

# Diagenesis in first-cycle desert alluvium of Cenozoic age, southwestern United States and northwestern Mexico

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## ABSTRACT

Petrographic studies of first-cycle desert alluvium of Cenozoic age in the southwestern United States and northwestern Mexico show that the mineralogy, texture, and chemical composition of the deposits have been changed diagenetically. The mineralogy has been changed by addition of mechanically infiltrated clay, partial removal of framework grains of feldspars and ferromagnesian silicates, and precipitation of authigenic potassium feldspar, zeolite, montmorillonite, quartz, hematite, and calcite. The texture has been changed by three processes: (1) infiltration of detrital clay and formation of authigenic montmorillonite, which form interstitial clayey matrix not present in the original sediment, (2) formation of voids where framework grains have been dissolved, and (3) *in situ* formation of silt and other fine-grained sizes. The chemical composition has been changed by infiltration of clay minerals that are richer in aluminum and lower in alkalis and alkaline earths than the original sediment and by removal in ground water of some of the ions released by dissolution and replacement of framework grains. These changes have significantly increased the mineralogical maturity and decreased the textural maturity of the sediments diagenetically.

Four major conclusions are drawn from the studies. (1) Some or all of the diagenetic alterations observed in these deposits probably occurred in many analogous ancient first-cycle alluvial deposits at a comparable time in their history. (2) Prolonged movement of ground water through first-cycle deposits may cause unstable minerals to be removed completely, or nearly so, leaving no direct evidence that they were important original constituents of the deposits. (3) Ancient first-cycle alluvium probably rarely, if ever, has the same mineralogy, texture, or chemical composition that the sediments had when deposited. (4) The present mineralogy, texture, and chemical composition of ancient first-cycle alluvial deposits probably do not accurately reflect lithology and climate in the source area or the nature of depositional currents and other environmental factors in the depositional basin.

## INTRODUCTION

Basin-fill deposits of Cenozoic age throughout the desert regions of the southwestern United States and northwestern Mexico consist predominantly of first-cycle sediments, which in most places are locally derived from crystalline rocks in adjacent mountain ranges. Sandstones and conglomerates in these deposits characteristically have high initial porosity and permeability. When first deposited, they are composed almost entirely of granitic and volcanoclastic detritus, which is rich in feldspars, ferromagnesian silicate minerals, and volcanic rock fragments. This material is not in chemical

equilibrium with the circulating ground water. Because of the high porosity and permeability of the sediments and the instability of the contained minerals, these deposits, in a relatively short time geologically, have undergone extensive postdepositional alterations that have greatly modified the characteristics of the sediments. Studies of samples collected from many localities throughout the region indicate that the deposits have been affected by at least four types of postdepositional processes: (1) mechanical infiltration of detrital clays into the interstitial pore spaces (Fig. 1), (2) dissolution of detrital framework silicate minerals (Figs. 2, 5), (3) replacement of detrital framework silicate minerals by clay (Figs. 3, 5); and (4) precipitation of a wide variety of stable and metastable authigenic cementing minerals (Figs. 6 through 11). These processes have not been equally important everywhere, but almost everywhere the effects of one or more of them have significantly altered the original texture and mineral composition of the deposits. We describe here the characteristics of the alterations that have resulted from these four processes as they are revealed by a combination of analytical techniques, including thin-section petrography, scanning electron microscopy, x-ray diffraction analysis, energy dispersive x-ray analysis, and microprobe chemical analysis.

The examples illustrated here are from samples that were collected from widespread localities and mainly from deposits of Pliocene(?) and Miocene age. Many of the examples are from outcrops of Pliocene(?) fanglomerates in Cañon Rojo on the west side of the Cocopah Mountains in northeastern Baja California (Walker, 1967, 1976); some are from subsurface cores of alluvial deposits in the Gila Group at Tucson, Arizona; and some are from outcrops of fluvial sandstones and associated eolian deposits in the Sante Fe Group at various localities in the Rio Grande trough in New Mexico. We purposely selected these examples to emphasize that the types of alterations described here are commonplace in deposits of Cenozoic age throughout the desert regions of the southwestern United States and northwestern Mexico, both in outcrop and in the subsurface. Other related aspects of some of these deposits have been discussed in previous papers (Walker, 1967, 1976; Walker and others, 1967).

## MECHANICAL INFILTRATION OF CLAY

Most coarse-grained desert alluvium (sand, sandy gravel, and gravely sand) is essentially free of interstitial clay when originally deposited, because sorting by stream flow, even in ephemeral streams, is a highly efficient process for separating clay from coarser grained detritus. We have examined the texture of modern alluvium on the surfaces of active alluvial fans, in arroyos, and in the channels of perennial streams at hundreds of locations throughout the region, and, as expected, the coarse-grained sediments characteristically are clay free. By analogy, we conclude that most of the Cenozoic alluvium also was clay free when deposited. Both Holocene and Cenozoic alluvium, of course, may be interbed-

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ded with clay-rich mudflows, particularly if source areas are clay rich, but such deposits were not the subject of our investigations.

The earliest diagenetic process that has significantly altered the initially clay free alluvium is postdepositional infiltration of clay into the interstitial voids of the sediments. Because of the arid climate, water tables in this region characteristically are low, particularly in the alluvial fans on the margins of the basins. Hence, whenever there is runoff from the adjacent highlands, water is recharged into the deposits. The pore spaces in the alluvium are many times larger than the clay particles that are suspended in the runoff, and therefore the clay is not filtered from the water as it seeps into the sediments. The infiltrated clay remains in suspension and moves with the water until interstitial conditions are encountered that allow it to settle interstitially and form secondary matrix, in the manner described by Walker (1976).

The mechanism by which clay is infiltrated into sand and the fabrics formed by the clay have been studied in laboratory experiments that simulate natural conditions of influent seepage (Crone, 1975). These studies show that infiltrated clay has a diagnostic clastic texture by which it can be readily distinguished from clay of other origins when samples are examined under a scanning electron microscope. The infiltrated clay typically is composed of mixed sizes of clay platelets that, except in meniscal bridges between

grains (Fig. 1A), are aligned parallel to the surface of framework grains. Thus, they form oriented coatings or "skins" around the grains (Figs. 1A-1C) that are similar to the soil cut illustrated by Brewer (1964, Fig. 43).

Alluvium that contains framework grains that are coated with infiltrated clay having the above characteristics is almost ubiquitous in basin-fill deposits throughout southwestern United States and northwestern Mexico. Although the clay does not occur in freshly deposited sediments, initial stages of its accumulation can be observed in most places within a few feet below the surface pits dug into Holocene alluvium. More advanced stages of accumulation can be observed almost any place that arroyos or man-made excavations have exposed analogous older deposits. A typical example, showing a fairly advanced stage of accumulation, is illustrated in Figures 1D-1F. Such clay occurs throughout porous permeable deposits, but it generally tends to be concentrated above beds that have low permeability and therefore have served as barriers to downward migration of the influent surface water. The clay is particularly abundant in deposits located at the proximal end of alluvial fans, where the sediments are coarsest grained and recharge is most frequent. In such places, infiltrated clay may fill or nearly fill the interstitial voids. As would be expected, the mineralogy of infiltrated clay varies from place to place because it is composed

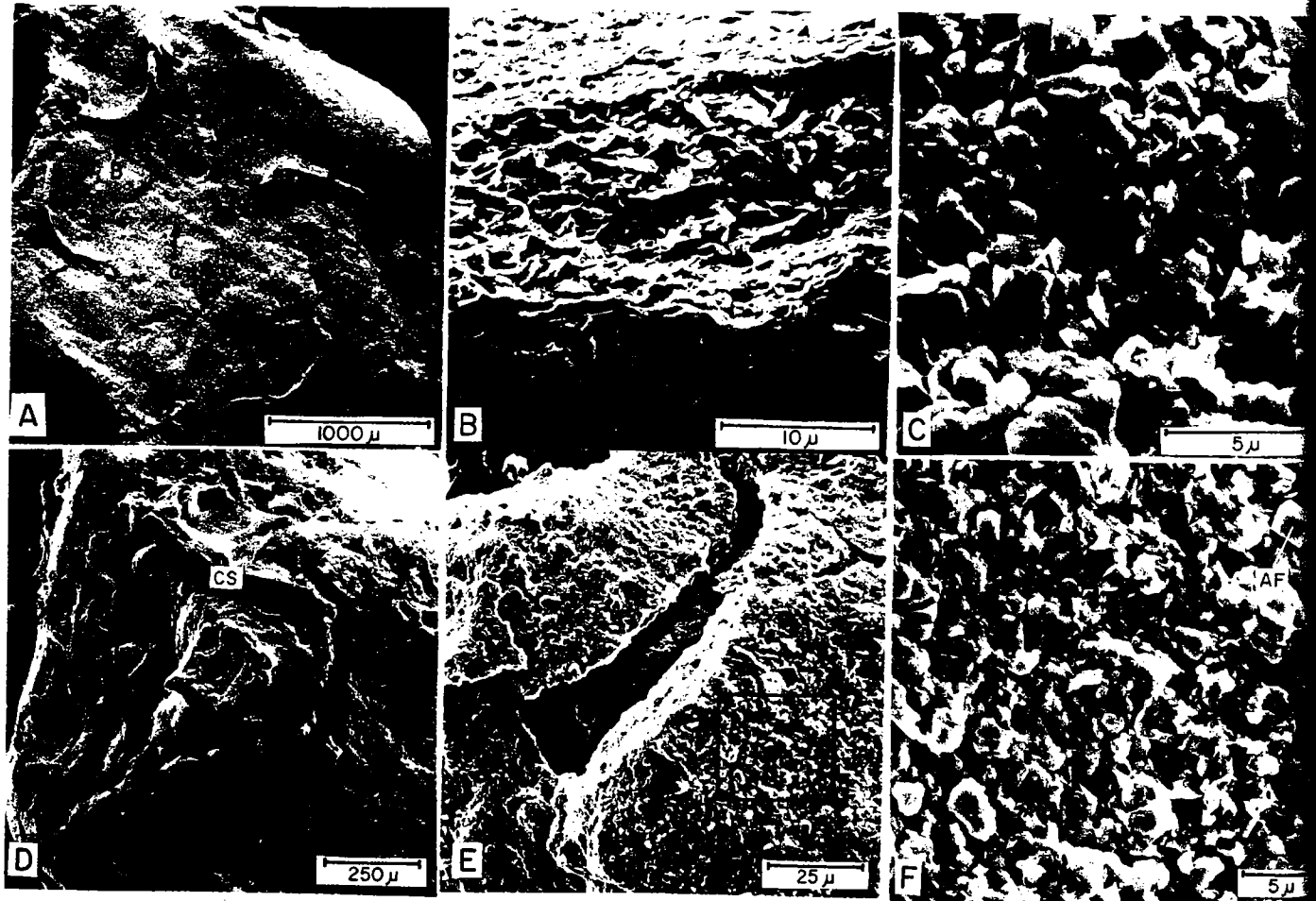


Figure 1. Scanning electron microscope photomicrographs of mechanically infiltrated clay. A-C: artificially infiltrated; D-F: naturally infiltrated. Clay-coated framework grain. Locations of B and C are shown by arrows. B. Enlarged view of clay platelets oriented perpendicular to grain surface. C. Enlarged view of clay platelets oriented parallel to grain surface in clay skin at location C in A. D. Framework grain coated with clay skin (CS) composed of naturally infiltrated clay. Sample SF-G-A-56, upper Gila Group (Pliocene-Pleistocene), from roadcut 6.4 km south of Red Rock, New Mexico. E. Enlargement of area outlined in D, showing broken clay skin. F. Enlargement of area outlined in E, showing clay platelets oriented parallel to grain surface, and one crystal of authigenic potassium feldspar (AF). Note similarity between artificially infiltrated clay in C and naturally infiltrated clay in F.

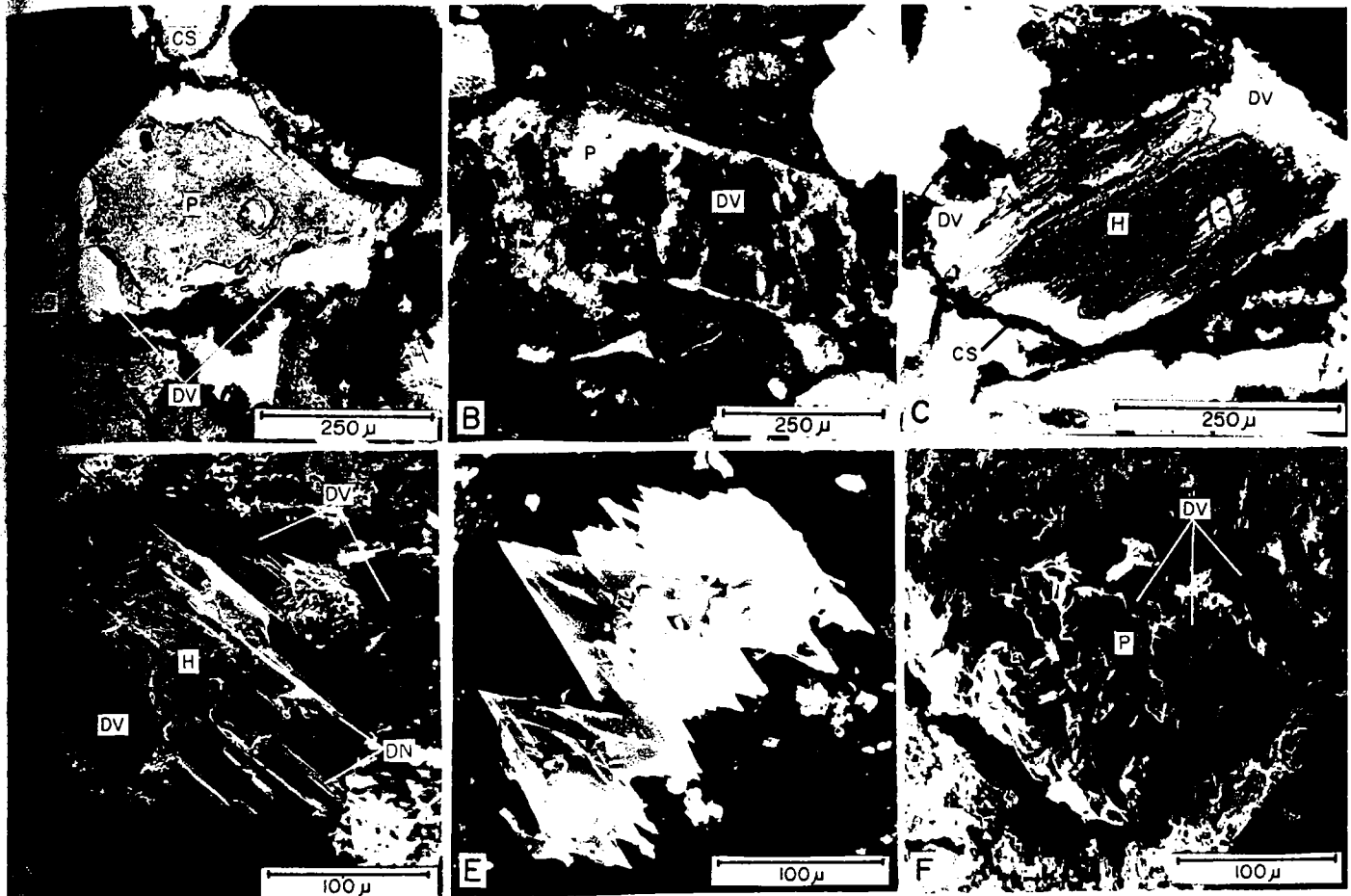


Figure 2. Partially dissolved silicate minerals. A-C: thin-section photomicrographs, D-F: SEM photomicrographs. A. Peripherally dissolved plagioclase grain. Dissolution voids (DV) occupy area between relict grain (P) and clay skin (CS) that marks original outline of grain. Sample PE-478, Pliocene(?) conglomerate, Cañon Rojo, Baja California. B. Internally dissolved plagioclase (P). Most of original area of grain is now occupied by dissolution voids (DV). Sample WFRB-5-2, Pliocene(?) fanglomerate, Cañon Rojo, Baja California. C. Peripherally dissolved hornblende, showing dissolution voids (DV) between relict grain (H) and clay skin (CS) that marks original boundary of grain. Note needle-like terminations on relict grain. Sample WFRB-5, Pliocene(?) fanglomerate, Cañon Rojo, Baja California. D. Partially dissolved hornblende grain (H), showing well-developed dissolution needles (DN) and dissolution voids (DV). Outlined area shows location of Figure 9F. Sample WFRB-5, Pliocene(?) fanglomerate, Cañon Rojo, Baja California. E. Partially dissolved augite grain, showing characteristic cockscomb terminations on relict grain. Sample BC-23, Hayner Ranch Formation (Miocene), San Diego Mountain, Las Cruces area, New Mexico. F. Partially dissolved plagioclase grain (P), showing well-developed dissolution voids (DV). Sample W-1a, Tropic Formation (Miocene-Pliocene), near Dixon, New Mexico.

a mixture of whatever clay minerals are available in the source areas.

Presumably, clay is mechanically infiltrated into porous, permeable sediments wherever influent seepage accompanies alluviation, and large amounts of clay are commonly added to sediments in this way. Indeed, sediments containing between 10% and 20% matrix clay of this origin are common among the older Pleistocene and Tertiary deposits in the region. The mechanism therefore has important effects on the texture, mineralogy, and even the bulk chemical composition of the original sediments. For example, infiltration of clay in this manner changes the original texture by adding important amounts of matrix to sediments that were essentially matrix free when deposited. The process thereby decreases the textural maturity (Folk, 1951) of sediments diagenetically, and it may lead to the erroneous conclusion that the sediments were deposited by mudflows or other currents that were incapable of the sorting action that accompanies normal stream flow. In addition, it changes the mineralogy by adding clay minerals that were not present in the original sediment. Finally, it changes the bulk chemical composition by adding minerals (clays) that normally are richer in aluminum and lower in alkalis and alkaline earths than those of the

original sediment, thus causing corresponding modifications in the bulk chemical composition.

#### ALTERATION OF DETRITAL SILICATE MINERALS

The framework silicate minerals in these deposits have been extensively altered intrastratally. The most altered minerals are those that occur low in Goldich's stability series (1938), such as plagioclase, augite, and hornblende. These minerals have been so pervasively altered and the alterations occur over such a widespread region that it seems certain that given enough time and assuming that the processes continue, some or all of these minerals will be completely removed from the sediments.

The detrital silicates have been altered mainly by two diagenetic processes, dissolution and replacement by clay.

#### Dissolution

Etched and hollow grains resulting from dissolution of detrital minerals are common (Fig. 2). Extensively dissolved grains may be seen with a binocular microscope, but they are more obvious in

petrographic thin section (Figs. 2A–2C), and their details are best observed using a scanning electron microscope (SEM) (Figs. 2D–2F). It should be emphasized that evidence of dissolution is easily overlooked when samples are examined only in thin section (for example, Figs. 2A, 2B), because the dissolution voids are easily misinterpreted as being formed by “plucking” during the process of making the thin section.

SEM examinations show that hornblende grains characteristically develop delicate needle-like terminations (Fig. 2D), which reflect preferential dissolution along lattice planes parallel to the *c* axis. Augite, on the other hand, characteristically develops “cockscorn” terminations, which reflect preferential dissolution along lattice planes transverse to the *c* axis (Fig. 2E). Plagioclase dissolution is more irregular, and although cockscorn structures may develop, the grains usually show irregularly etched and (or) fluted surface textures (Fig. 2F). The characteristics of the relict grains indicate that dissolution may either proceed inward from the periphery of the grains (for example, Fig. 2A) or, particularly in the case of plagioclase, it may selectively remove the interior of the grains (Fig. 2B). In these deposits, plagioclase, hornblende, and augite show evidence of extensive dissolution, but potassium feldspar has not been discernibly affected. Other writers (Miller and Folk, 1955; Heald and Larese, 1973; Flesch and Wilson, 1974), however, have reported evidence of dissolution of potassium feldspar in much older rocks — some as old as early Paleozoic. This evidence suggests that dissolution is a potentially important process of intraratal alteration of all types of silicate minerals wherever they are bathed by interstitial waters with which they are not in equilibrium, regardless of the age of the deposits.

Grains that are completely removed by dissolution may leave no clue that they were original components of the sediment other than the voids that are left in the spaces that the grains once occupied. Commonly, however, the outlines of dissolved grains have coatings of either infiltrated clay minerals, such as those discussed above, or chemically precipitated clay. Both types of clay coatings normally are chemically more stable in the interstitial water than many of the silicate mineral grains that they coat. Hence, upon dissolution of the framework grains, the clay coatings form hollow shells, which are molds of the dissolved grains (Figs. 2A, 2C). Compaction of the sediment, of course, may cause the clay shells to collapse; when that happens, the clay coatings may become indeterminate components of the interstitial matrix. The voids resulting from dissolution of grains may at any stage become filled with cement; if it seals off the permeability, this cement may halt further removal of the mineral grain, unless a change in the chemistry of the interstitial water creates new conditions under which the cement is soluble. For example, Figure 11A displays a hornblende grain that has been partly dissolved, as shown by its needle-like terminations. Subsequently, the void space between the clay skin and the relict of the grain has been filled with calcite cement. Similarly, Figure 11B shows calcite filling a void left by the complete dissolution of a grain of an unknown mineral (perhaps an especially unstable grain such as olivine). In this instance, the only evidence that remains of the dissolved grain is the preserved clay coating that marks its original outline. Without the formation of clay coatings and their subsequent preservation in a more or less intact condition, no evidence of completely dissolved grains would remain in the sediments.

The dissolution of unstable silicate minerals has several important effects on the original characteristics of the sediments. For example, it increases the porosity of the sediments, assuming of course that (1) the dissolution voids have not been subsequently eliminated by compaction or filled with cement, and (2) that the formation of voids has not been compensated by the loss of interstitial porosity owing to the reprecipitation of the dissolved ions as stable authigenic minerals. Inasmuch as dissolution removes

feldspars in preference to quartz, the process increases the quartz to feldspar ratio and thereby increases the mineralogical maturity of the sediments diagenetically. In addition, dissolution changes the texture of the sediments by decreasing the grain size of the affected minerals (Fig. 2A) and by producing relicts of silt and other fine-grained minerals of grain sizes that were not present in the original sediment (Fig. 2B). Finally, dissolution may significantly change the bulk chemistry of the deposits owing to the removal of ions by migrating groundwater, assuming that all of the released ions have not been precipitated as cementing minerals.

### Replacement by Clay

Another commonly observed evidence of diagenetic alteration is the occurrence of silicate grains that have been replaced *in situ* by clay (Fig. 3). The clay replaces grains irregularly (Fig. 3A), preferentially along crystallographic planes (Fig. 3C), incipient fractures, grain contacts in rock fragments (Fig. 3D), or any other zone of weakness in the grains. The contacts between the daughter clay and relicts of the parent minerals normally are very abrupt; that is, there seems to be no transition zone of partial replacement at the margin of the relicts (Fig. 3E). Moreover, the contacts typically are blunt (Fig. 3E); cockscorn and needle-like terminations such as those that typify dissolution of augite and hornblende are not developed by replacement.

Using identification criteria developed by Reynolds and Howarth (1970), x-ray analyses show that in basin-fill deposits throughout the southwestern United States and northwestern Mexico, the diagenetically formed clay characteristically is randomly interstratified mixed-layer illite-montmorillonite containing between 80% and 95% expandable layers (Fig. 3I).

Replacement has affected the texture and mineralogy of the deposits in several important ways. For example, it has greatly increased the amount of interstitial matrix, and hence it has increased the textural maturity of the sediments diagenetically. Modal analyses of thin sections show that in places, clay of replacement origin commonly exceeds 15% of the volume of the sediments. Such sediments commonly have a graywacke-like texture; that is, they are matrix rich, and the nonclay fraction consists of a poorly sorted mixture of angular and subangular grains (Fig. 3F). Indeed, these deposits in places provide outstanding examples of sediments that show early stages in the diagenetic development of graywacke-type textures in the manner proposed by Cummins (1962). The daughter clay rarely occurs as intact pseudomorphs of the parent grains (exceptions are shown in Fig. 3F), because compaction resulting from overburden pressure causes the soft clay to become squeezed between more rigid framework grains (Figs. 3G, 3C). Thus, the daughter clay ultimately becomes interstitial matrix. That the matrix clay forms by replacement of framework grains becomes increasingly obscured with time, with degree of alteration and particularly with depth of burial.

Replacement also has demonstrably reduced the average grain size of the nonclay fraction of the sediments, owing to the common breakdown of framework grains into smaller sizes by incomplete replacement. The replacement of plagioclase grains has been particularly important in causing this type of textural change because plagioclase makes up a major percentage of the sediments, and it has undergone extensive alteration (Figs. 3A, 3C). Similarly, replacement by clay has been important in the disaggregation of rock fragments. Granitic rock fragments are the most extensively disaggregated owing mainly to the selective replacement of the most unstable mineral components (Figs. 3G, 3H). In places, even such stable rock types as quartzite have been partly disaggregated because of replacement along contacts in the composite grains (Fig. 3D).

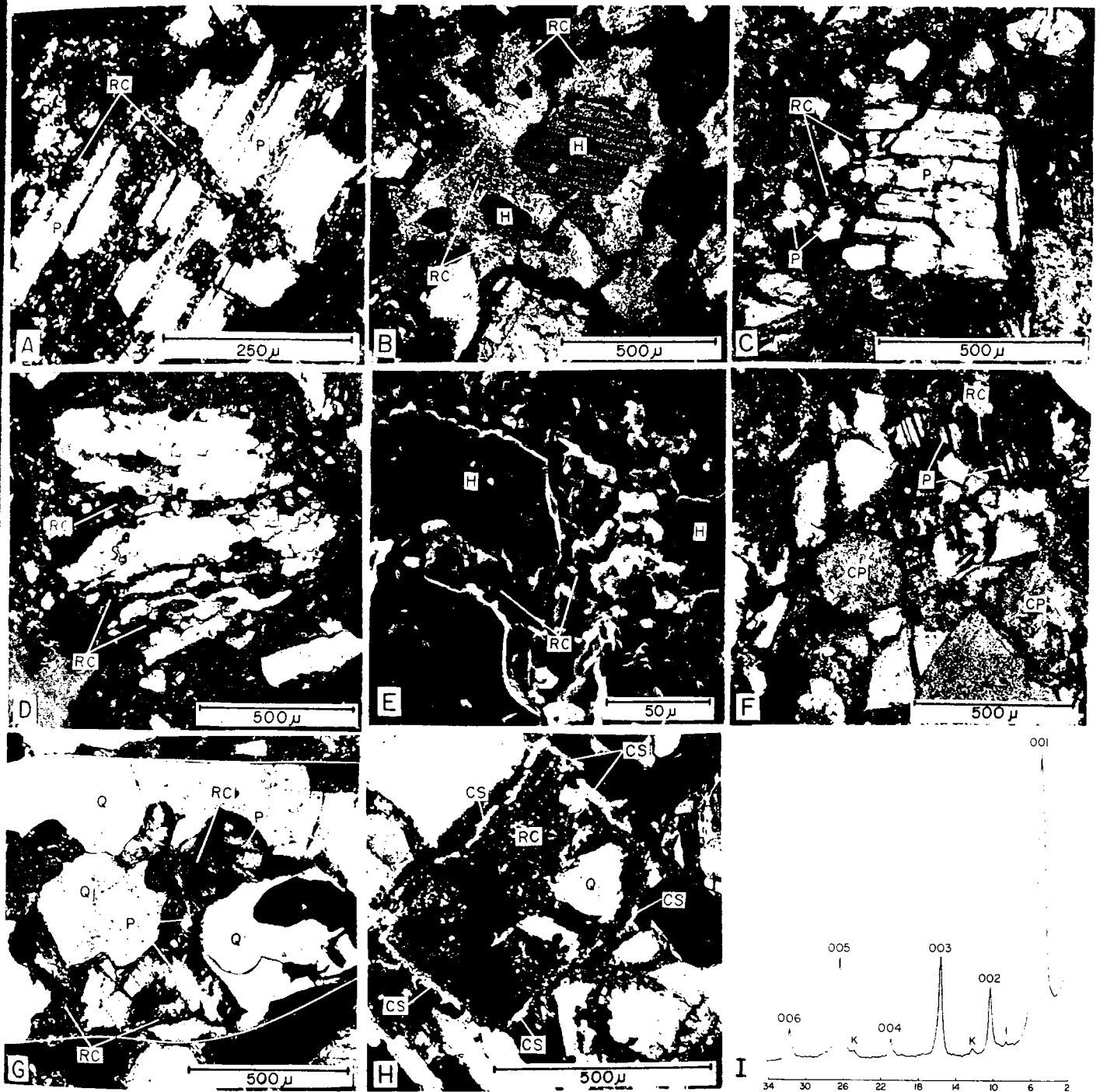


Figure 3. Photomicrographs (A-D and F-H are thin section, E is SEM) of silicate minerals replaced by clay. A. Plagioclase grain (P) that has been replaced irregularly by clay (RC). Sample BE-32S, Pliocene(?) fanglomerate, Cañon Rojo, Baja California. B. Hornblende grain (H) that has been peripherally replaced by clay (RC), which in turn has become squeezed between other framework grains. Sample V-CR-M-1-2, Pliocene(?) fanglomerate, Cañon Rojo, Baja California. C. Plagioclase grain (P) that has been preferentially replaced by clay (RC) along cleavage planes. Note development of silt-sized relicts of plagioclase. Core sample GS-5-423, from Gila Group at depth of 138 m, Tucson, Arizona. D. Quartzite rock fragment, showing partial disaggregation due to replacement by clay (RC) along grain contacts. Core sample GS-5-228, from Gila Group at depth of 74 m, Tucson, Arizona. E. Hornblende (H) replaced by clay (RC), showing characteristic abrupt and blunt contacts between parent hornblende and daughter clay. Sample WFRB-1, Pliocene(?) fanglomerate, Cañon Rojo, Baja California. F. Clay pseudomorphs (CP) of grains that probably originally were hornblende, extensively replaced plagioclase grain (P), and many other grains showing partial replacement. Sample BE-32, Pliocene(?) fanglomerate, Cañon Rojo, Baja California. G. Part of granitic rock fragment (border marked by white line) in which plagioclase grains (P) have been selectively replaced by clay (RC), and quartz (Q) has remained essentially unaltered. Sample BB-2, Tesuque Formation (Miocene-Pliocene), near Truchas, New Mexico. H. Granitic rock fragment now composed mainly of clay (RC), which has replaced all original minerals except quartz (Q). Original outline of rock fragment is marked by an oriented clay skin (CS). Same sample as in G. I. X-ray diffraction trace of  $-2 \mu\text{m}$  fraction of sample illustrated in C, solvated with ethylene glycol,  $\text{CuK}\alpha$  radiation.

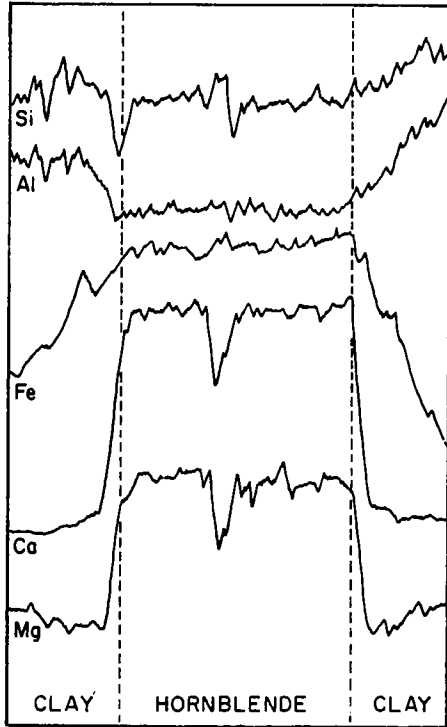


Figure 4. Five-element microprobe chemical analysis along traverse across hornblende grain that has been peripherally altered to clay, as in Figure 3B. Core sample GS-2-204, from Gila Group at depth of 67 m, Tucson, Arizona.

In some instances where compaction has not distorted the replaced grains, it is possible from thin-section examination to distinguish between the parallel-oriented clay in the clay coating, which was emplaced by earlier mechanical infiltration, and the randomly oriented clay, which was formed by subsequent replacement of silicate minerals (Fig. 3H). These distinguishing characteristics, however, probably cannot survive intense compaction.

Replacement has significantly changed the mineral composition of the deposits because the more unstable framework minerals, such as plagioclase and ferromagnesian silicates, have been extensively removed, and the more stable minerals, such as quartz and potassium feldspar, have been proportionally increased. Thus, replacement has increased the mineralogical maturity of the sediments diagenetically, in the same way as dissolution. Replacement also has changed the mineralogy by adding mixed-layer illite-montmorillonite, which was not originally present in the deposits.

Finally, replacement has changed the chemical composition of the sediments because ions are released in the conversion of parent silicates to daughter clay, and some of these ions are carried off in solution in the ground waters. The major chemical changes that occur in the conversion of hornblende to clay are indicated in Figure 4, which shows a five-element microprobe traverse across a hornblende grain that has been altered peripherally to clay in a manner analogous to the alteration of the grain shown in Figure 3B. The analysis shows that as hornblende alters to clay, there is a loss of magnesium, calcium, and iron and a corresponding relative increase in aluminum and silica (see also Walker and others, 1967). We have not made comparable microprobe analyses of replaced

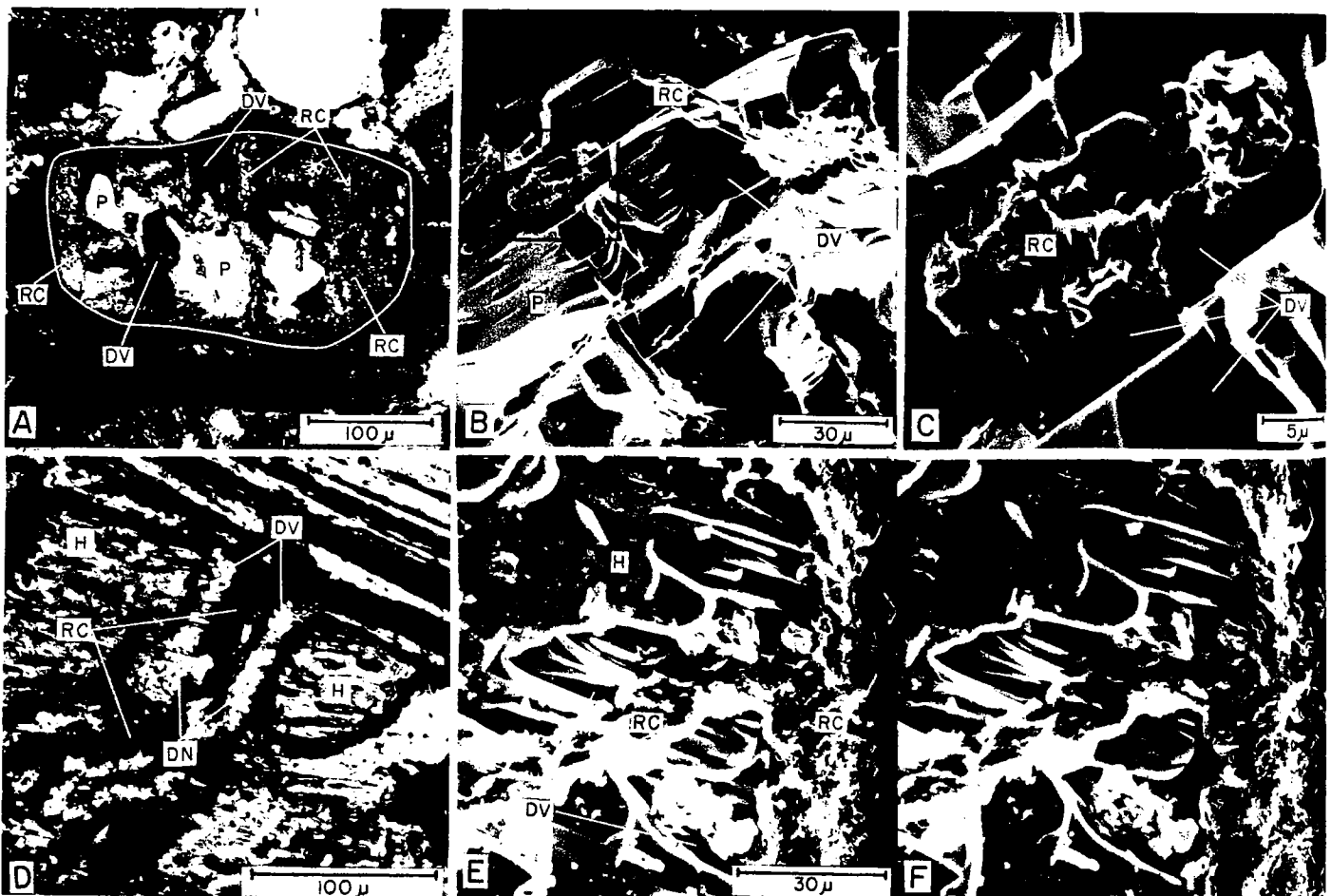


Figure 5. Partially replaced and partially dissolved silicate minerals. A. Thin-section photomicrograph of plagioclase (P) that has been partially replaced by clay (RC) and subsequently dissolved, forming dissolution voids (DV). Sample WFRB-1, Pliocene(?) fanglomerate, Cañon Rojo, Baja California. B. Same alteration sequence viewed under SEM. C. Enlargement of part of B. D. Thin-section photomicrograph showing effects of similar replacement-dissolution sequence in hornblende (H) and development of dissolution voids (DV) and dissolution needles (DN). Sample PE-424, Pliocene (?) fanglomerate, Cañon Rojo, Baja California. E-F. Same alteration sequence viewed in SEM stereopair.

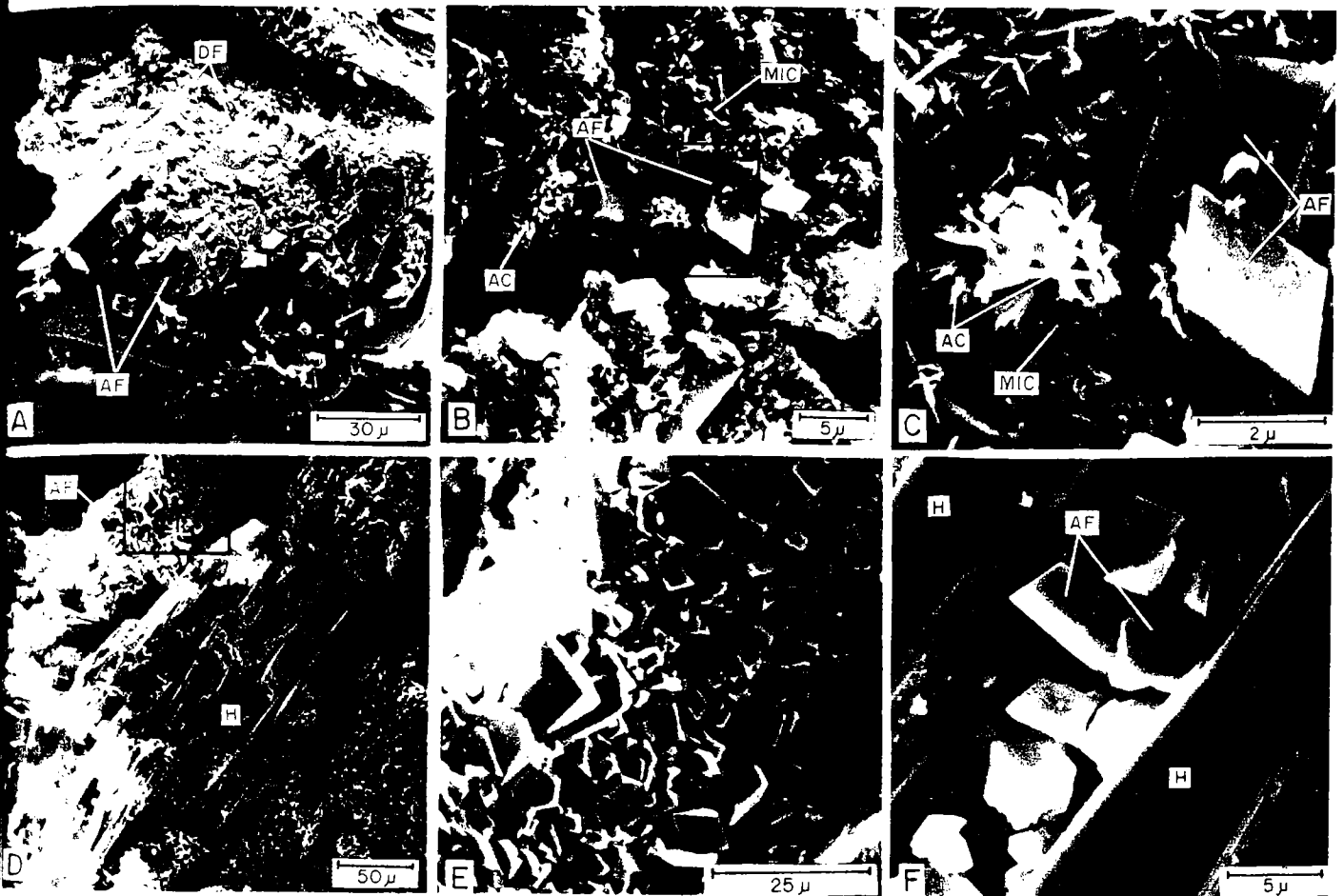


Figure 6. SEM photomicrographs of authigenic potassium feldspar. A. Authigenic potassium feldspar overgrowth (AF) on detrital feldspar grain (DF). Sample VCR-M-1, Pliocene(?) fanglomerate, Cañon Rojo, Baja California. B. Bright-red interstitial matrix consisting of mixture of mechanically infiltrated clay (MIC), authigenic clay of uncertain mineralogy (AC), and authigenic potassium feldspar (AF). Sample PE-248, Pliocene(?) fanglomerate, Cañon Rojo, Baja California. C. Enlargement of outlined area in B. D. Relict of partly dissolved hornblende grain (H) surrounded by reddish "matrix" composed predominantly of authigenic potassium feldspar (AF). Sample BC-27B, Hayner Ranch Formation (Miocene), San Diego Mountain, near Las Cruces, New Mexico. E. Enlargement of upper outlined area in D, showing close-up of authigenic feldspar "matrix." F. Enlargement of lower outlined area in D, showing authigenic potassium feldspar (AF) partly filling dissolution voids in relict hornblende grain (H).

plagioclase grains, but presumably those alterations also involve a loss of some of the elements, particularly calcium and sodium. Depending upon the chemistry of the interstitial ground waters, the released ions may be removed in solution or precipitated as authigenic minerals. To the extent that these ions have been removed in solution, the chemical composition of the altered sediments has been changed from what it was at the time of deposition.

### Multiple Alterations

In some places, the framework silicate minerals show evidence of having undergone both replacement and dissolution, in that order. Evidence of such multiple stages of alteration of plagioclase and hornblende, as they are revealed in thin section and under the SEM, is illustrated in Figure 5. The relationships shown in these photomicrographs support the following interpretation. Initially, both the plagioclase and hornblende were replaced by clay on the periphery of the grains and along fractures, cleavages, and other zones of weakness within the grains. At this stage of alteration, the grains probably resembled those shown in Figure 3A, 3B, and 3E, with the daughter clay occurring in close contact with the relicts of the partly replaced parent minerals. Subsequently, however, replacement halted and dissolution began. The previously formed clay, being a product of intrastatal alteration, remained stable, but as the parent silicate minerals began to dissolve, the surfaces of the relicts of the parent silicate mineral grains receded away from the earlier formed clay, leaving voids such as those shown in Figure 2.

The reason for the change in type of alteration from replacement to dissolution is uncertain, but it almost certainly reflects a change in the chemistry of the interstitial water. However, the nature of the change and the time at which it occurred cannot be ascertained from available data.

Presumably, dissolution can also precede replacement, but we have seen no evidence of this sequence of alteration. It may be very difficult to preserve such evidence, because replacement would tend to destroy characteristics that are diagnostic of earlier dissolution, such as the cockscomb terminations on augite and the needle-like terminations on hornblende. We conclude that both alteration sequences are possible, but we can convincingly document only one of them.

### DESTINY OF THE RELEASED ELEMENTS

Dissolution and replacement of silicate minerals result in the release of some or all of the chemical elements contained in the altered grains. Obviously, greater percentages of the contained elements are released if the grains are dissolved than if they are replaced by clay. Nevertheless, even replacement, when it is pervasive and widespread, is accompanied by the release of important amounts of material. The destiny of the released elements depends on the physicochemical conditions of the interstitial water that is in contact with the altering minerals. The elements may stay in solution as soluble ions and migrate with the ground water, or they may precipitate as authigenic minerals that are stable under the condi-



tions existing in the interstitial environment. The most commonly formed authigenic minerals in these deposits are potassium feldspar (Fig. 6), zeolites (Fig. 7), clay minerals (Fig. 8), quartz (Fig. 9), hematite or its precursor oxides (Fig. 10), and calcite (Fig. 11). All of these minerals are common constituents of the interstitial matrix that typically binds together the framework grains in the basin-fill deposits. Indeed, the so-called "matrix" in these deposits seems typically to be composed either entirely of one or more of these authigenic minerals or, alternatively, of a mixture of one or more of these minerals superimposed on platelets of clay, which have been mechanically infiltrated into the deposits some time after deposition. For most of the sandstones and conglomerates in these deposits, the interstitial fine-grained matrix seems rarely to be an original detrital component.

The authigenic minerals in these geologically young sedimentary deposits are so extremely fine grained that except where they form overgrowths on framework grains (Fig. 6A), they normally are very difficult or even impossible to recognize without SEM examination. We have found that the nature of the interstitial matrix is best studied using an SEM for determination of crystal morphology, an attached energy dispersive x-ray analyzer for determination of the elemental composition of the material examined, and a standard x-ray diffractometer for determination of the mineralogy of fractional concentrates.

#### Authigenic Minerals

**Potassium Feldspar.** This is one of the most common authigenic minerals. Along with zeolites and clay minerals, it accounts for most of the reprecipitated aluminum and potassium and much of the reprecipitated silica. Authigenic potassium feldspar occurs in part as overgrowths on detrital feldspar grains (Fig. 6A); therefore, it commonly is coarse enough to be recognized in thin section. Detrital grains of both potassium feldspar and plagioclase serve as host grains for the overgrowths. Sheppard and Gude (1969, p. 21) have illustrated analogous overgrowths on plagioclase in the Barstow Formation (Miocene) in southern California. Authigenic potassium feldspar also occurs as pseudorhombic crystals (Figs. 6B, 6C) that typically are so small that only the larger ones are discernible in thin section, and even these are easily overlooked. Crystals of potassium feldspar in many places are important components of the interstitial matrix. They usually are associated with other matrix material, such as mechanically infiltrated and authigenic clay (Figs. 1F, 6B, 6C) and authigenic quartz (Fig. 9E). In some places, however, the interstitial matrix is composed almost entirely of aggregates of potassium-feldspar crystals (Figs. 6D, 6E). In places, the crystals have partly filled the voids formed by dissolution of framework grains (Figs. 6D, 6F), indicating that partial dissolution

of the detrital silicate minerals has preceded precipitation of the authigenic potassium feldspar.

Energy dispersive x-ray analyses of several representative crystals indicate that they are composed of potassium, aluminum, and silica. X-ray diffraction traces of samples containing concentrates of the crystals confirm that the mineral is potassium feldspar.

Other investigators (Hay, 1966; Sheppard and Gude, 1968, 1969, 1973) have shown that potassium feldspar is a common authigenic mineral in other deposits of Cenozoic age, particularly in tuffaceous lake sediments, in many places in the southwestern United States.

**Zeolite.** Authigenic zeolite, which x-ray analysis indicates is *clinoptilolite* (a high-silica zeolite that usually contains sodium), is a common cementing mineral in areas where the basin-fill deposits contain volcanoclastic detritus. Locally it forms the major cement. The formation of zeolites seems to be fostered by the alteration of volcanic glass, but in the appropriate chemical environment, they probably can form from the alteration of any type of aluminosilicate mineral (Hay, 1966, p. 83). The clinoptilolite in these deposits typically forms tiny clear euhedral crystals with low birefringence and high relief. These crystals line interstitial voids and, if abundant, they can be recognized in thin section (Fig. 7A); however, they are best seen under the SEM (Figs. 7B, 7C). Wherever it occurs, clinoptilolite accounts for the precipitation of significant amounts of aluminum and silica. To our knowledge, it and montmorillonite are the only authigenic minerals in these deposits that are likely to contain significant amounts of sodium. The occurrence of authigenic zeolites in tuffaceous deposits of Cenozoic age in the desert regions of the western United States has long been known, and their nature and genesis have been discussed in detail by Hay (1963, 1966), Sheppard (1971), and Sheppard and Gude (1968, 1969, 1973).

**Clay Minerals.** Much of the released aluminum and silica and probably some of the sodium also are reprecipitated as clay (Fig. 8). Throughout the region, the most common precipitated clay is randomly interstratified mixed-layer illite-montmorillonite containing 80% to 95% expandable layers. This clay yields x-ray traces that are similar to those obtained from the clay of replacement origin described above (compare Fig. 3I). Thus the precipitated clay is nearly pure montmorillonite. In rare instances, it forms radiating crystals that line interstitial voids; in such instances, its authigenic origin is obvious in thin section (Fig. 8A). In most instances, however, when examined in thin section the precipitated clay is an unrecognizable part of the interstitial matrix, and its authigenic origin cannot be ascertained by its optical properties. However, under the SEM, it has a diagnostic boxwork texture that confirms its authigenic origin (Figs. 8B, 8C). The texture is similar to the texture of montmorillonite illustrated by Borst and Keller (1969, Fig. 33).

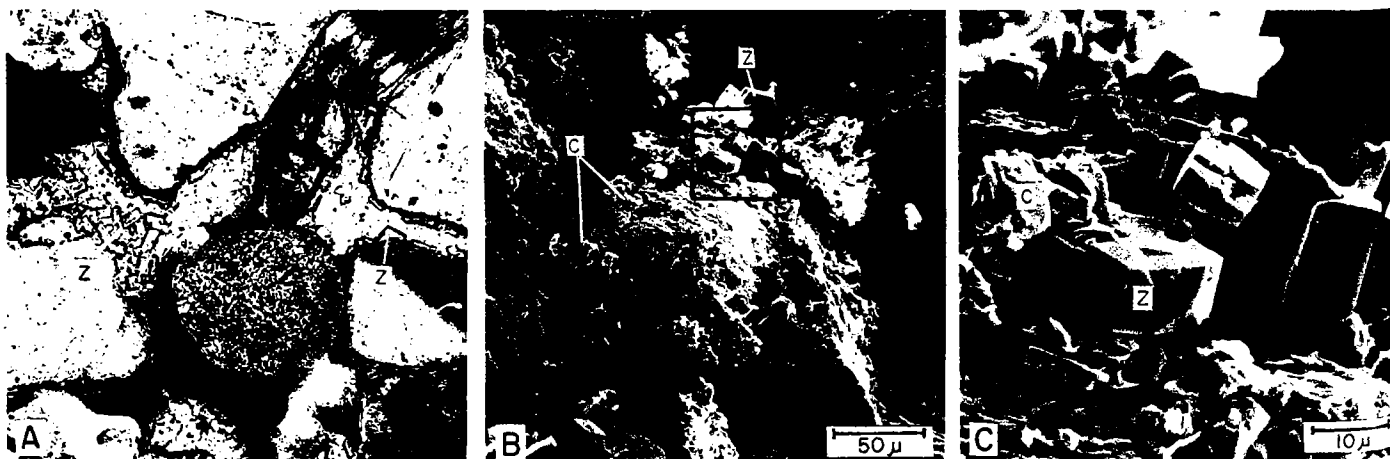


Figure 7. Authigenic zeolite. A. Thin-section photomicrograph showing authigenic clinoptilolite (Z) that partly fills interstitial voids. Sample Penr. 46, Tesuque Formation (Miocene-Pliocene) near Dixon, New Mexico. B. SEM photomicrograph of same sample, showing authigenic clinoptilolite (Z) and associated clay. C. Enlargement of outlined area in B.



and it is strikingly different from the texture of mechanically infiltrated clay (compare Figs. 8 and 1).

Our SEM studies show that chemically precipitated clay is an important contributor to the interstitial matrix in these deposits in many places. In some places, clay that is solely of this origin has nearly filled the interstitial voids (Figs. 8D-8F, 8H, 8I). In some in-

stances, the clay forms curved sheet-like masses that appear to have originated as grain coatings. For some reason, perhaps repeated wetting and drying, the grain coatings have become crumpled, and they have separated from the surfaces of the framework grains (Figs. 8D-8F); the interstitial voids have become nearly filled with contorted platy fragments of authigenic clay (Fig. 8E) that perhaps

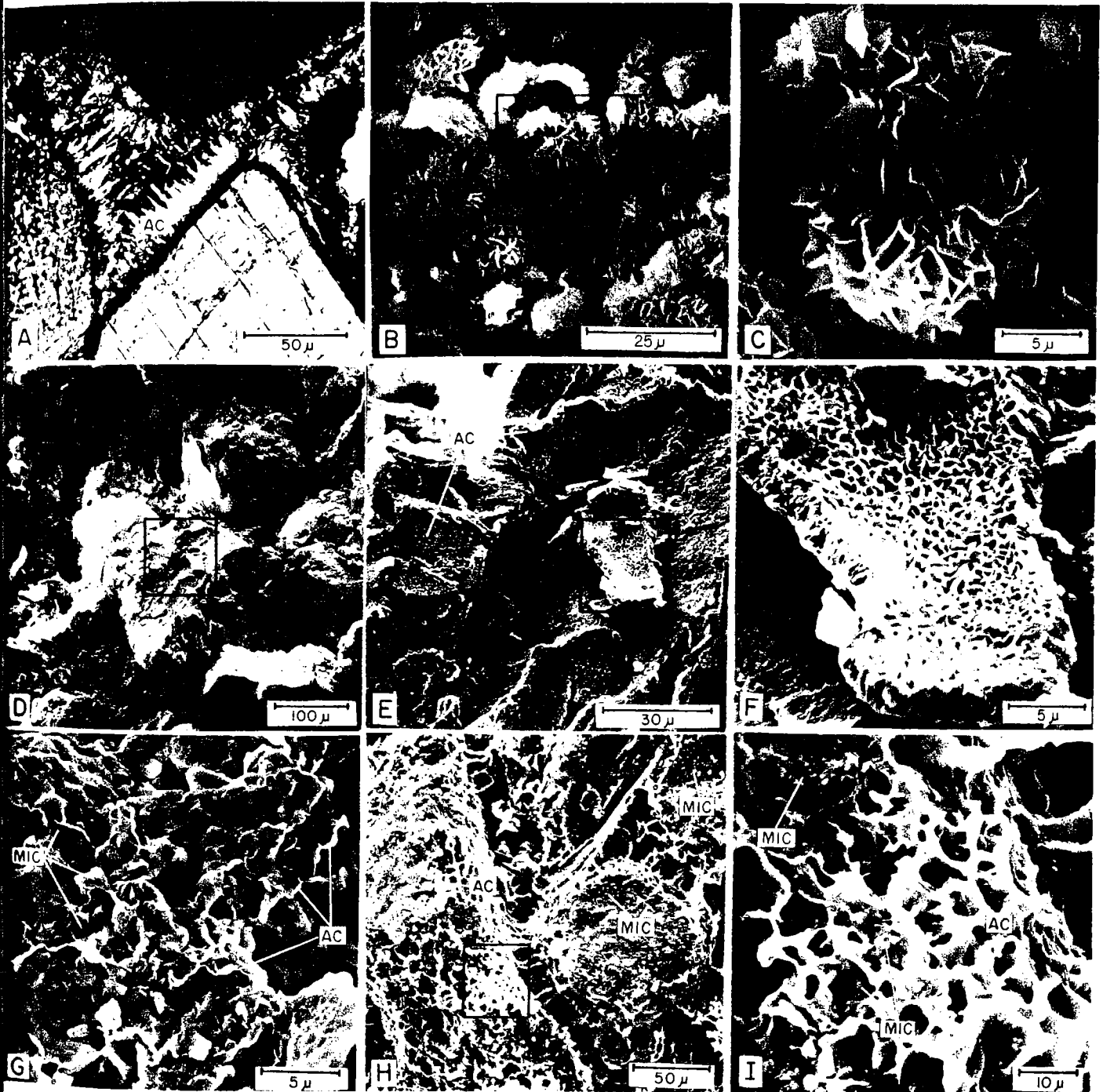


Figure 8. Thin-section (A) and SEM photomicrographs (B-I) of authigenic clay. A. Authigenic clay (AC) occurring as coarse-grained void liner. Sample VLE-285, Cuchara Formation (Eocene), SW $\frac{1}{4}$ sec. 36, T. 31 S., R. 68 W., Las Animas County, Colorado. B. Authigenic clay forming interstitial matrix in core sample GS-4-701.5, Gila Group, Tucson, Arizona. C. Enlarged view of outlined area in B. D. Authigenic clay forming matrix in well-sorted (colian?) sand associated with alluvium. Sample T-31, Ojo Caliente Member, Tesuque Formation (Miocene-Pliocene), in roadcut on U.S. Highway 64, 1.9 km south of junction with New Mexico Highway 75 near Dixon, New Mexico. E. Enlarged view of outlined area in D, showing crumpled, sheet-like character of the clay. F. Enlarged view of outlined area in E, showing boxwork texture of the clay. G. Incipient stage of development of authigenic clay (AC) on previously deposited mechanically infiltrated clay (MIC). Sample SF+G-A-56, Gila Formation near Red Rock, New Mexico. H. Advanced stage of development of authigenic mixed-layer clay (AC) on mechanically infiltrated clay (MIC). Sample BC-23a, Hayner Ranch Formation Miocene, San Diego Mountain, near Las Cruces, New Mexico. I. Enlarged view of outlined area in H, showing boxwork texture of the clay.

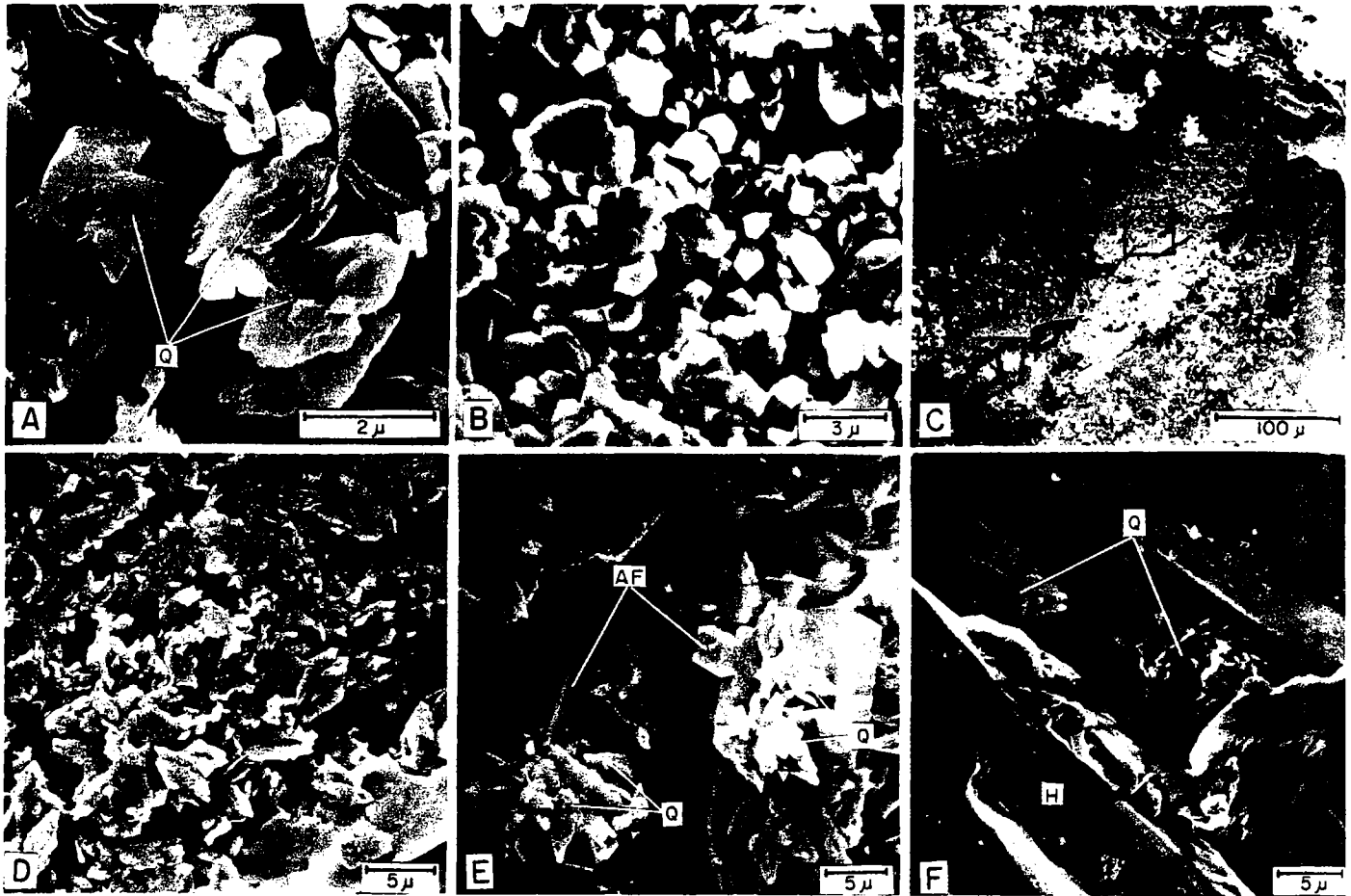


Figure 9. SEM photomicrographs of authigenic quartz. A. Tabular crystals of quartz which are flattened parallel to basal pinacoid and have unequally developed prism faces. Concentrate of red silt fraction from sample WFRB-1, Pliocene(?) fanglomerate, Cañon Rojo, Baja California. B. Prismatic crystals of quartz which are terminated with rhombohedral faces. Bright-red matrix in El Rito Formation (Eocene?), 7.4 km northwest of El Rito, New Mexico, on New Mexico Highway 110. C. Bright-red interstitial matrix containing abundant authigenic quartz and authigenic potassium feldspar. Sample WFRB-5, Pliocene(?) fanglomerate, Cañon Rojo, Baja California. D. Enlarged view of upper outlined area in C, showing interstitial matrix composed mainly of authigenic quartz (for example, at arrows) of type shown in A. E. Enlarged view of lower outlined area in C, showing authigenic quartz (Q) encrusting authigenic potassium feldspar (AF). F. Enlarged view of outlined area in D in Figure 2, showing authigenic quartz of type shown in A here; quartz partly fills dissolution voids in relict hornblende grain. Sample WFRB-5, Pliocene(?) fanglomerate, Cañon Rojo, Baja California.

represent multiple generations of formation. In other instances, the interstices contain aggregates of curled-up plates that may be of similar origin (Figs. 8B, 8C). Many of the samples we have examined contain authigenic clay that has been precipitated on surfaces coated with clay platelets, which we interpret to have been mechanically infiltrated. An example showing an initial stage in the formation of authigenic clay having this association is shown in Figure 8G; an advanced stage of formation of similar clay that has almost concealed the underlying mechanically infiltrated clay is shown in Figures 8H and 8I.

We emphasize that occurrences such as these illustrate that the mineralogy of the matrix clay in these and analogous ancient fluvial sandstones and conglomerates is unreliable evidence of source-area climate. For example, the matrix in these deposits contains both authigenic and mechanically infiltrated clay. The authigenic clay obviously reflects the interstitial chemical environment, not the weathering environment in the source area. It is true that the mechanically infiltrated clay is derived from the source area, but even this clay, assuming that it can be differentiated from authigenic clay, does not necessarily reflect the weathering conditions, because the clay commonly is reworked from shale and other clay-rich bedrock formations rather than from soils that have formed by *in situ* weathering. Furthermore, if the source areas are arid to semiarid, such as those of concern here, much if not most of even the clay in the soils probably is not a product of *in situ* weath-

ering; rather, it is probably airborne clay that has been reworked by wind from many different sources, which may lie far outside the watershed. The importance of airborne dust in desert soils has been stressed by Yaalon (1973). Accordingly, we believe that the mineralogy of the matrix clay in these and other analogous ancient deposits is of dubious value in making climatic interpretations.

**Quartz.** Most of the free silica is precipitated interstitially as two crystal forms of quartz: (1) tabular crystals that are flattened parallel to the basal pinacoid and have unequally developed prism faces that give them a pseudo-orthorhombic form (Fig. 9A), and (2) prismatic crystals that are terminated with rhombohedral faces (Fig. 9B). That both types of crystals are composed of  $\text{SiO}_2$  (commonly with impurities of iron) has been confirmed by energy dispersive x-ray analysis of several representative crystals; both types consist of quartz, as confirmed by x-ray diffraction analysis of concentrates of the crystals.

Both types of quartz crystals are at least locally important constituents of the interstitial matrix, and each type dominates the matrix in places (Figs. 9B-9D). We have not observed the two crystal forms occurring together, and therefore we conclude that one is not the precursor of the other. We also conclude that the two forms reflect either different concentrations of  $\text{SiO}_2$  in the ground waters, or other differences in the interstitial chemical conditions, but we do not know the nature of the differences. Both quartz forms commonly are associated with authigenic potassium feldspar (Fig. 9E).

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We have only rarely observed authigenic overgrowths on detrital quartz grains.

As in the case of authigenic potassium feldspar, crystals of authigenic quartz have partly filled the voids that are formed by dissolution of framework grains (Figs. 2D, 9F). This indicates that dissolution of the detrital silicate minerals has preceded the precipitation of authigenic quartz.

**Hematite and Its Precursor Oxides.** Inasmuch as iron is a common constituent of some of the most unstable detrital minerals, such as augite and hornblende, the alterations in these deposits have led to the release of important amounts of iron. The behavior of the iron depends upon the interstitial chemical conditions that control the state of the iron. If conditions favor reduction, the iron is carried as ferrous ions, which remain in solution and migrate with the ground water. If conditions favor oxidation, the iron precipitates as hematite or as a limonitic precursor oxide, which should convert to hematite upon aging, in the manner described by Berner (1969) and Langmuir (1971). Conditions generally have favored the precipitation of these iron oxides; the basin-fill sediments throughout the region characteristically are stained reddish yellow to bright red, and they commonly contain identifiable hematite crystals (Fig. 10).

Hematite occurs as drusy linings of voids that are left by the dissolution of some of the framework minerals, presumably ferromagnesian silicates (Figs. 10A–10C), and as tabular concentrates

occurring between cleavage laminae in biotite grains (Figs. 10D–10F). More commonly, however, iron oxide occurs as stain, which is either amorphous or occurs in crystals that are too fine grained to be detected under an SEM. These oxides, which commonly are bright red, stain all types of framework and matrix grains, including the previously described mechanically infiltrated clay (Figs. 1D–1F, 6B, 6C), authigenic feldspars (Fig. 6E), authigenic clays (Figs. 5D–5F, 8H, 8I), and quartz (Fig. 9). Throughout most of the region, the precipitation of iron oxide in these various ways has converted the deposits into red beds (Walker, 1976).

**Calcite.** Authigenic calcite, although not present in all samples, is common, and it is the mineral that most commonly causes induration of the deposits. (All of the authigenic minerals at least locally cause induration.) The calcite generally is coarsely crystalline (Fig. 11), and it fills both interstitial voids and the voids that are formed by dissolution of framework silicate minerals (Figs. 11A, 11B). If dissolution voids are filled by calcite prior to compaction, and if the original outlines of the dissolved grains are marked by coatings of clay and (or) iron oxides, the calcite preserves the outlines of the dissolved grains (Figs. 11A, 11B). If calcite seals off the permeability, it protects relicts of partially altered minerals from further alteration (Fig. 11A). The protection may persist until a subsequent change in the chemistry of the interstitial water causes later dissolution of the calcite and re-exposes the mineral to the interstitial waters. An example of authigenic calcite that has been partly re-

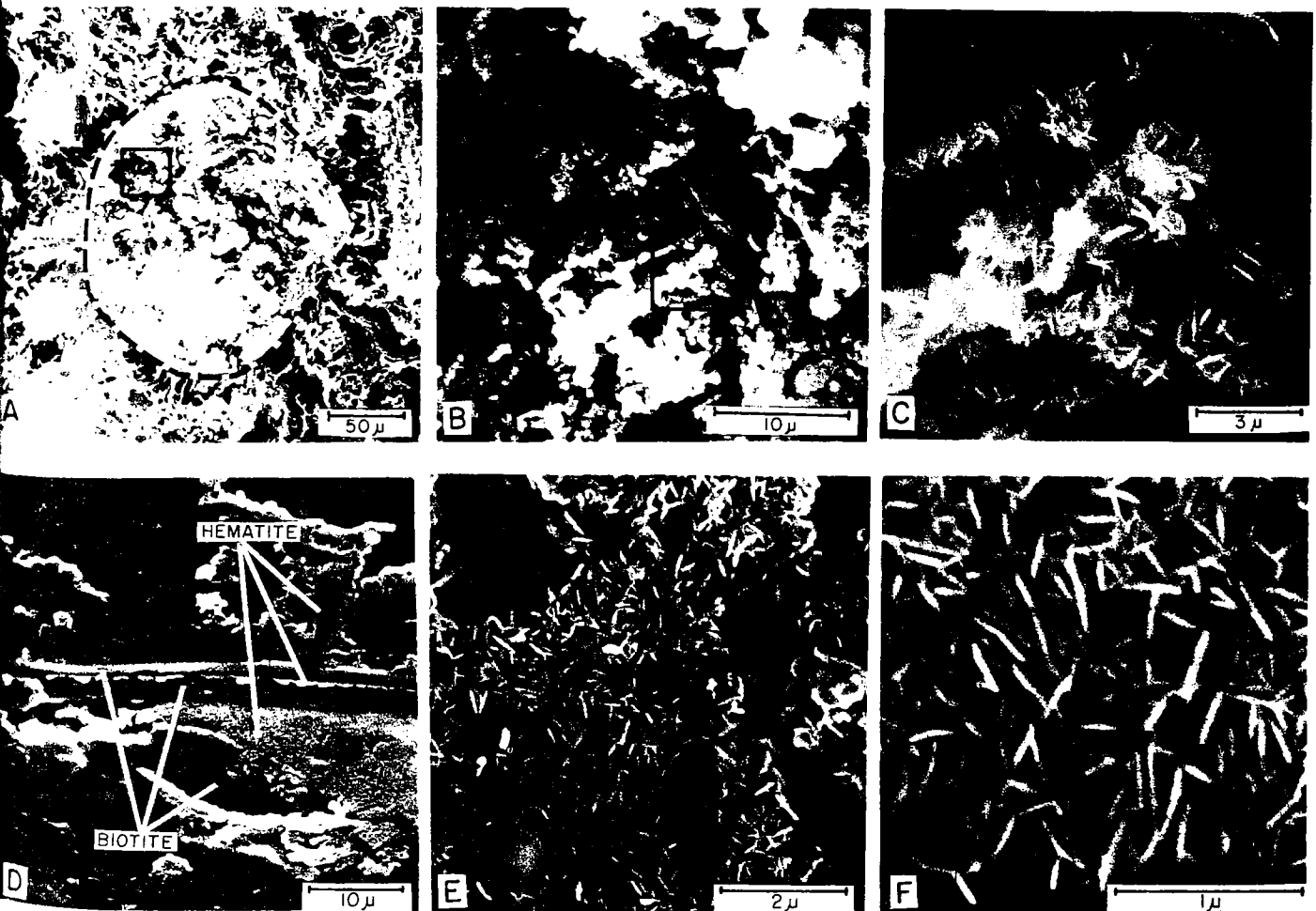


Figure 10. SEM photomicrographs of authigenic hematite. A. Hematite crystals lining void (encircled area) left by dissolution of framework grain, probably ferromagnesian silicate mineral. Sample BC-23a, Hayner Ranch Formation (Miocene), San Diego Mountain, near Las Cruces, New Mexico. B. Enlarged view of outlined area in A. C. Enlarged view of outlined area in B, showing clusters of euhedral hematite crystals. D. Intercleavage hematite in biotite. E. Enlarged view of part of D. F. Enlarged view of part of E. Sample L-27, Tertiary red fanglomerate (Miocene?), Laguna Mountains, near Yuma, Arizona. Crystals in both examples are hematite, as indicated by red color, hexagonal shape, and iron-oxide composition as confirmed by energy dispersive x-ray analysis.

moved by dissolution is shown in Figures 11C–11F. Evidence displayed in these photos supports the following interpretation. Prior to precipitation of the calcite, the interstices of this sediment were partly filled with a mixture of mechanically infiltrated clay and authigenic feldspar (Figs. 11E, 11F), in a manner analogous to the interstitial mixture illustrated in Figure 6B. Subsequently, coarsely crystalline calcite was precipitated, and it engulfed the clay and feldspar mixture and nearly filled the remaining void space (Fig. 11C). Still later, in response to what must have been a change in the chemistry of the interstitial water, the calcite has been partly dissolved, and the included clay and feldspar are now exposed on the etched surfaces of the calcite (Figs. 11E, 11F). Presumably, continued dissolution of calcite in this manner would ultimately allow protected relicts of unstable silicate minerals, such as the hornblende relict shown in Figure 11A, to be bathed again by ground water with which the minerals are not in equilibrium, and the destruction of these minerals would resume. Such episodic removal of unstable minerals probably is common during the history of many ancient first-cycle deposits. The example illustrates the complexity of postdepositional alterations in first-cycle deposits.

### SUMMARY AND CONCLUSIONS

The alterations described here, which presumably are still in progress in the Cenozoic deposits in the southwestern United States

and northwestern Mexico, provide excellent examples of mineral responding to changes in their environment. The deposits are composed predominantly of minerals derived from crystalline rocks; therefore, the minerals were formed under physical and chemical conditions that were strikingly different from the conditions under which they have existed since they were deposited in the sedimentary basins. Most of the minerals are not in equilibrium with their surroundings; hence, the unstable minerals are being destroyed and new minerals that are stable under the new conditions are being formed. Presumably, these mineralogical changes will continue until an equilibrium mineral assemblage has formed.

The degree to which partly altered minerals remain in these deposits, some of which are tens of millions of years old, indicate that the formation of an equilibrium mineral assemblage is a slow process, even on a geologic time scale. Indeed, on a geologic time scale, equilibrium probably is only a temporary condition, because each subsequent change in the interstitial environment tends to upset the equilibrium and initiate new alterations. The physicochemical characteristics of the interstitial water can change repeatedly during the long history of a deposit, and each change produces a new set of conditions which in turn leads toward the formation of a new equilibrium mineral assemblage. Intrastrata alterations probably are most rapid during the early history of first-cycle deposits, because at this time the sediments are likely to contain the most unstable minerals. However, alterations can occur

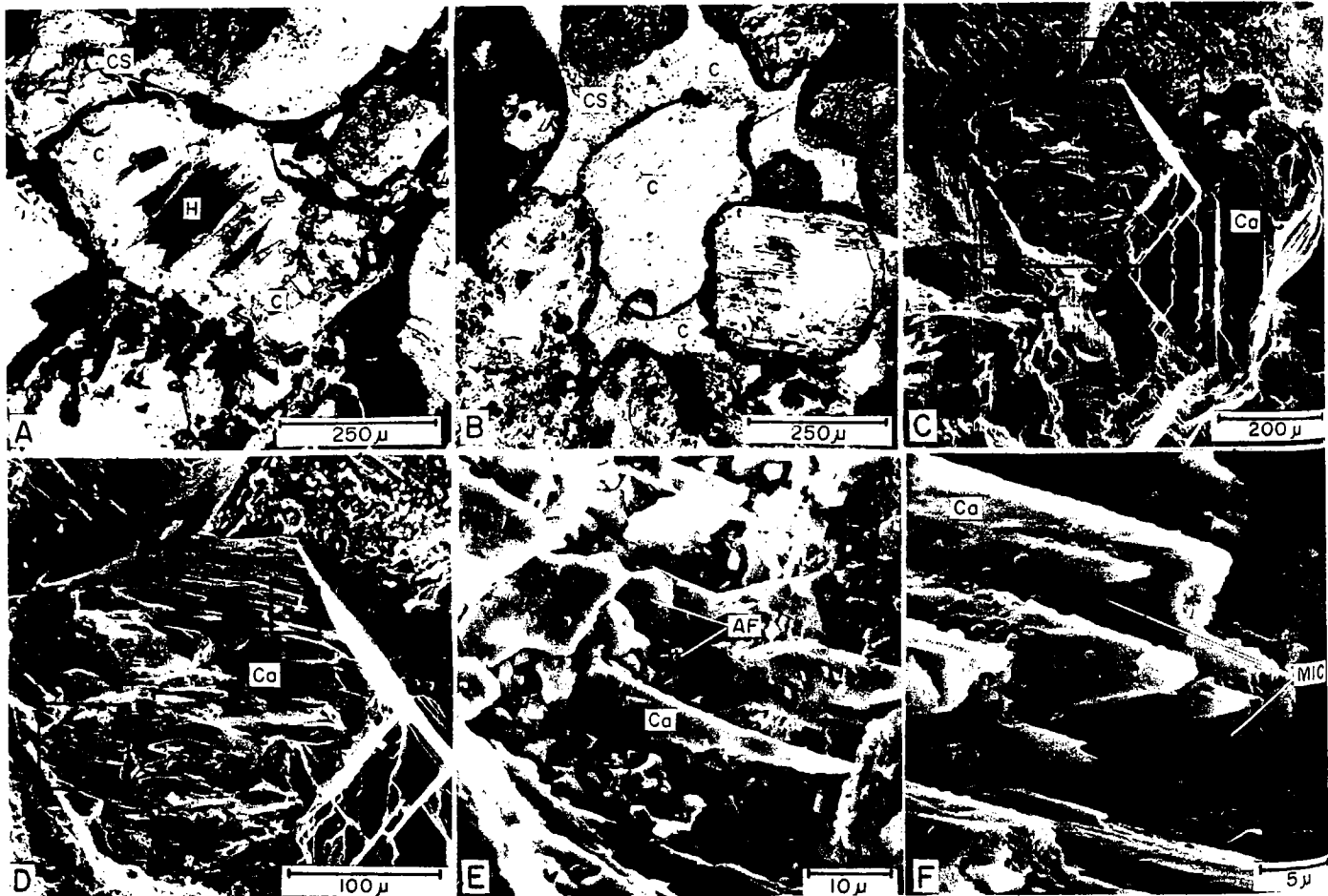


Figure 11. Thin-section (A and B) and SEM (C–F) photomicrographs of authigenic calcite. A. Partially dissolved hornblende grain (H), with original outline marked by preserved clay skin (CS). Dissolution void is filled with calcite (C). Sample WFRB-3-7, Pliocene(?) fanglomerate, Cañon Rojo, Baja California. B. Probable completely dissolved silicate grain, with original outline marked by preserved clay skin (CS). Dissolution void is filled with calcite (C), which is optically continuous with interstitial calcite. Sample BC-27b, Hayner Ranch Formation (Miocene), San Diego Mountain, near Las Cruces, New Mexico. C. Calcite cement (Ca) in Pliocene(?) fanglomerate, Cañon Rojo, Baja California. Freshly broken surfaces that follow cleavage planes are products of sample preparation. Sample PA-38-1. D. Enlarged view of outlined area in C, showing naturally etched surface of calcite. E. Enlarged view of lower outlined area in D, showing inclusions of authigenic potassium feldspar (AF) in calcite (Ca) which have been exposed by the etching. F. Enlarged view of upper outlined area in D, showing exposed inclusions of mechanically infiltrated clay (MIC).

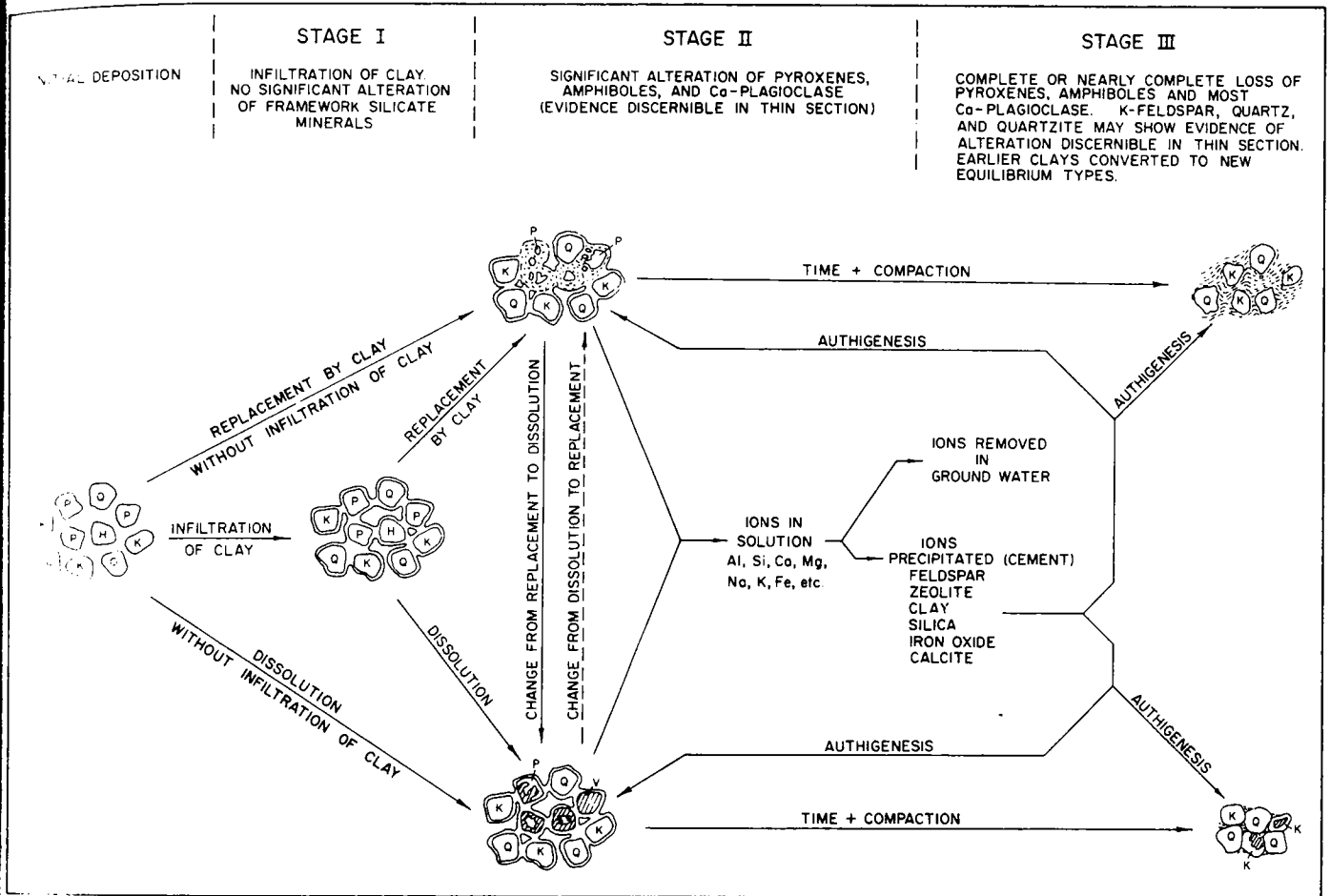


Figure 12. Diagrammatic interpretation of observed and inferred stages of diagenetic alterations in first-cycle desert alluvium.

any time the minerals are not in equilibrium with the interstitial waters. Moreover, none of the common rock-forming minerals is immune to alteration; even such a relatively stable mineral as quartz is altered in some ground waters (Fig. 3D).

Changes in the interstitial environment may have both internal and external causes. For example, the chemistry of the interstitial water is significantly affected by the ions that are released by intrastatal alterations. As unstable minerals are removed, important sources of ions are eliminated, and the composition, pH, and so forth of the interstitial water change correspondingly. In addition, chemical changes are affected by such external causes as climate and tectonism. Climate affects the amount and temperature of water that circulates through the sediments and the amount of organic matter included in the original sediments, and these factors in turn affect rates of intrastatal alteration, concentration of dissolved ions, pH, and redox potential. Tectonism affects the rate of ground-water circulation, interstitial temperature, and pressure. For example, uplift tends to increase the circulation rate and decrease the temperature and pressure. Conversely, deep burial tends to decrease the circulation rate and increase the temperature and pressure. All of these factors affect mineral stability, and changes in any of them are likely to promote mineralogical changes in the sediments.

We conclude that the history of intrastatal alteration can be exceedingly complex, the alterations can occur at any time during the history of the sediments, and the alteration processes probably are never truly terminated.

Figure 12 summarizes diagrammatically the major types of post-depositional alterations that have occurred in the sands, sandstones, and conglomerates of Cenozoic age in the desert regions of the southwestern United States and northwestern Mexico (stages I and II). It includes our interpretation of the nature of the

sandstones that eventually will be formed if the same types of alterations continue (stage III). Admittedly, the alterations are more complex than this simplified diagram implies, but the diagram summarizes the major types of changes that can be expected during the history of many ancient nonmarine first-cycle deposits.

When initially deposited, the sands are composed of an essentially clay-free mixture of grains that range from very unstable minerals (represented in the diagram by hornblende) to relatively stable quartz. Feldspars, including both plagioclase and potassium feldspar, are the most abundant minerals; in places they initially compose more than 50% of the framework grains. (Sands that are derived from volcanic source areas also would be very rich in volcanic rock fragments, but, for the sake of simplicity, these have been omitted from the diagram.) Accordingly, when initially deposited, the sands characteristically are texturally mature and mineralogically immature.

Almost immediately following deposition, clay begins to be mechanically infiltrated into the sands wherever influent seepage accompanies alluviation (stage I). The clay becomes concentrated around the framework grains in coatings composed of clay platelets that are oriented parallel to the grain surfaces. This is the earliest diagenetic alteration of the sands. It changes the characteristics of the original sediment because it increases the clay fraction and thereby decreases the textural maturity of the sand. This stage is bypassed if conditions are unfavorable for influent seepage of surface water.

As the sands are bathed by interstitial water, both above and below the water table, the unstable minerals are altered by replacement and (or) dissolution (stage II). The minerals may be altered exclusively by either type of process, or first by one and subsequently by the other. We infer that either process may be followed by the other, but we have seen evidence proving only that

dissolution can follow replacement. Both processes tend to destroy unstable minerals and thereby proportionately increase the percentage of stable grains. Thus, both processes increase the mineralogical maturity of the sediment. Dissolution increases the porosity, but the gain may be offset by subsequent compaction and (or) cementation. Replacement increases the amount of matrix clay, and thus, along with the clay that is mechanically infiltrated, it decreases the textural maturity. Similarly, replacement by clay decreases the porosity and permeability of the deposits.

Both replacement and dissolution release ions to the interstitial waters. Depending upon the chemistry of the water, the ions may remain in solution and migrate with the ground water, or they may precipitate as authigenic minerals. The types of minerals precipitated are controlled by the chemistry of the interstitial water. The precipitation may occur at any time, assuming, of course, that the sources of ions remain available and the chemical conditions are favorable for precipitation.

As these processes continue, minerals that originally were important constituents of the sediments are removed completely or nearly completely (stage III), in the reverse order of their stability. (The relative stability of the minerals, of course, may vary, depending upon the interstitial physicochemical environment.) If the unstable minerals are replaced by clay, the resulting rocks consist of stable types of framework grains, such as quartz and potassium feldspar, set in a clayey matrix which, if abundant, gives the rock a graywacke-like texture. If the unstable minerals are removed mainly by dissolution, the resulting sandstone consists primarily of quartz and potassium feldspar without abundant matrix, and any of the various authigenic minerals may serve as cements.

The authigenic minerals that are formed during the early post-depositional history of first-cycle deposits, such as those described herein, are not likely to remain unchanged throughout the later history of the deposits. For example, if the temperature and the pressure change because of deeper burial and (or) if changes occur in the composition of the interstitial water, the minerals formed earlier may no longer be in equilibrium with their surroundings; therefore, they will alter to a still younger generation of authigenic minerals that are in equilibrium with the new conditions. For example, the mixed-layer clays may convert to clays containing higher percentages of illite layers, in a manner analogous to changes described in Gulf Coast sediments by Perry and Hower (1970). Alternatively, the clays may convert to chlorite or to authigenic feldspars. The zeolites probably convert to feldspars. Some of the first-generation minerals such as potassium feldspar, quartz, calcite, and hematite probably remain stable under a wider range of conditions, and their only change may be recrystallization and (or) the development of overgrowths.

We conclude that diagenetic alterations such as those observed in the Cenozoic deposits are likely to have occurred in many analogous ancient first-cycle fluvial deposits at a comparable time in their history. We further conclude that where there is prolonged movement of ground water through first-cycle deposits, the more unstable silicates such as pyroxene, amphibole, and calcic plagioclase in time can be removed completely or nearly completely and leave no direct evidence that they were important original constituents of the deposits. Accordingly, we conclude that ancient first-cycle nonmarine arkose probably rarely has the same mineralogical, textural, or bulk chemical characteristics that the sediments had when they were deposited. Finally, we conclude that the mineralogy, texture, and chemical composition of ancient nonmarine arkoses and other first-cycle deposits do not necessarily indicate the lithology and climate in the source area or the nature of depositional currents and other environmental factors in the depositional basin.

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