcite precipitated in the studied sandstones was commonly corrosive to both the detrital quartz grains and their authigenic overgrowths. Evidence of quartz replacement by calcite has been reported in some samples in which quartz grains appear floating in a groundmass of calcite cement. These grains exhibit conspicuous embayed outlines (PL. 4.44A-C), which are typical of detrital quartz grains in some samples free of calcite cement (PL. 5.1D). It seems that corrosion started at the peripheral or weaker parts of the detrital grain, where calcite makes solid solution with quartz. These parts firstly have the same appearance of quartz in plane light and the same optical continuity and interference color of calcite under polarizing light (PL. 5.4). With aging and prolonged attack by calcite-saturated pore waters, these parts were replaced by calcite which resulted in the development of embayments. Girard et al. (1989) believed that the sharp outlines and embayments in quartz grains cannot be inherited, but reflect intense *in situ* dissolution and replacement by calcite. Some quartz overgrowths show similar notches and irregularities that were also the result of partial replacement by calcite (PLs. 4.25B; 4.45D; 5.2B, C).

5.1.3.6 Gypsum and Halite

Both gypsum and halite cements were recorded in the studied sandstones. It is most likely that these cements were formed very late in their diagenetic history. Because the formation of gypsum requires the presence of sufficient dissolved sulfate, the environment cannot be too reducing. Calcium and sodium ions as well as sulfates and chlorides came probably from the younger Middle Miocene evaporites that are exposed in the area. A less likely possible source of these salts is the sea water during their transgression over the Paleozoic rocks that accompanied the Suez rift in the Oligocene time.

A study of the origin of salts in ground water in southern Sinai using strontium isotopic composition was carried out by Strainsky et al. (1983). These authors concluded that the only possible major source of salts into the water is the air-borne particles that are derived from carbonate beach sands and sabkha sediments along the coast of the Sinai Peninsula. The dissolved minerals are halite, gypsum, anhydrite, barite, calcite and aragonite. About 10 % of the salts in water are derived directly from sea spray, while 90 % originate from air-born salts.

The results of the petrographic study carried out in the present work indicate a late diagenetic origin for gypsum and halite cements. There are three possible sources for the sulfate and chloride salts: 1) recycling from the uplifted Miocene evaporites, 2) sea water transgression during rifting, and 3) sea spray and air-born salts. The first source is more likely, but contributions of sea spray and air-borne salts cannot be neglected.

5.1.4 Compaction

Compaction is the process of volume reduction and consequential pore-water expulsion within sediments. Normally this takes place in response to vertical shear-compressional stresses owing to increasing weight of overburden, but the same processes operate under tectonic compressional forces. Two types of compaction are known: mechanical and chemical (Houseknecht, 1987).

The relative importance of these two compactional processes is difficult to establish and is, therefore, rather controversial. Depending on one's point of view, mechanical compaction may be a significant process that reduces intergranular volume to as little as 26 % or may be relatively insignificant process that hardly modifies the original intergranular volume established by the depositional medium (Houseknecht, 1987). If we assume that mechanical compaction can reduce intergranular volume to about 26 %, then any further reduction can occur only through chemical compaction. Alternatively, if we assume that mechanical compaction does not significantly reduce intergranular volume, then any reduction of intergranular volume below 45 % can be attributed to chemical compaction.

Herein, I believe that mechanical and chemical compaction work together to reduce the intergranular volume of the studied sandstone during shallow burial, an idea supported by other workers (cf. Füchtbauer, 1967). At the same time, any reduction of intergranular volume below about 26 % should be attributed to chemical compaction, which most commonly involves intergranular pressure solution. However, it appears that mechanical compaction does not necessarily reduce intergranular volume to the minimum possible value before pressure solution begins. For this reason, it is difficult to assess precisely how much of the reduction in intergranular volume results from mechanical compaction and how much results from chemical compaction, even in sandstones that have large intergranular volumes preserved (cf. Houseknecht, 1987).

The total amount of porosity lost by compaction of a sandstone can be expressed by the following equation (McBride et al., 1991):

$$\begin{aligned} \text{Total Ø lost by compaction} &= \text{Ø lost by pressure solution} \\ &+ \text{Ø lost by ductile grain deformation} \\ &+ \text{Ø lost by grain rearrangement} \end{aligned}$$

Framework grains in Araba Sandstone have typically undergone only minor pressure solution and have less than 1 % ductile grains. Therefore, it is assumed that compaction mostly occurred through grain rearrangement, and that both pressure solution and ductile grain deformation played only a minor role in reducing porosity. This hypothesis is supported by the high average intergranular volume (31.62 %) of Araba Sandstone. Only 18 % of Araba samples have intergranular volumes less than 26 % (Table 4.4). Grain rearrangement during the first tens of meters of burial can produce a loss of 5-12 % porosity (von Engelhardt, 1960), further rearrangement occurs when the grains move into tighter packing patterns.

Ninety-three percent of Naqus samples have intergranular volumes less than 26 % and 24 % of the samples have intergranular volumes less than 15 %. These data, together with the presence of appreciable concavo-convex and some sutured contacts may suggest that pressure solution has played an important role in reducing porosity, but still less than the role played by mechanical compaction. Ductile grain deformation played only a minor role in the porosity loss by compaction.

On the other hand, the subsurface sandstone is believed to have undergone so much chemical compaction. The nonspherical and poorly rounded appearance of most of its quartz grains can be attributed to shape modification by intergranular pressure solution that occurred throughout most of the sandstone. Additionally, the predominance of concavo-convex and sutured contacts, stylolitization and the near absence of floating grains indicate that a significant amount of pressure solution has occurred and played an important role in the reduction of porosity of the studied subsurface sandstone. This lithic-poor sandstone has an average intergranular volume of 17 %. Hundred percent of the subsurface samples have intergranular volumes less than 26 % and 27 % of the samples have intergranular volumes less than 15 % (Table 4.4). These results strongly suggest that significant porosity loss has taken place through pressure solution. Similar to the surface sandstone, ductile grain deformation has played a minor role in reducing porosity. No attempt was made to quantify this role.

From the aforementioned discussion, it is clear that compaction in the studied sandstones has occurred in response to four processes: grain rearrangement (rotation and slippage), ductile grain deformation, brittle grain fracturing and intergranular pressure solution (cf. Füchtbauer, 1967; Wilson and McBride, 1988). Ductile grains are essentially absent except as few clay-rips and mica fragments, and fractured grains, although present, are not abundant. Thus, compaction was essentially the product of grain rearrangement and intergranular pressure solution (PLs. 4.40A-B; 5.5; 5.6A-C). However, squeezed clay rips, mica and gypsum fragments (Pls. 4.6D; 4.18D; 4.19B-D) and fractured quartz grains (PLs. 4.2C, D; 4.3A; 4.50B; 5.6D) are also good features of pronounced compaction.

5.2 Diagenetic History

Paragenetic sequence is the interpreted order in which diagenetic processes occur successively. It is a simple way of relating a potentially complex series of events in a time series. The proposed generalized sequence of diagenetic events for the studied sandstones is illustrated in Figure 5.1. This sequence is based on the textural relationships of the various diagenetic products. It is important to note that textural relationships are relative and cannot be used to establish the actual timing of authigenesis. The proposed evolutionary pathway of the studied sandstones is presented in Figure 5.2. These two figures show that the paragenetic sequence may be traced as follows:

Shortly after deposition of the sandy sediments, suspended clay particles carried in alluvium were mechanically infiltrated into pore spaces. This infiltration occurred probably in the shallow subsurface, perhaps in the 0-50 m zone.

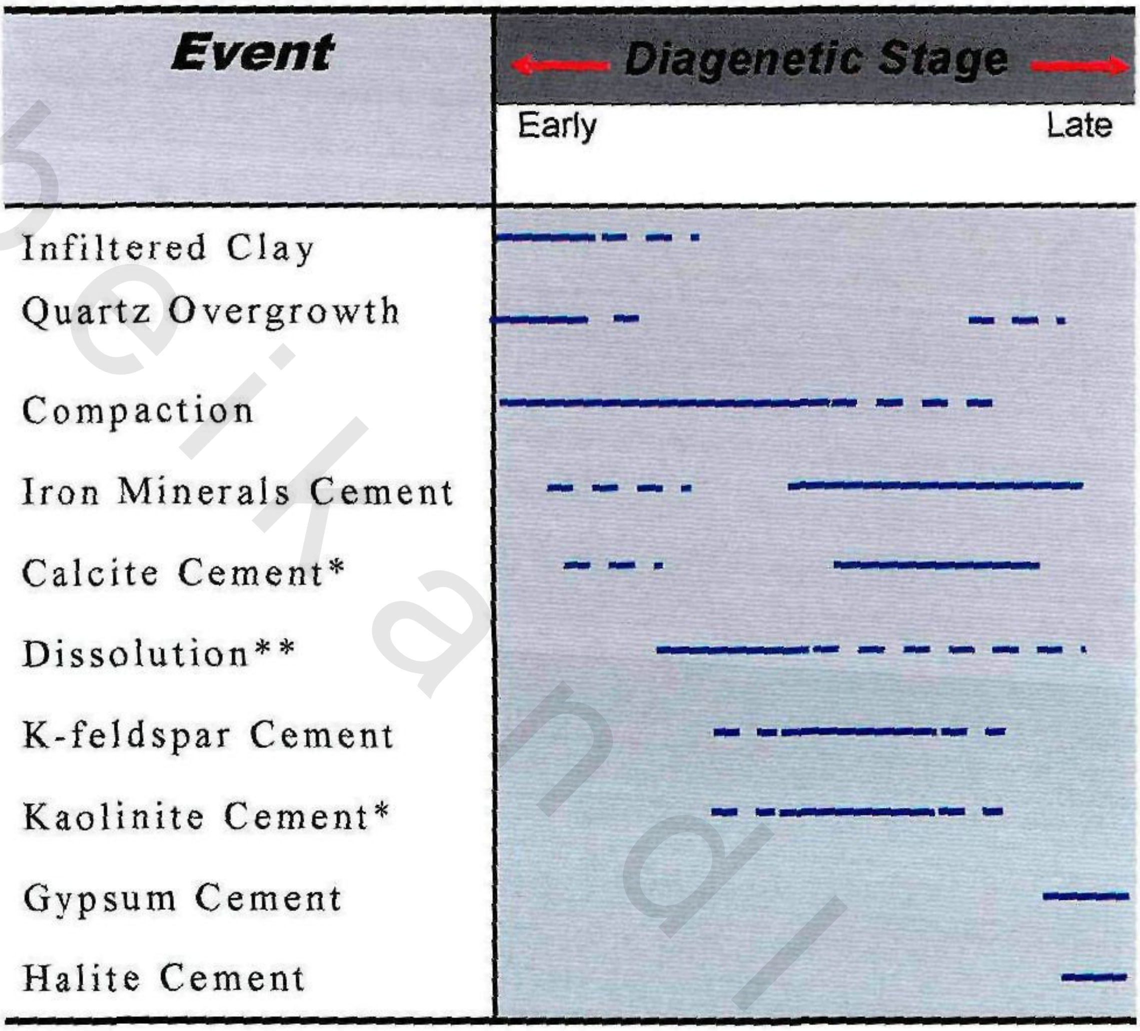


Fig. 5.1: Schematic diagram showing the relative timing of diagenetic events in the studied sandstones.

* Authigenic phase present as both pore-filling cement and replacive mineral.

** Dissolution of feldspars, rock fragments and calcite.

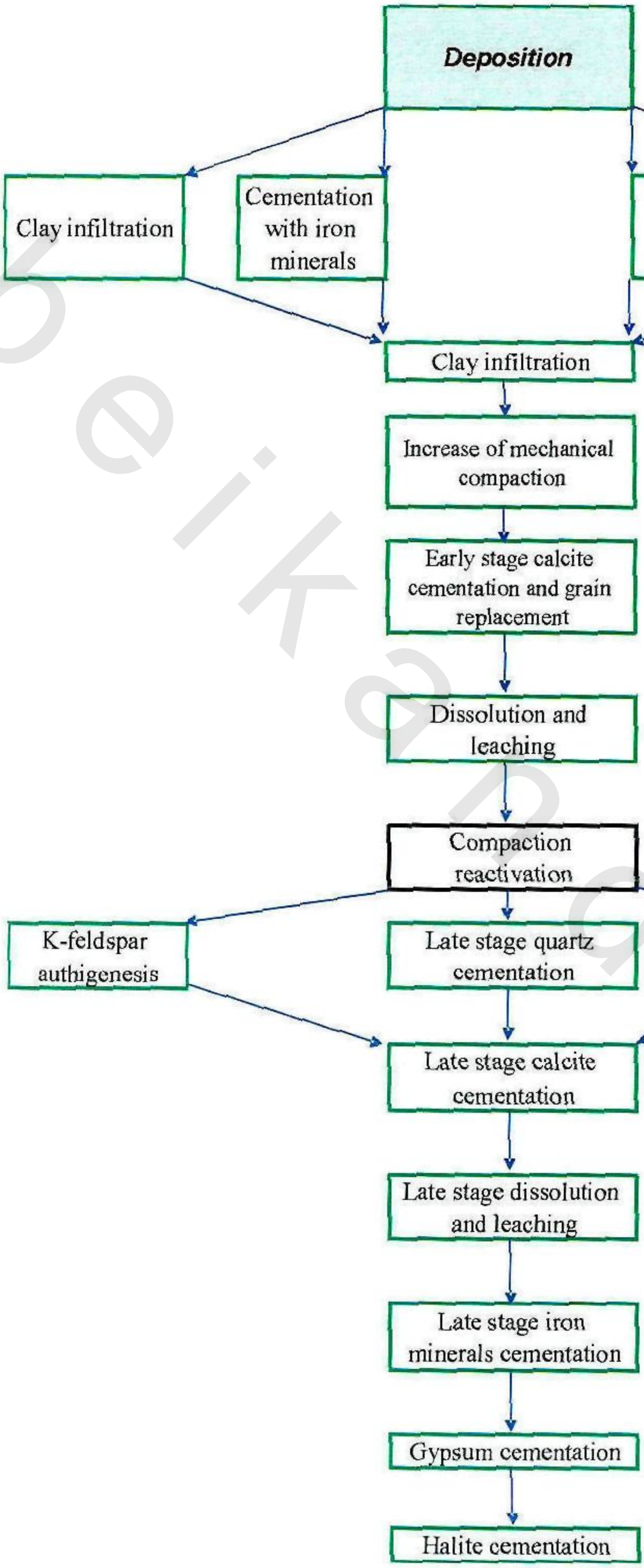


Fig. 5.2: Diagenetic evolution pathway for the studied sandstones.

Compactional processes started early and resulted in grain rearrangement, minor ductile grain deformation, rigid grain fracturing, pressure solution and an overall reduction in porosity. Compaction continued progressively but at a much lower rate and its effect was significant only in rocks that had not been made rigid by the introduction of quartz or calcite cement.

The early precipitation of quartz cement in the studied sandstones occurred soon after their deposition at shallow depths before the real burial compaction took place. However, the presence of dust lines between detrital cores of some quartz grains and their overgrowths may suggest the introduction of minor amounts of mechanically infiltrated clays before cementation by quartz overgrowths. Quartz overgrowths in calcite-cemented samples are commonly overlapped by calcite cements (PL. 4.25B-D; 4.45D) which indicates their earlier development. This is further confirmed by the fact that quartz cement does not encompass any other authigenic phase. The source of silica for the early diagenetic quartz cement is uncertain. Silica released during the major feldspars dissolution was generated too late to be the main source. Therefore, it is likely that most of the silica was imported possibly from meteoric water which was introduced during the early burial stage (cf. Blatt, 1979).

Another short episode of quartz cementation in cement-free sandstones took place at a late stage. In the subsurface sandstone, silica that had released from pressure solution was responsible for late stage quartz cementation. Heald (1956) mentioned that a limited amount of clay is believed to accelerate pressure solution, while in the presence of large amounts of clay hardly any pressure solution contacts are formed, probably due to the "Cushioning" effect of clay (Siever, 1959). De Boer (1977a, b) and Oelkers et al., (1996) pointed out that the dissolved silica must diffuse out of the grain contact, the presence of clay layers might conceivably promote the process. In silcrete samples, two successive stages of quartz cementation were observed. The first is the ordinary quartz overgrowth, and the second is the quartz overprint which is exclusively formed in silcrete.

Cementation with iron minerals commenced early in the burial history and continued for a long time. This cement is mainly represented by hematite that has two different generations: 1) early-stage hematite that was precipitated before significant compaction, and 2) late-stage hematite that postdated compaction and cementation with most other authigenic cements except for gypsum, halite and occasionally poikilotopic calcite. Late stage iron minerals cement was deposited in secondary pores and filled fractures in quartz grains (PLs. 4.2C, D; 4.40D; 4.41A, B). The iron mineral pigments in the studied red sandstone beds are diagenetic and originated through the intrastratal alteration of detrital iron-bearing silicates. However, Abdelwahab (1998) reported that iron minerals in sandstones may be introduced at any stage in the diagenetic history and, therefore, their existence is valueless for any diagenetic sequence interpretation.

Two different generations of calcite cement were recorded in the studied sandstones. These are the micrite and the coarsely-crystalline poikilotopic spar (PL. 4.44). The early-formed calcite was almost completely dissolved leaving behind a few micritic relics disseminated in clay and iron mineral cements. The loose packing of quartz grains, the floating of quartz grains in poikilotopic calcite cement in most calcite-cemented samples and the high intergranular porosity (pre-cement porosity) are good criteria for the deposition of an early-calcite cement that was dissolved. Most cementation by calcite preceded significant mechanical compaction and followed quartz cementation. The cross-cutting relationships between calcite cement and quartz overgrowths indicate that calcite emplacement took place after the main episode of quartz cementation (PLs. 4.25B-D; 4.45D). Occupation of secondary pore spaces by calcite is not common, indicating that cementation by calcite was prior to significant framework grain dissolution.

The significant part of feldspars and rock fragments dissolution probably occurred after the main episode of calcite cementation and resulted in changes in water chemistry due to the release of ions into pore water. Calcite dissolution started also at this stage. Kaolinite, K-feldspars and late stage iron minerals cement were deposited following the dissolution process probably during the late mesodiagenesis. Kaolinite commonly breaches secondary pore spaces (PLs. 4.35; 4.36) and is interpreted mainly as a late stage cement, however, the possibility of formation of early stage kaolinite cannot be excluded. Silica and alumina in the kaolinite were probably derived from detrital feldspars after invasion of the sandstones by meteoric water. Silica and alumina required for kaolinite genesis were probably derived from detrital feldspars as a result of their dissolution upon invasion of the sandstones by meteoric water. On the other hand, the components necessary for the precipitation of K-feldspar cement were derived from the early dissolution of detrital feldspars.

Plate 5.1

A) Chert aggregate displaying a sign of dissolution as expressed by the development of intragranular porosity (arrows). Note oversized pore (OSP). Plane light, sample Na-22.

B) A in crossed polars.

C) Photomicrograph showing relics of dissolved K-feldspars (arrows) in an oversized pore. Plane light, sample Na-31.

D) Photomicrograph showing a detrital quartz grain corroded and replaced by calcite which was later dissolved. Note the sharp embayments at the boundary of the quartz grain (arrows) and the developed high secondary porosity (stained blue). Plane light, sample Na-15.

Plate 5-1

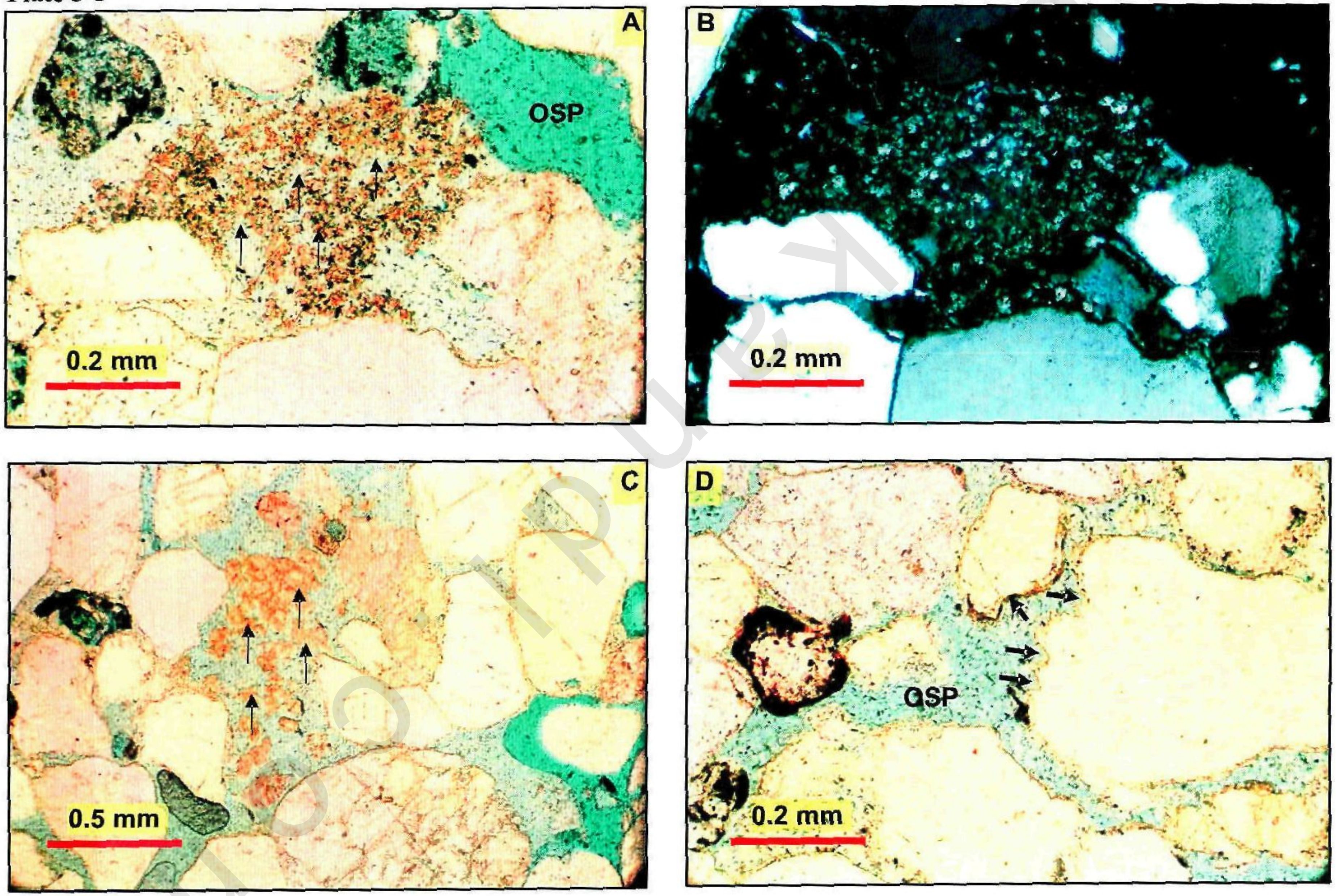


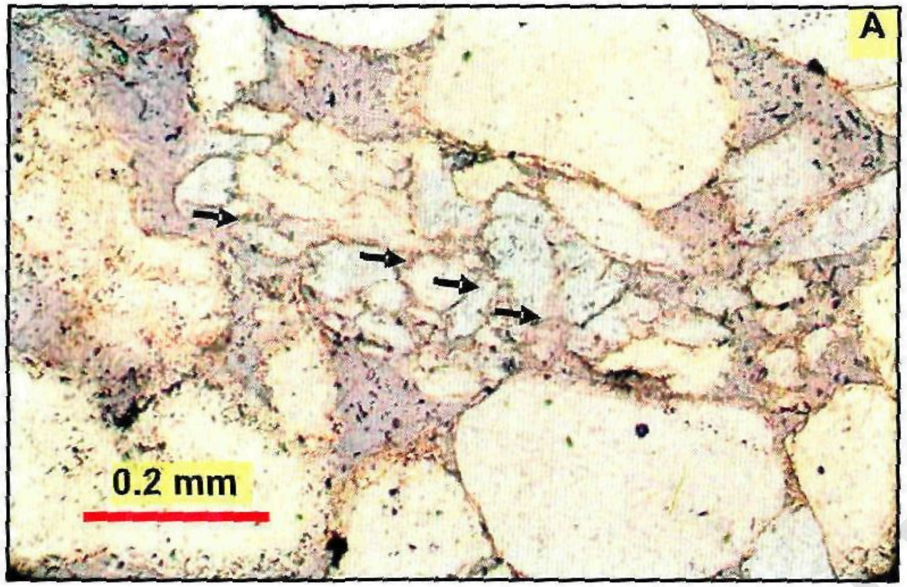
Plate 5.2

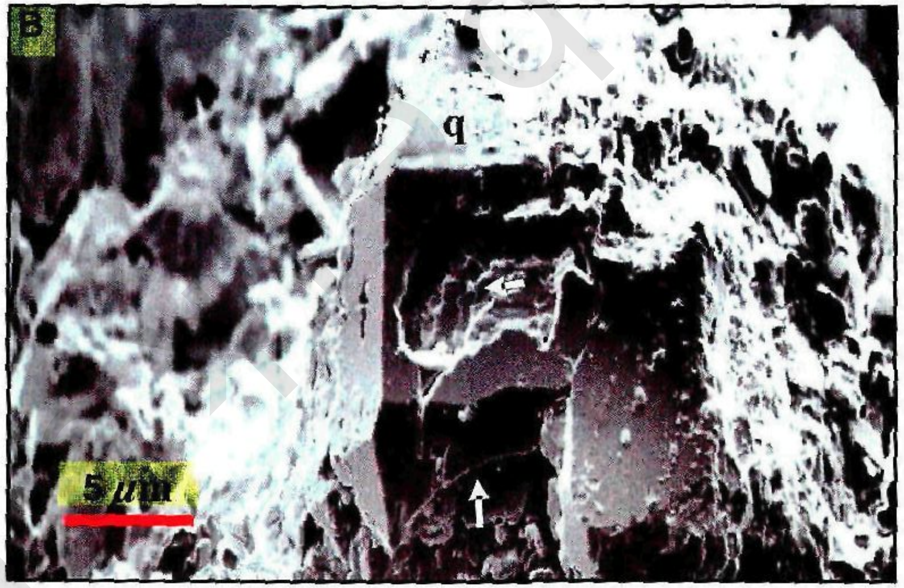
A) Photomicrograph showing selective dissolution along the intercrystalline boundaries of a polycrystalline quartz grain (arrows). Plane light, sample Na-80.

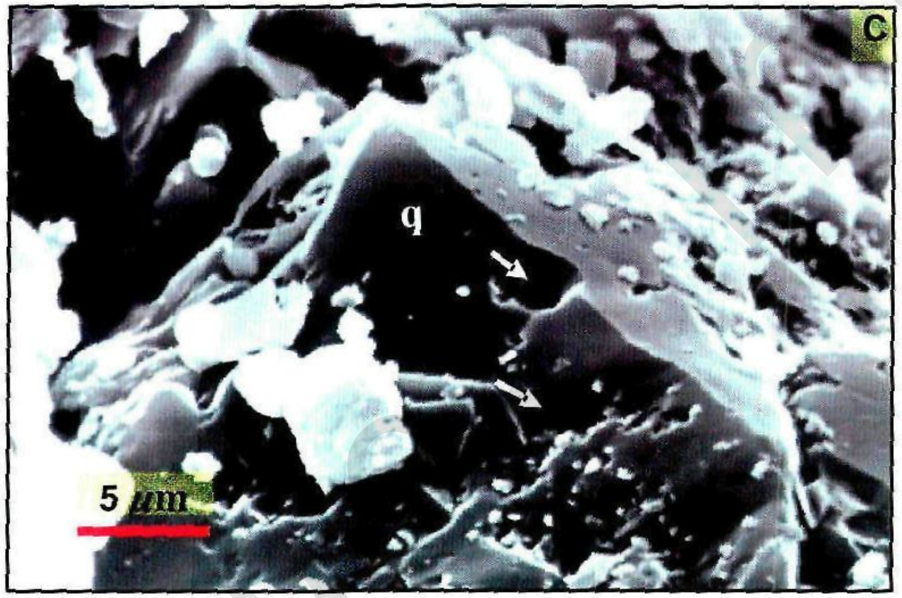
B, C) SEM micrographs showing corrosion and etch patterns (arrows) in quartz overgrowth (q). Samples Na-51 and Ar-4, respectively.

D) Photomicrograph showing loose packing. Plane light, sample Na-80.

Plate 5-2







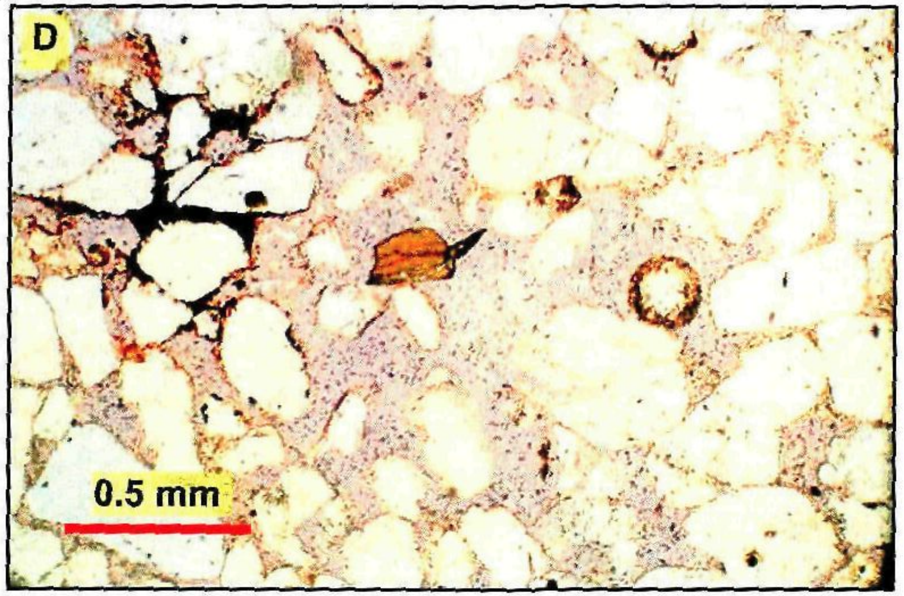


Plate 5.3

A) Photomicrograph showing dissolution in clay clast. Plane light, sample Na-80.

B) A in crossed polars.

C) SEM micrograph showing partial dissolution (Dis) of halite cement (Ha) which postdated kaolinite (Ka). Sample Na-26.

D) Photomicrograph showing partial dissolution of a zircon grain at its periphery which resulted in the development of embayments (arrows). Oversized pore (stained blue) probably resulted from dissolution of calcite cement. Plane light, sample Na-32.

Plate 5-3

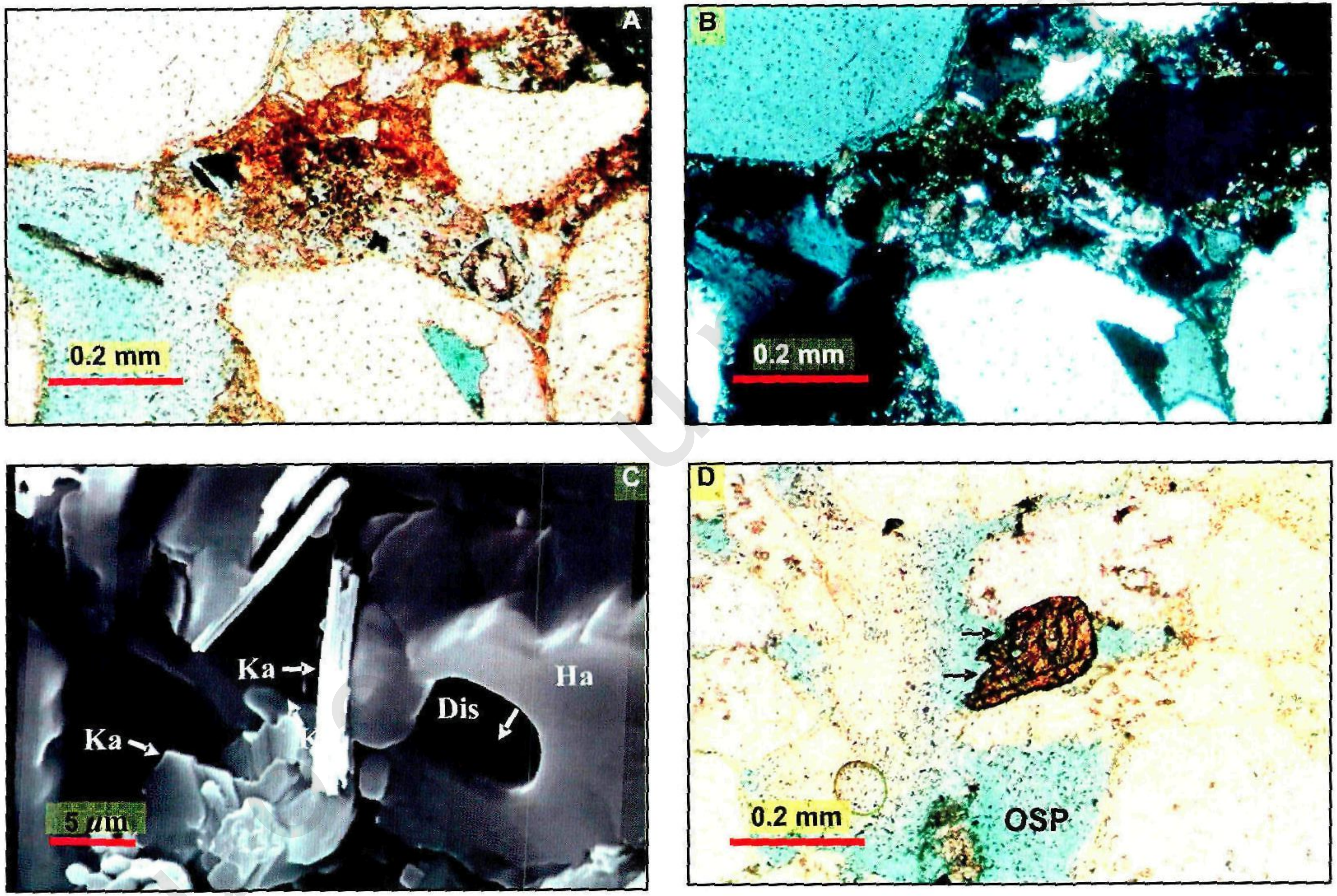


Plate 5.4

A, C) Photomicrographs showing poikilotopic calcite cement which corroded detrital quartz grain. The corroded parts have the same optical properties as the quartz grains (arrows). Plane light, sample Ar-7c.

B, D) Crossed polars of A and C. Note, the corroded parts are filled with calcite of optical properties as that of calcite cement.

Plate 5-4

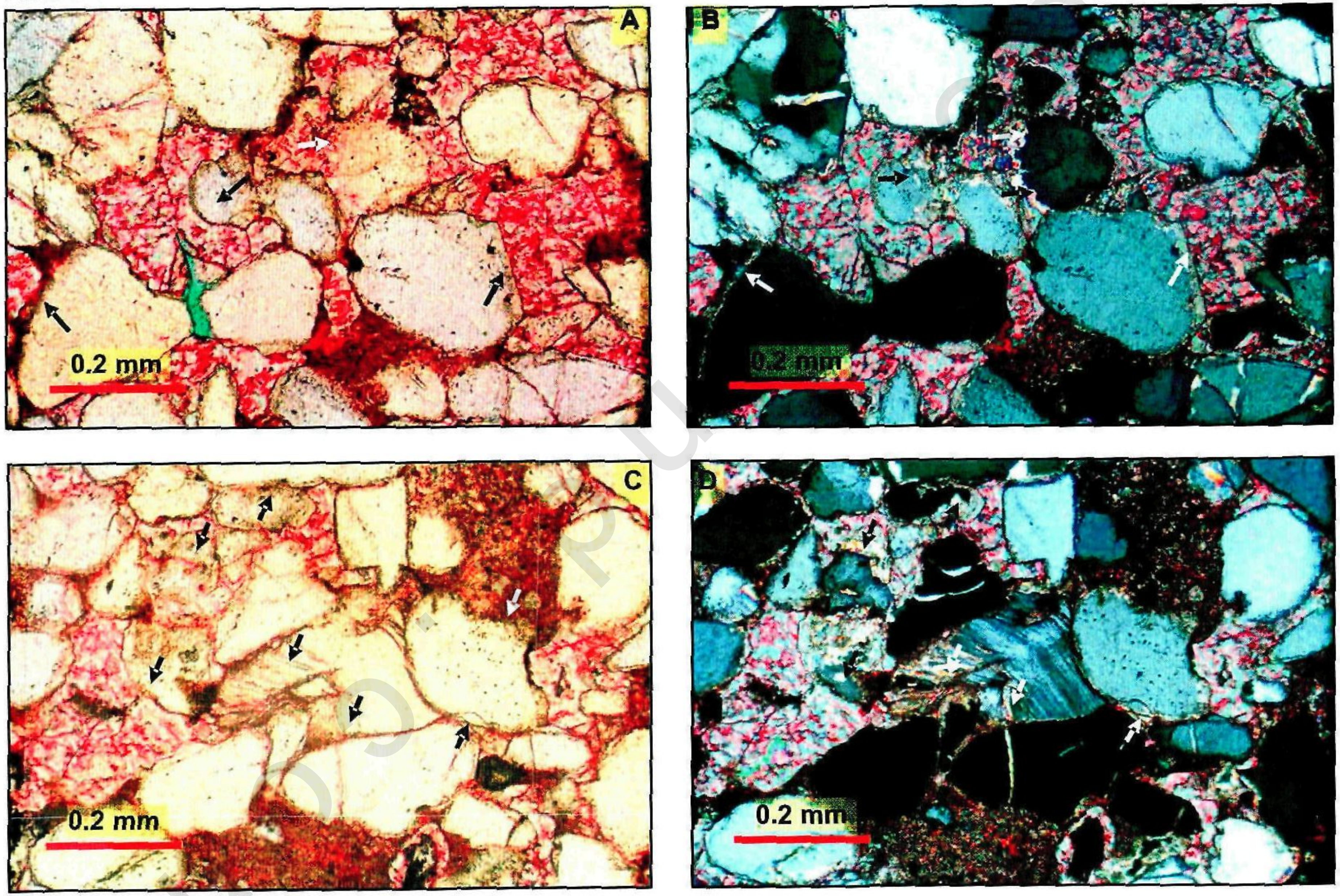


Plate 5.5

Photomicrographs showing probable grain-compaction and pressure solution texture. Quartz grains have irregular line contacts (long, concave-convex & sutured contacts) with adjacent grains rather than the point-to-point contacts found in undeformed sediments. Note chert fragment (Ch) penetrated by quartz grain indicating high durability of quartz rather than chert. The samples also show very little primary porosity (green stain) left after compaction. Plane light, samples RB-A1 (12341 ft) and RB-A5 (12808 ft), respectively.

Plate 5-5

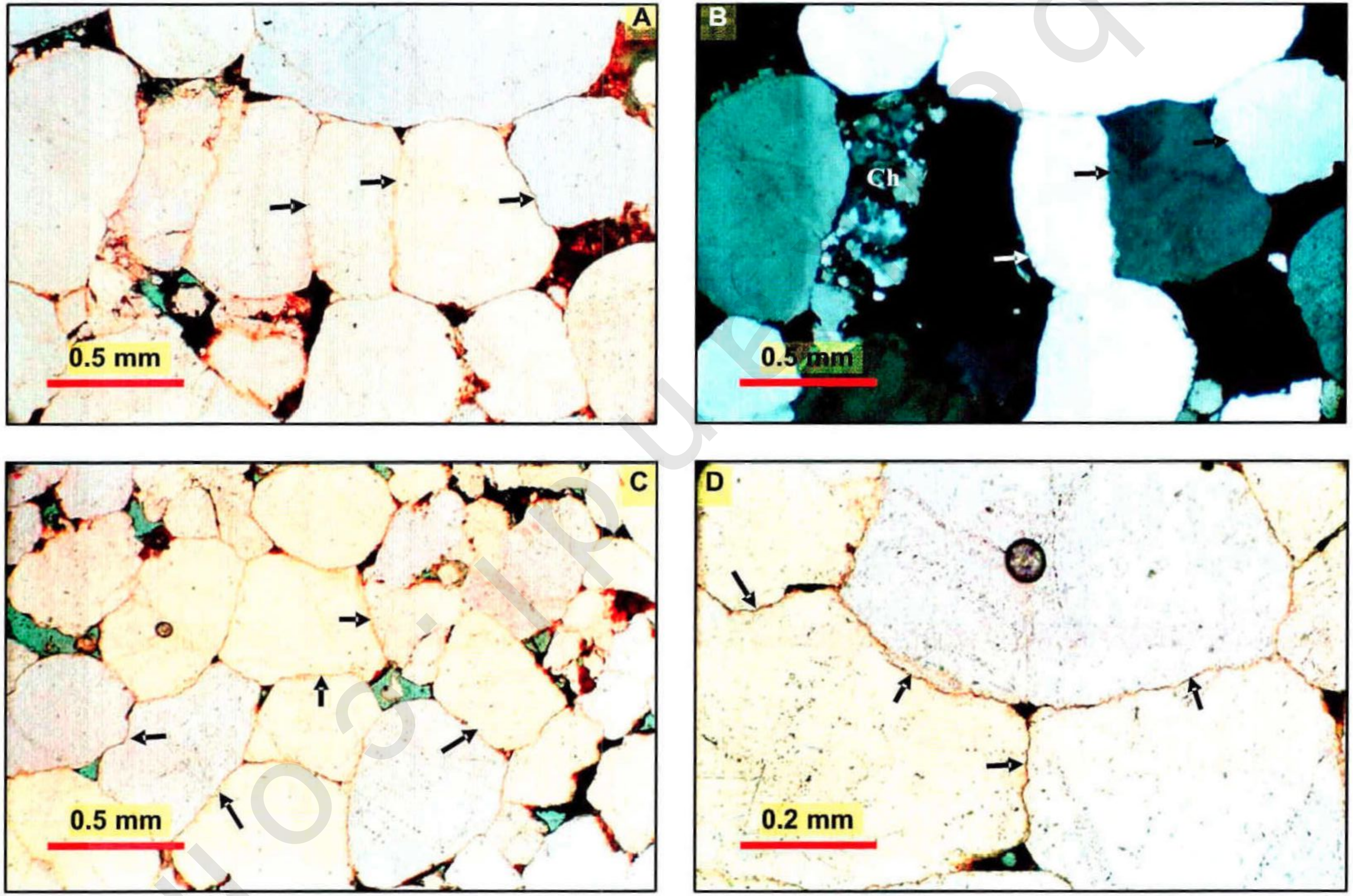


Plate 5.6

A-C) Photomicrographs showing stylolitization due to intergranular pressure solution. Plane light, samples RB-B4 (12404 ft) and RB-A5 (12832 ft), respectively.

D) Photomicrograph showing widespread fracturing of a detrital quartz grain due to compaction. Clay material filled the fracture planes (arrows). Plane light, sample RB-B3 (12392 ft).

Plate 5-6