THE EVOLUTION OF THE CANON CITY EMBAYMENT, COLORADO

By

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EMBAYMENT, COLORADO

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CHAPTER I

INTRODUCTION

Statement of Purpose

For many years geologists have investigated the sedimentation, stratigraphy, and structure of the Canon City area in Colorado. However, the geological evolution, especially that of the Mesozoic and Cenozoic eras, and the petrology and diagenesis of its rocks have not been studied in detail, nor have the general subsidence and thermal histories been synthesized.

Furthermore, the origin of oil in the Florence Oil Field--the second oldest west of the Mississippi River in the United States--has not been studied in detail, nor has the productive capacity of this ancient oil field been evaluated. However, the known geological development and thermal maturation of source rocks in the adjacent Denver basin provides a useful parallel in attempting to solve the above two problems.

This study, then, will discuss the following aspects in detail: (1) the geological history and setting of the Canon City area; (2) the petrographic description, diagenesis, depositional environment, and correlation of the sedimentary rocks; (3) the thermal maturation of the Florence Oil

Field's source rocks, the origin of its oil and the possible distribution of undiscovered petroleum. A detailed description of tectonics and paleontology is beyond the scope of this research.

In this work, materials regarding the evolution of the studied area are presented first in a chronological sequence. Next, the subsidence history, thermal maturation, and petroleum geology are discussed.

The terms "Canon City area" or "Canon City embayment" are restricted to the geographical area between Grape Creek, three miles west of the Arkansas River, and Beaver Creek, approximately fifteen miles northeast of Canon City (Figure 1). The area, a topographical and structural lowland, is located at the junction of the Wet Mountains and the Front Range of the Rocky Mountains.

Methods of Investigation

One hundred and twenty sedimentary rock samples were collected from outcrops in the Canon City area, covering formations ranging in age from the Ordovician to the Cretaceous. Of these, 83 were selected for thin section making. Over one hundred articles regarding the studied area were reviewed.

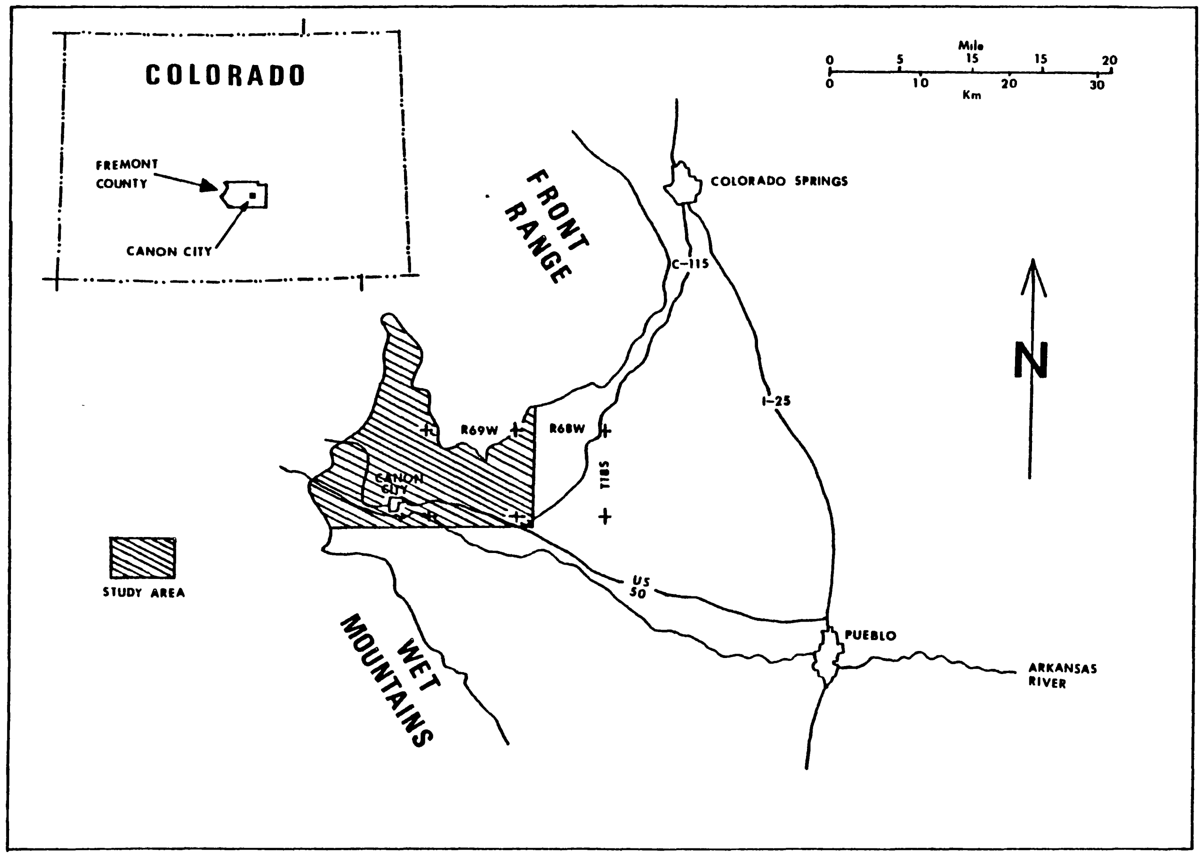


Figure 1. Location of the Canon City Embayment

CHAPTER II

LITERATURE SURVEY

Hayden (1869) first investigated the areal geology of the Canon City area during the United States Geological and Geographical Survey of the then territories of the West. Endlich (1874) first detailed the surface geology, and he prepared a geological map in 1877. According to Gerhard (1967), this map of the Colorado Territory showed the form of the area, but failed to give either the correct ages of the pre-Dakota sedimentary rocks or an acceptable description of tectonic structures.

Walcott (1892), in a detailed stratigraphic study, first described and named the Harding Sandstone and Fremont Limestone. In 1894, Cross mapped the area and named the Manitou Limestone and the Fountain Formation. Fenneman (1905), Frederickson (1956), and Brainered (1957) further described the above formations, and in addition Fenneman (1905) named the Lykins Formation. Frederickson (1956) investigated the formation and the overlying Ralston Creek Formation which was first differentiated by Leroy (1946). Cramer (1962) conducted a comprehensive stratigraphical study of the Ralston Creek Formation both in the Canon City area and along the southern Front Range from Canon City into

El Paso County, Colorado. Emmons and Eldridge (1896) first defined the Morrison Formation while Brady (1969) studied the stratigraphy and petrology of the formation in the embayment area. Meek and Hayden first described the Dakota Group in 1862 from exposures in Dakota County, Nebraska. Subsequently, Morris (1961) studied the Dakota Group along the northeast flank of the Canon City embayment from Phantom Canyon to Colorado Springs. His report covered the lithology, stratigraphy, age, depositional environments and economic aspects of the Dakota Group. Weimer (1970) conducted a regional synthesis on the Dakota Group stratigraphy in the southern Front Range area. He recognized three genetic units within the group and presented correlations and an environmental interpretation.

In addition to the preceeding specific investigation many other geologists have analyzed some parts of this area, including Lee (1915), Lovering and Johnson (1933), Powers (1935), Van Tuyl and Lovering (1935), Wahlstrom (1947), and Lovering and Goddard (1950). Boos and Boos (1957) published generalized maps of the Front Range and discussed the tectonics of the area.

Many geologists have discussed the ancestral Rocky Mountains (e.g., Lee, 1923; Melton, 1925; Ver Wiebe, 1930; Schuchert, 1930; Glockzin and Roy, 1945; Maher, 1953; Curtis, 1958; and Mallory, 1958), but none of their studies have detailed local structures and their relation to those of Laramide age except that of Gerhard (1967).

Washburne (1908) first noticed volcanic rocks in the Canon City area; however, these rocks were not studied in detail until 1967 when Epis and Chapin (1968) presented an article on volcanic rocks at a symposium on volcanism held in Colorado.

Washburne first examined the Canon City coal field in 1910; he discussed the field geology, the stratigraphic positions of the coal beds, and the characteristics of the coal, relating them to economic conditions at that time. Billingsley (1978) interpreted the depositional environments of the sandstones, shales, and the coal beds in the Walsenburg area.

Youmans (1888) first described the Florence Oil Field and assigned the Pierre Shale a Middle Cretaceous age. Washburne (1909) detailed the stratigraphy and the mode of oil occurrence in the field. The following geologists have described parts of the field: Newberry (1888), Fenneman (1905), Day (1908, 1909, and 1912), Foote (1915), Clayton, and Swetland (1980).

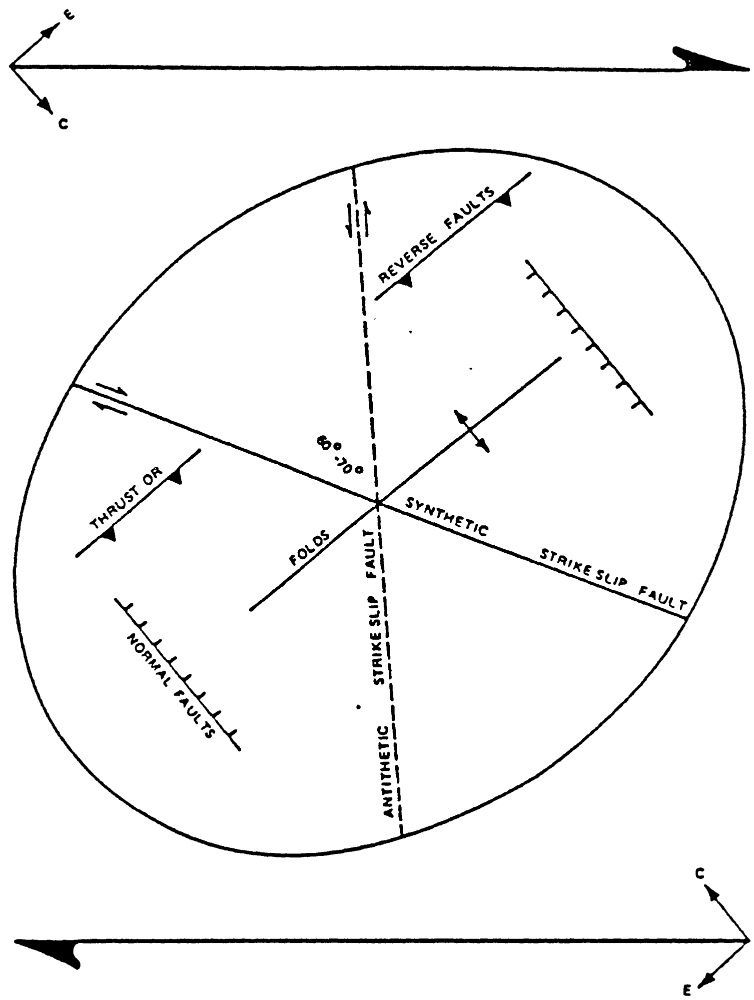
CHAPTER III

PRE-CAMBRIAN

Geological Setting

According to Baars (1980), the framework of the Colorado Plateau was created either by a pure compressional force, or as a sequence of regional right lateral wrench couples during early Pre-Cambrian time (Figure 2). A conjugate set of transtensional rifts was generated with a dominant north-westerly series of major fractures and a subordinate series of north-easterly trending fractures. The northwest-trending faults have a dextral sense of displacement while the northeasterly faults have a sinistral sense. Second-order structures are north-south oriented normal faults in the present sites of the Central and Southern Rockies, and east-west oriented compressional folds in the present site of the Central Rockies.

This basement framework controlled the pattern of sedimentation throughout the Rocky Mountains region during the Paleozoic until Mississippian time; however, the faults were active until the Laramide revolution. In the embayment, faults related to this initial framework include the northwest trending Cooper Mountain fault, the northeast trending Twin Mountain fault, the east-west faults of the



C-Vector of compression
derived from wrenching

E-Vector of extension
derived from wrenching

Figure 2. Wrench Assemblage from a Right-Lateral Wrench Couple (After Harding, 1974)

Royal Gorge fault zone, and the smaller north-south normal faults which bound the Oil Creek graben (a re-entrant of the Canon City embayment) (Figure 3).

In the Late Pre-Cambrian time, a continental arch developed across Colorado on which there was a lowland named the Colorado Sag. The present sites of the Wet Mountains, the Front Range of the Rockies, and the Canon City area were within this sag (Figure 4).

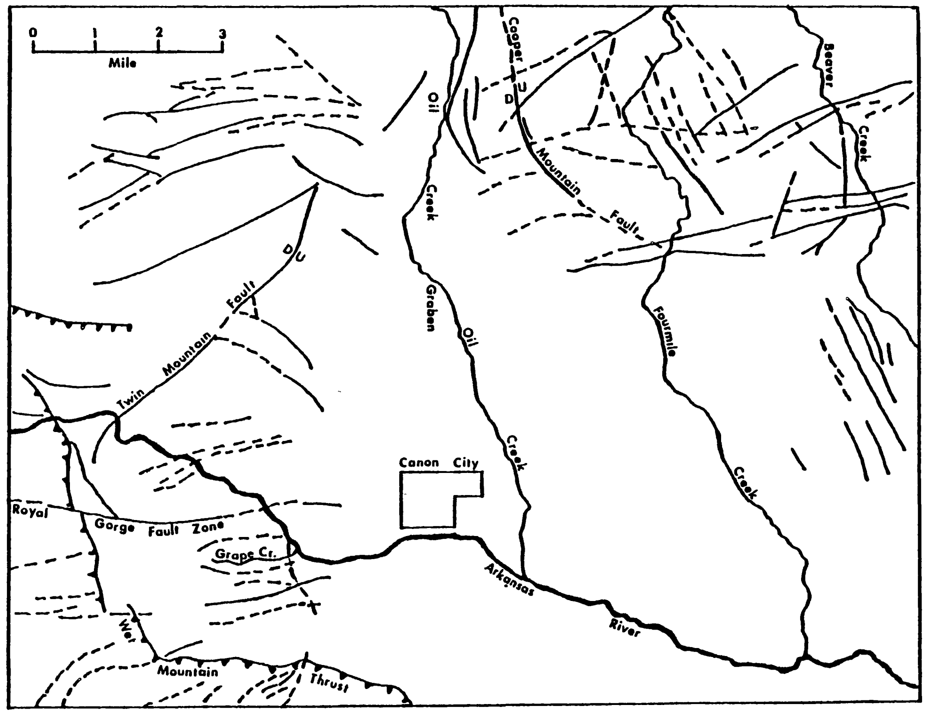


Figure 3. General Structural Map of the Canon City Embayment. Showing Some Geographic Features Mentioned in Text (After Boos and Boos, 1957)

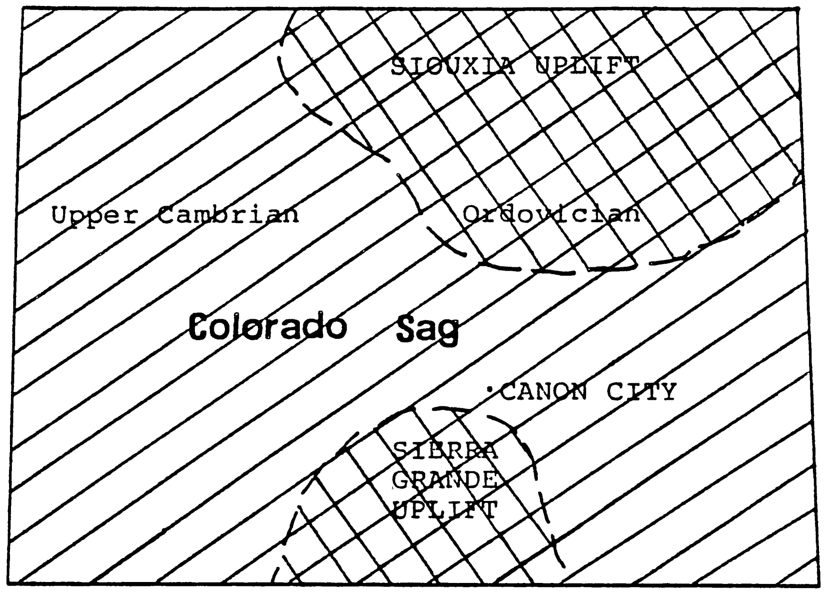


Figure 4. Worm's Eye Map of the Cambrian and Lower Ordovician Rocks on the Pre-Cambrian at the Close of the Early Ordovician in Colorado (After Haun and Kent, 1965)

CHAPTER IV

EARLY PALEOZOIC ERA

Geological Setting

Beginning in Cambrian time, there was a progressive spreading of shallow epicontinental seas eastward from the Cordilleran Geosyncline and northwestward from the Anadarko basin. Late in Cambrian time when the Sawatch was deposited in the embayment, the Western sea may have united with the southeastern sea in the Colorado Sag (Figure 4). At the close of the Cambrian, gentle uplift in the studied area exposed the Sawatch Formation and created the unconformity between the Sawatch and the overlying Manitou.

In the Early Ordovician, the sea again transgressed, due to either downwarping of the land or to a rise in sea level, and then the Manitou Dolomite was deposited in a shallow marine environment. The unconformable contact between the Manitou Dolomite and Harding Sandstone and the pre-Harding weathering and erosion of the Manitou indicate that the Canon City area was uplifted before the deposition of the Harding Sandstone in a northwestward transgressing sea. The area was uplifted again at the close of the Harding time, which is evidenced by the following observations: (1) the Fremont Dolomite unconformably overlies the

Harding Sandstone; (2) the Fremont becomes thicker from northeast to southwest, which could indicate a northeastward transgressing sea; (3) east of Phantom Canyon, the Fremont, which occurs only in patches, is upper Fremont, i.e., thinning in this area is due to both stratigraphic overlap and pre-Fountain erosion, suggesting that a high land existed in this area before the deposition of the Fremont (Lee, 1923; Monk, 1954).

According to Haun and Kent (1965), the entire Rocky Mountain region was possibly covered by the Silurian sea. The absence of Silurian sediments in the Canon City area could be the result of pre-Devonian and subsequent periods of erosion. During the Late Devonian and Mississippian times, the Williams Canyon Formation and possibly some younger rocks were deposited. However, the pre-Fountain erosion eliminated much of both, making it difficult to delineate the structural features which were generated before the rise of the Ancestral Rocky Mountains.

Petrology and Depositional Environment

Sawatch Sandstone Formation Outcrop and Stratigraphic Relations

The oldest sedimentary rock in the studied area is the Sawatch Sandstone, which crops out east of the mouth of Grape Creek. The formation is overlapped by the Manitou Dolomite, and rests unconformably on the Pre-Cambrian.

Lithology. The Sawatch contains six inches of basal conglomerate, which changes sharply into a shale layer measuring one foot. White to red and medium-to-coarse-grained sandstone overlies this shale. According to Miller (1951), the total thickness of the formation varies from six to ten feet.

Depositional Environment. The basal conglomerate may record the initial transgression of the Cambrian Sea and the overlying thin layer of shale deposited in lagoons. The sandstone, the youngest unit, may have accumulated in a turbulent near-shore environment.

Age. Based on the study of trilobite, which is found in this formation, Darton (1904) concluded that the Sawatch is of the Late Cambrian age and may correlate to the Reagan Sandstone in Oklahoma.

Manitou Formation

Outcrop and Stratigraphic Relationships. The outcrop of the Manitou is nearly continuous in a narrow belt throughout the embayment. Frederickson (1961) observed that the Manitou is absent about one mile south of Priest Canyon, which is either due to the pre-Harding erosion or to the onlap of younger beds. Figure 5 shows the correlation of this formation at locations in the embayment which are indicated on Figure 6.

The Manitou Formation rests nonconformably on the surface of Pre-Cambrian granites and metamorphic rocks throughout the embayment, with the exception of the location east

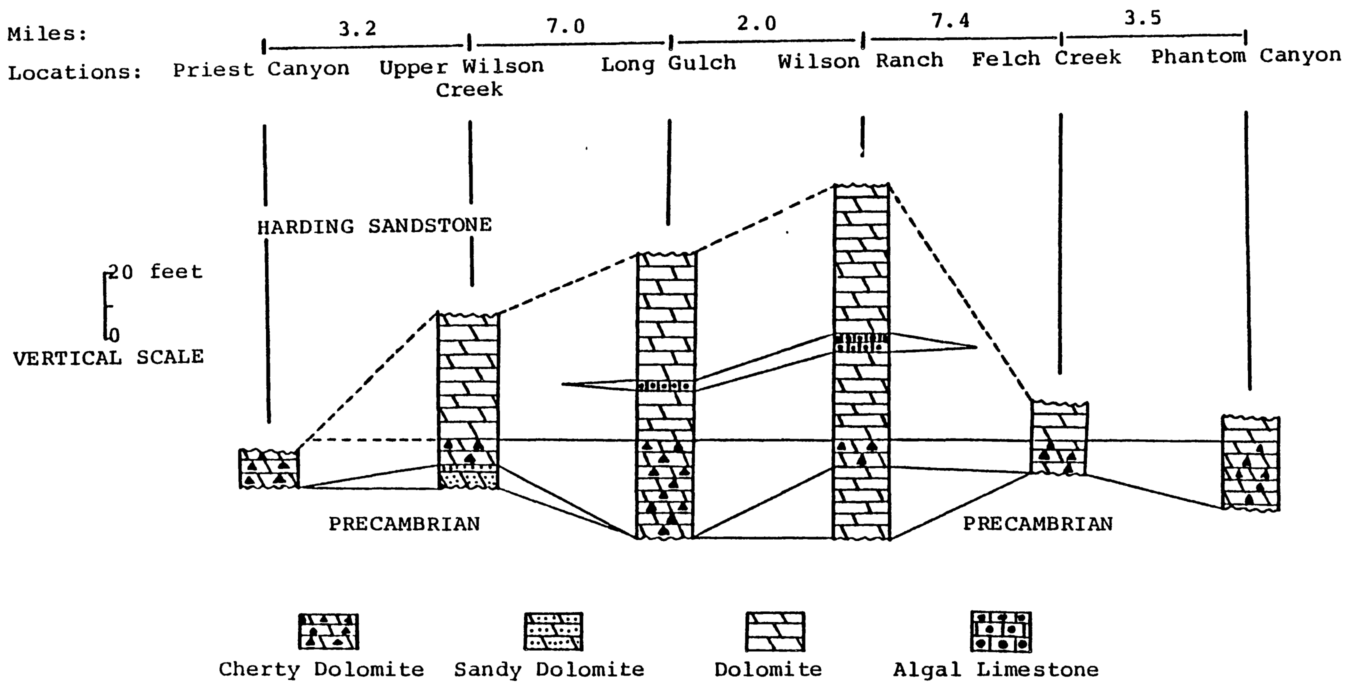
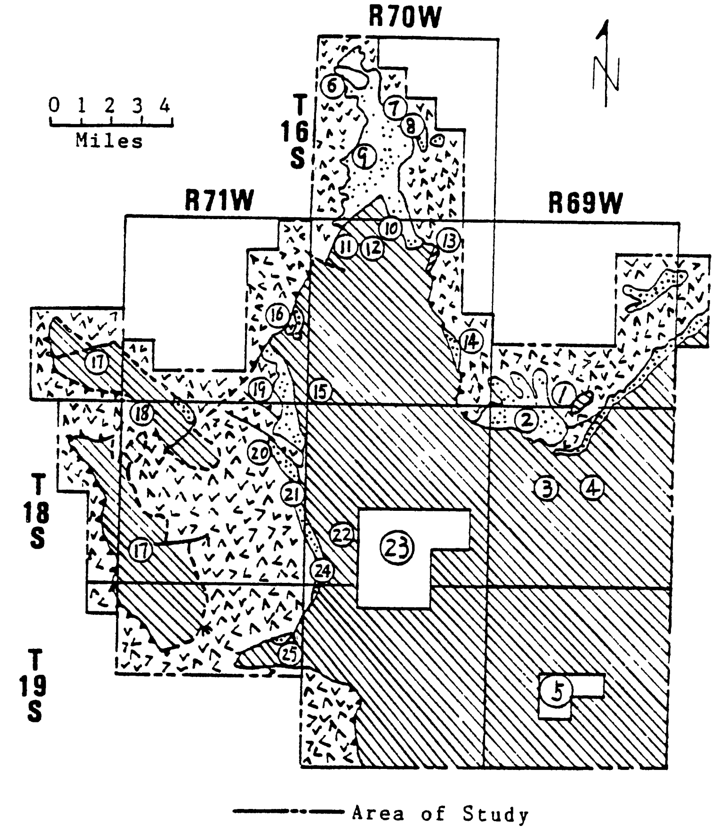


Figure 5. West-East Cross Section of the Manitou Dolomite in the Canon City Embayment. Locations Shown on Figure 6 (After Frederickson, 1961, and Gerhard, 1967)



1. Phantom Canyon
2. Six Mile Creek
3. Indian Spring
4. OK Geology Camp
5. Florence
6. Red Hill
7. Helena Canyon
8. Wilson Ranch
9. Long Gulch
10. K.U. Camp
11. Red Canyon Park
12. Garden Park
13. Cooper Mountain
14. Felch Creek
15. Shaw's Park
16. Upper Wilson Creek Area
17. Parkdale Area
18. OK Corral
19. North Twin Mountain
20. South Twin Mountain
21. Priest Canyon
22. Skyline Drive
23. Canon City
24. Harding Quarry
25. Mixing Bowl

Figure 6. General Geological Map of the Canon City Embayment, Showing Geographic Features Mentioned in Text. Pre-Cambrian Rocks Are Checked, Pre-Pennsylvanian Rocks Are Stippled, and Pennsylvanian and Younger Rocks Are Lined.

of the Grape Creek mouth where it unconformably overlies the Sawatch Sandstone Formation. The contact between the Manitou Dolomite and underlying crystalline rocks is very sharp. Locally, the basal Manitou beds contain fragments of granites and metamorphic rocks (especially pebbles of vein quartz); however, the dense, finely crystallined dolomite is more commonly in direct contact with the Pre-Cambrian basement (Frederickson, 1961). This condition was explained by Crosby (1899) as being due to a gradual transgression of the Cambrian Sea with accompanying planation (flattening) and reduction of the resulting debris to fine sand grains and clay-size particles. Because the Sawatch/Pre-Cambrian contact is almost a perfect plane, Brainerd, Baldwin, and Ketye (1933) thought that an intervening period of erosion completely removed the Sawatch Sandstone over a large area and that the Manitou Dolomite therefore rests on the surface of the Pre-Cambrian rocks. The Manitou is disconformably overlain by the Harding sandstone. Flower (1952) found reworked silicified cephalopods from the Manitou Formation in the lower beds of the Harding Sandstone.

Lithology. The formation is divided into three lithologically distinct members (Gerhard, 1967). The bottom member, named the Helena Canyon member, crops out in the Helena Canyon. It is composed of massive pink dolomite and is about 100 feet thick. Locally, a basal conglomerate of subangular to round quartz fragments occurs. The middle unit, named the Cherty member, is composed mainly of thin- to block-bedded pink to white dolomite and some chert beds which are irregularly bedded. This member is up to 45 feet thick. In locations where the Helena Canyon member is not present, the cherty unit onlaps the Pre-Cambrian basement with an extensive weathering surface separating the two (Frederickson, 1961). This member conformably overlies the Helena Canyon. The top member, named the Massive member, consists of a lower massive-bedded dolomite with shaly partings and a higher thin-bedded dolomite with finely etched banding. It is gray to red with a thickness of up to 75 feet. Frederickson (1961) found an algal limestone bed composed of mottled pink and white limestone that is full of concentrically banded and spherical limestone masses with diameters of about .05 inches. The Massive member conformably overlies the Cherty member.

Age. The cephalopods in the Manitou Formation indicate a Middle or Late Canadian age. Maher (1953) believed that the formation can be correlated with the Arbuckle Group in the subsurface of Eastern Colorado and Western Kansas. Cloud and Barnes' (1948) paleontological study also suggested an Early Ordovician age and a correlation with the Arbuckle Group.

Depositional Environment. Based on his study of bedding types, sedimentary structures, and kinds of fossils in the Manitou Formation, Roghani (1975) concluded that the formation was deposited mostly in subtidal and partly in intertidal and supratidal environments. He divided the whole sequence into twelve units and determined the environment for each (Figure 7). In the Canon City area, the absence of oolite, grainstones, and well-sorted structures and the predominance of mudstones and wackstones, strongly suggests that the water body was quiet during the deposition of the Manitou. Roghani's (1975) study also showed that the variation both in the textural types of bedding and in the amount of fossils from the subtidal to supratidal rocks is gradual, and that only a thin zone of the rocks which were formed in the intertidal environment is present. The thinness of the intertidal rock zone suggests that the tidal range was very small.

Harding Formation

Outcrop and Stratigraphic Relationships. Exposures of the Harding Formation continue almost uninterrupted from the Grape Creek area west of Canon City to beyond Beaver Creek. Minor disturbances of its outcrop are due either to faulting or to erosional onlap of the Fountain Formation. The Harding is locally repeated by faulting.

The Harding Formation is disconformably overlain by the Fremont Formation. The exception is near the Phantom Canyon area where Pre-Fountain erosion removed the Fremont dolomite and a portion of the upper Harding Sandstone. Figure 8 shows the correlation of the Harding in the Canon City area.

Lithology. Tolgay (1952) divided the Harding Sandstone into six units: a basal conglomerate and a lower, middle,

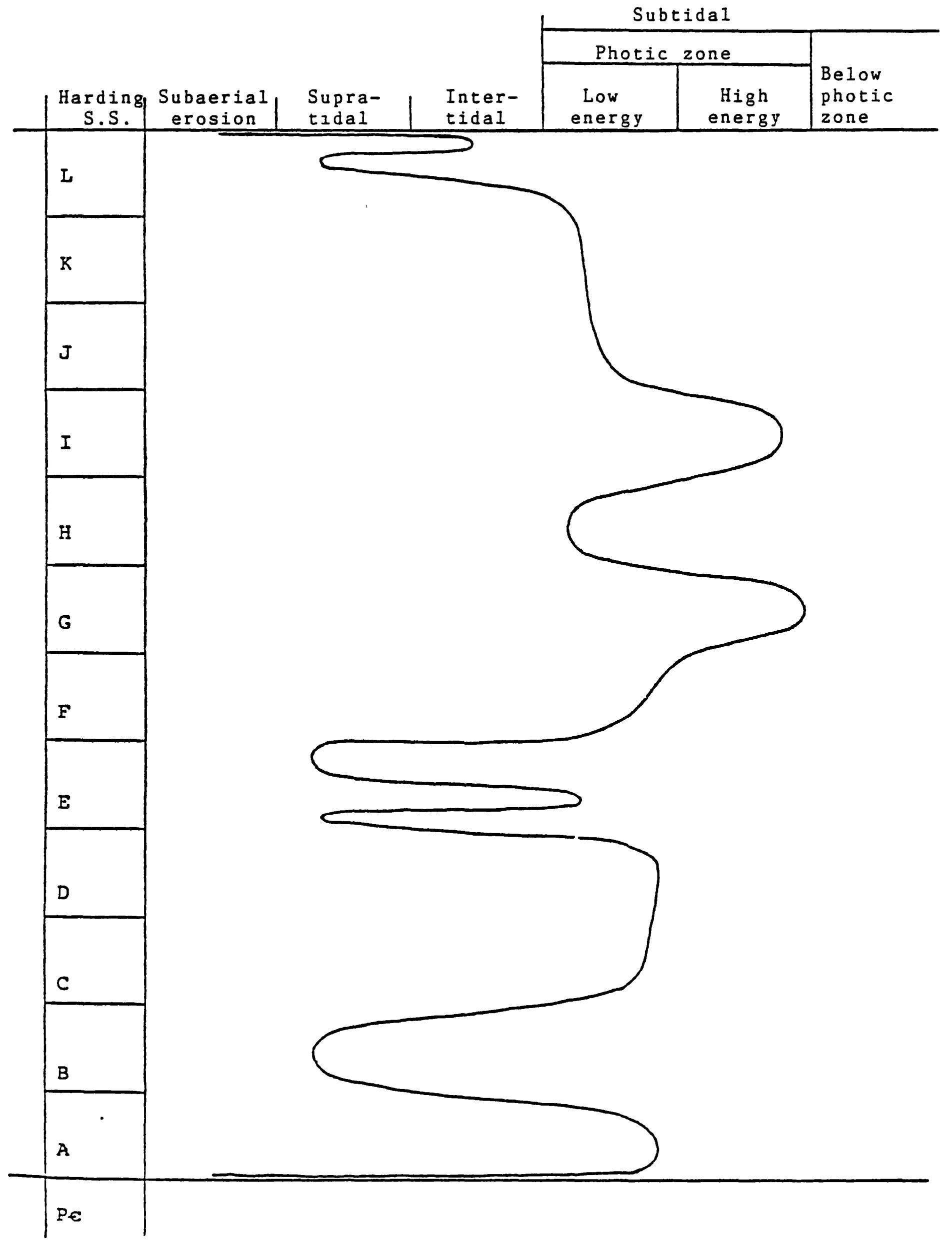


Figure 7. Depositional Environments of the Manitou Dolomite. The Whole Sequence is Divided into Twelve Units (A, B, C,...) (After Roghani, 1975)

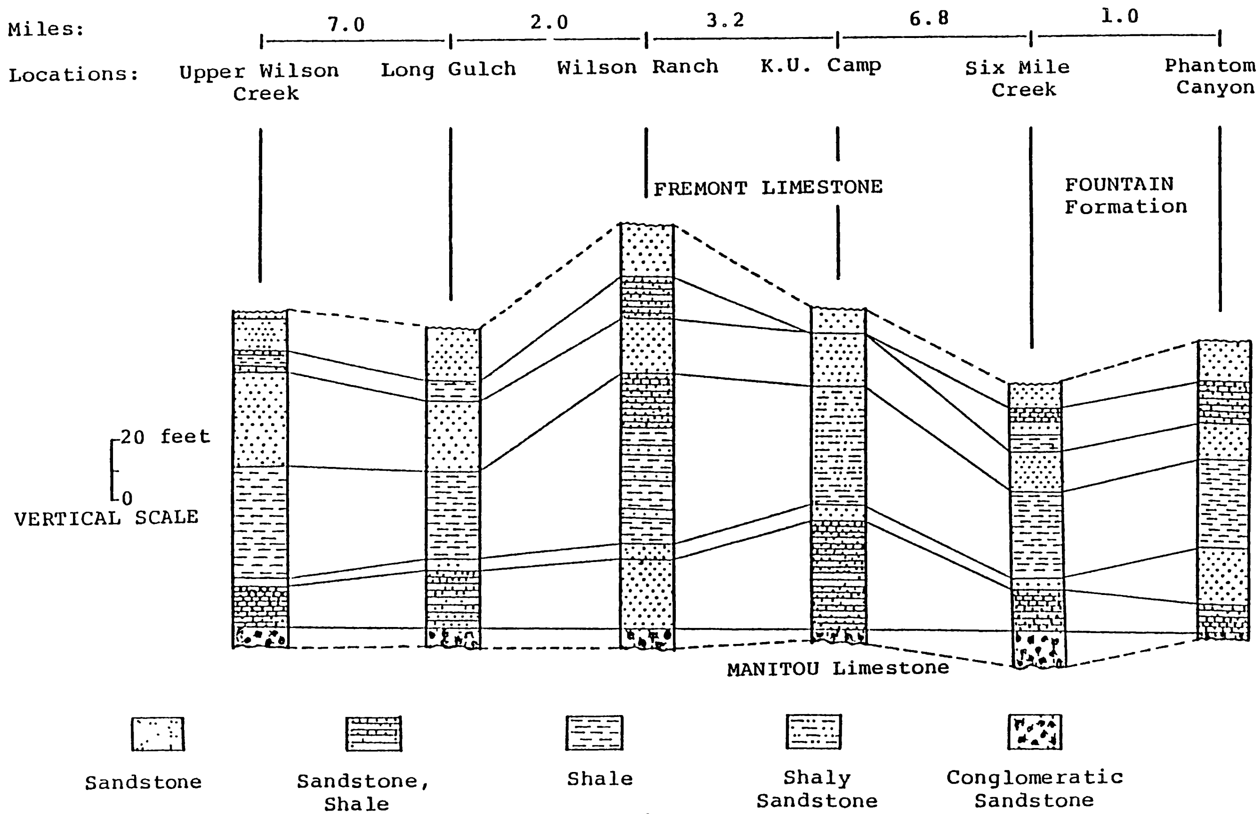


Figure 8. West-East Cross Section of the Harding Sandstone in the Canon City Embayment (After Frederickson 1961, and Gerhard, 1967).

and upper sandstone interrupted by two shale layers. The basal unit is a quartzose, bimodal, coarse-grained sandstone composed of angular to well-rounded quartz and chert pebbles in a fine-to-medium quartz sand matrix. This unit has an average thickness of six feet, and fines upward into the lower sandstone unit without a distinct break. The three sandstone units, which are very fine-grained, occur consistently throughout the embayment. The lower sandstone thins northeastward and contains abundant iron oxide (hematite). The middle sandstone is massively bedded. Frederickson (1961) found that the middle sandstone contains interbedded shale in the Garden Park area. The thickness of this middle unit varies from 5 to 28 feet. The upper sandstone unit is characterized by the red and yellow color of beds and bluish-white phosphatic fish plates. This unit ranges in thickness from 12 to 20 feet, and its beds are massive and are laminated by iron oxide residues which were left by the moving front of the interstitial water (Tolgay, 1952). The lower of the two intervening shale layers forms a definite bench in exposure and is green, brownish to pink, fissile, and 15 to 43 feet thick. The upper is green, deep red to pink, and 2 to 14 feet thick.

Thin-Section Description. Thirty-five samples from the Mixing Bowel, Twin Mountains and Phantom Canyon areas were analyzed under a polarizing microscope. Monocrystalline quartz (71.5-94%) is the main component and feldspar and polycrystalline quartz are in trace amounts. Microcline and albite twins are common. The polycrystalline quartz grains are present in a small amount and composed of two or three single quartz crystals. Secondary minerals are calcite and quartz overgrowths, both of which heavily cement the rocks. Fragments of phosphatic fish bone are common in some specimens (Figure 9). Generally, the bottom portion of the rock is well- to moderately-sorted and texturally mature (Figure 10), while the top portion is poorly sorted and texturally immature. Obviously, in the Twin Mountains area, the monocrystalline quartz content (Figure 11) decreases upwards in the sampled section as the contents of carbonates and matrix increase (Figure 12); the variation of secondary silica is not apparent (Figure 12); Figure 13 shows the variation of matrix/monocrystalline quartz, the upward increase of the ratio indicates that depositional environments became quiet at the close of the Harding time. In the Mixing Bowl area, polycrystalline quartz disappears wile monocrystalline quartz increases upwards (Figure 14), apparently indicating that the compositional maturity of the Harding Sandstone increases upwards and that the dynamics of the sedimentational process became stronger later on. However, this relationship is more probably related to an upward decrease in grain size (Figure 15). Figure 16 shows that two intervals of the Harding in the Mixing Bowl area have a high ratio of matrix/quartz, suggesting the presence of a low-energy environment. The distribution patterns of carbonates and secondary silica suggest that their formations were controlled by the permeability of the beds as

Figure 9. Fragments of Fish Bone and Fine Quartz and Zircon Grains in the Harding Sandstone, 10 x 10 (-) (Twin Mountains Area)

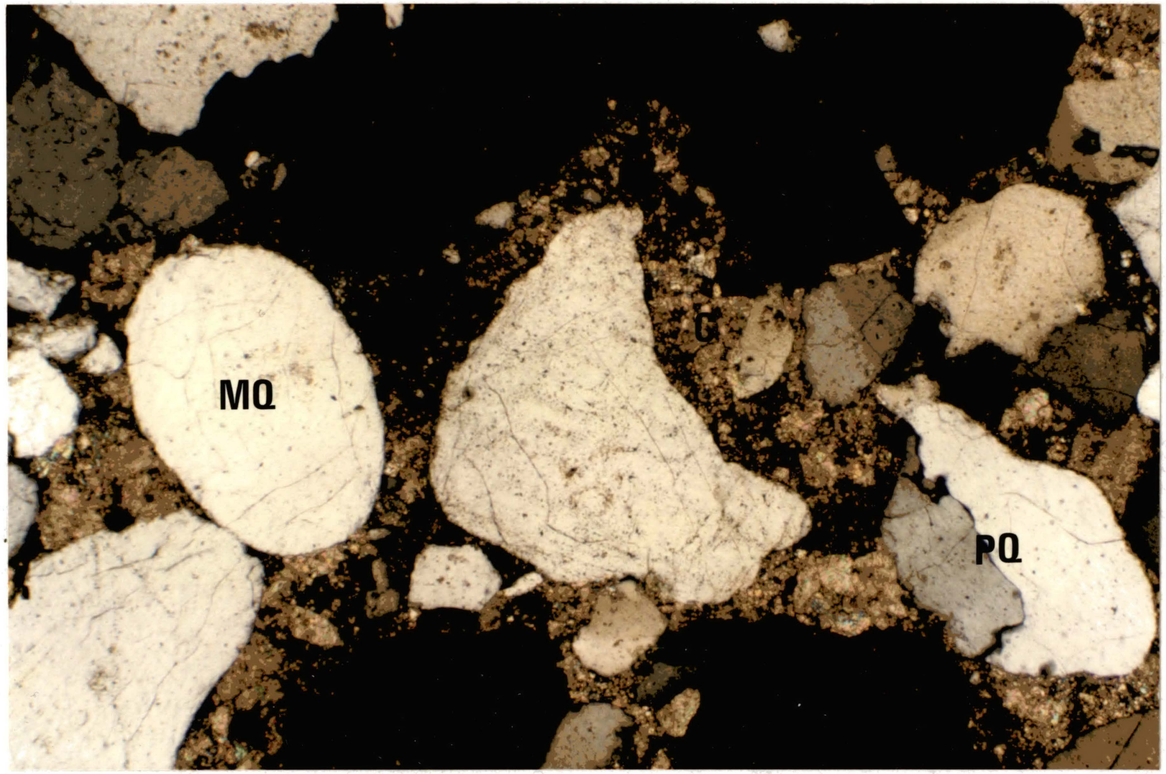


Figure 10. Texturally Mature Harding Sandstone, MQ: Monocrystalline Quartz, PQ: Polycrystalline Quartz, and C: Calcite, 10 x 10 (+) (Mixing Bowl Area)

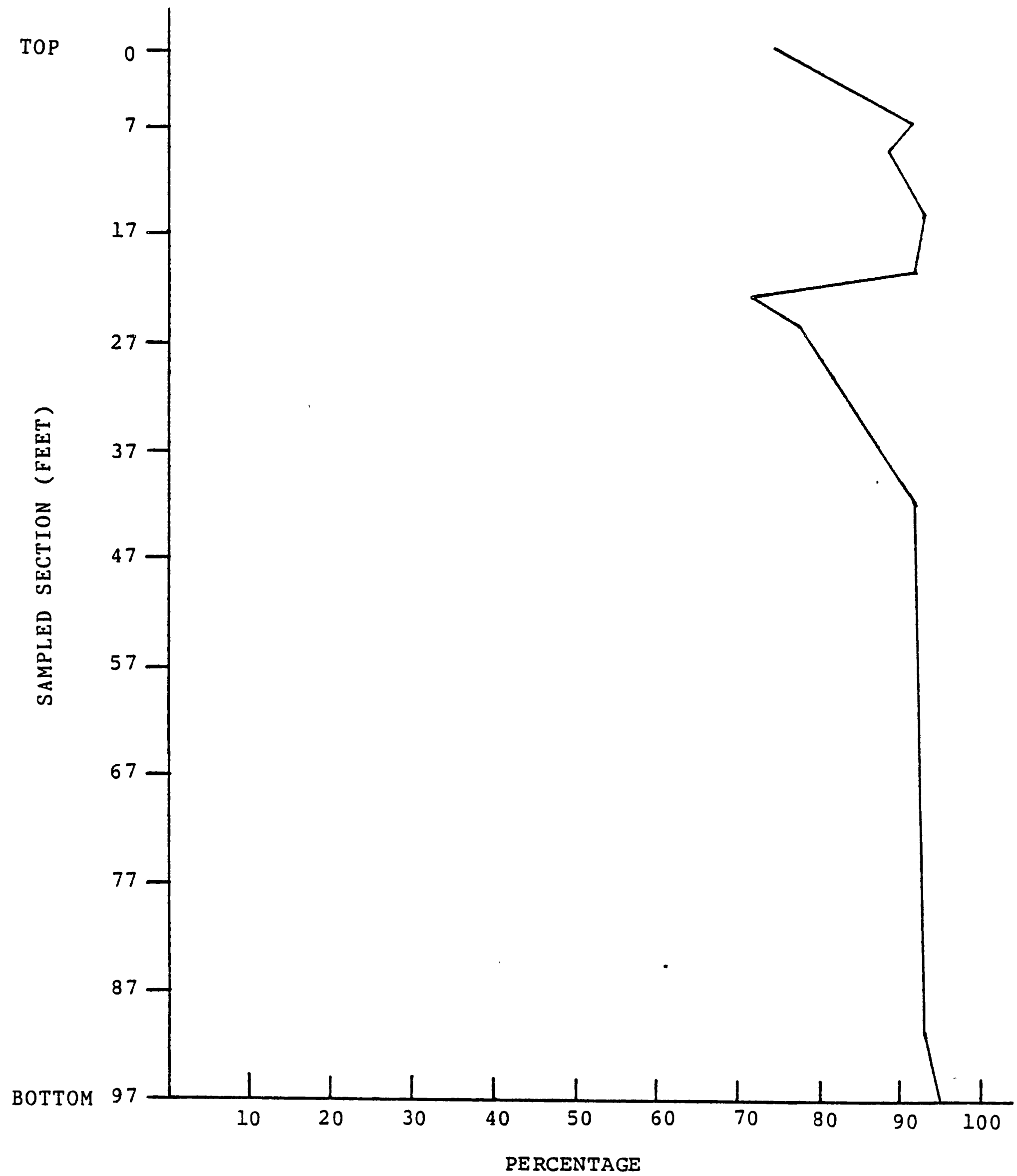


Figure 11. Vertical Variation of Monocrystalline Quartz in the Sampled Section of the Harding Sandstone (Twin Mountains Area)

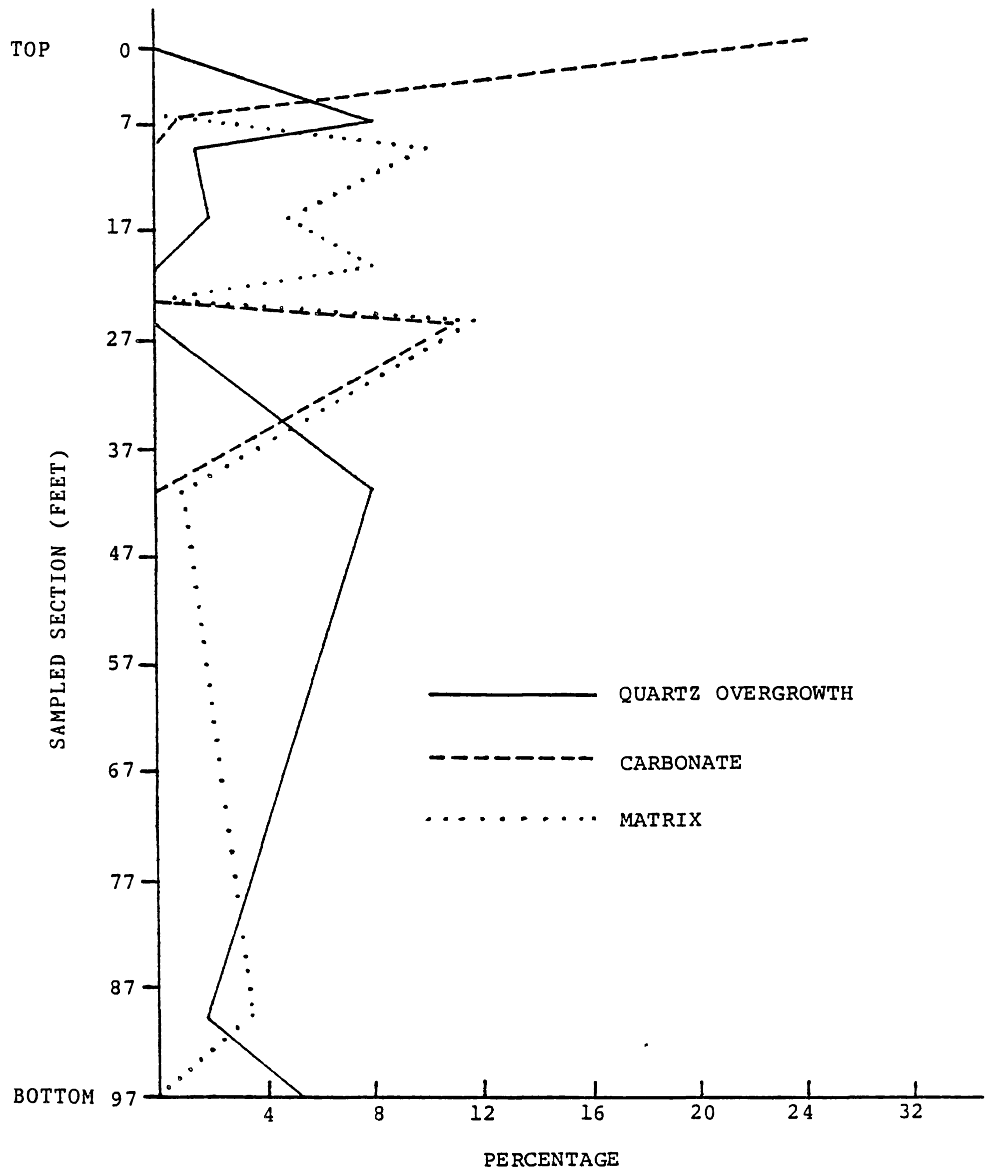


Figure 12. Vertical Variations of Secondary Minerals and Matrix in the Sampled Section of the Harding Sandstone (Twin Mountains Area)

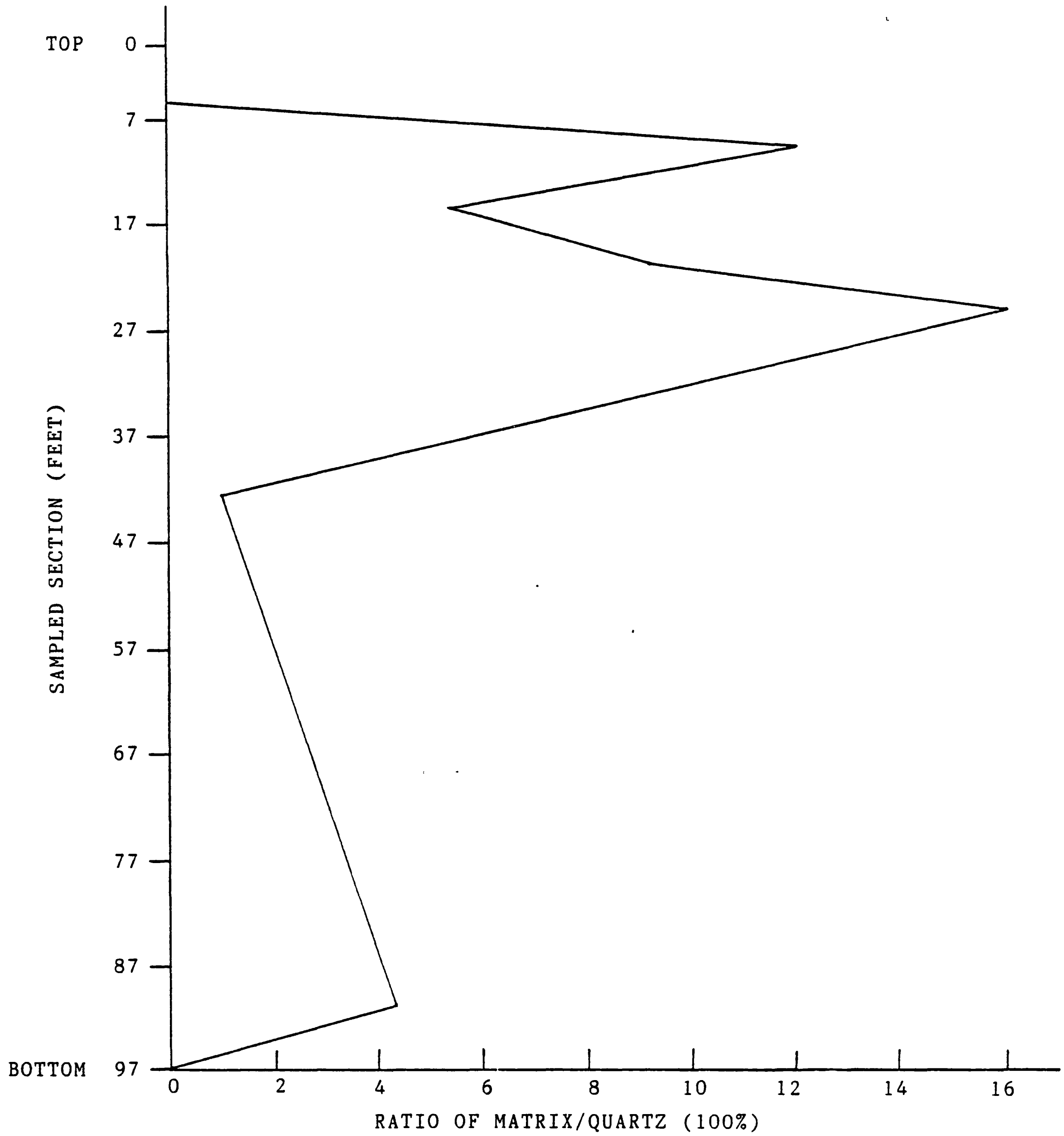


Figure 13. Vertical Variations of Matrix/Monocrystalline Quartz in the Sampled Section of the Harding Sandstone (Twin Mountains Area)

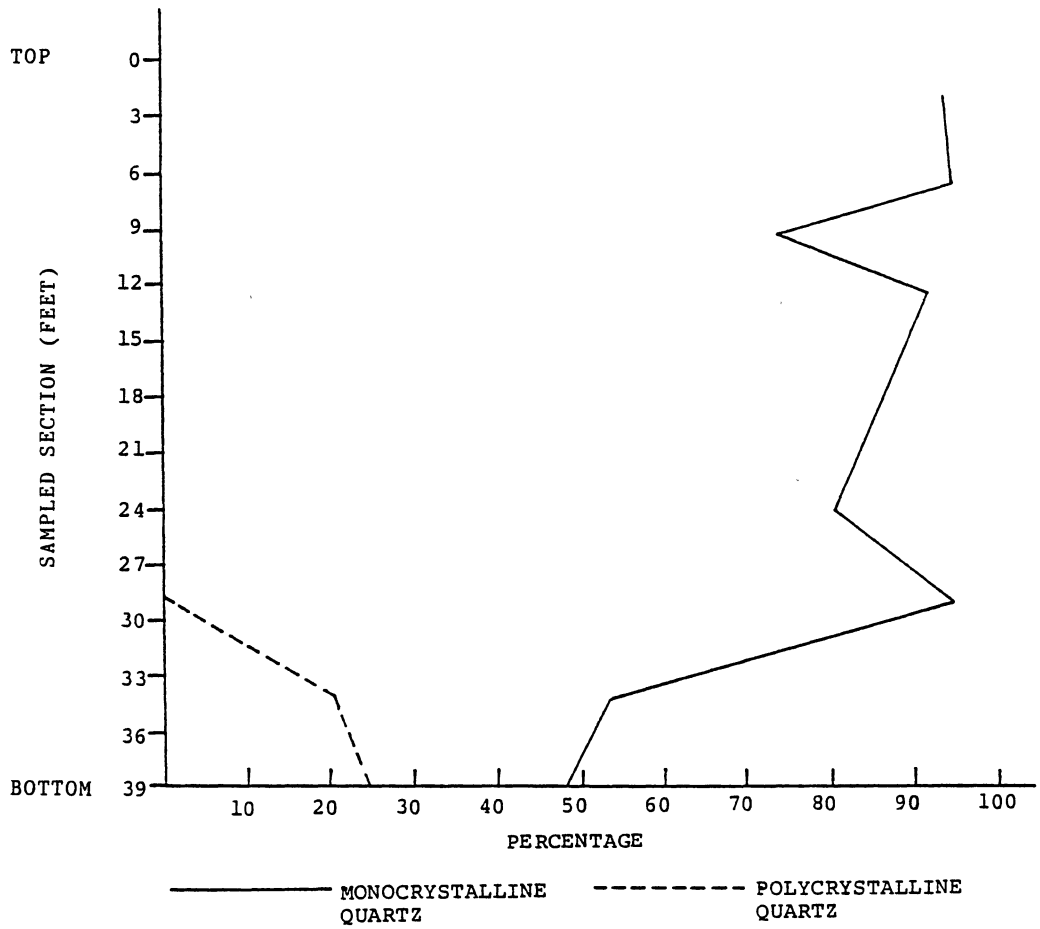


Figure 14. Vertical Variations of Monocrystalline Quartz and Polycrystalline Quartz in the Sampled Section of the Harding Sandstone (Mixing Bowl Area)

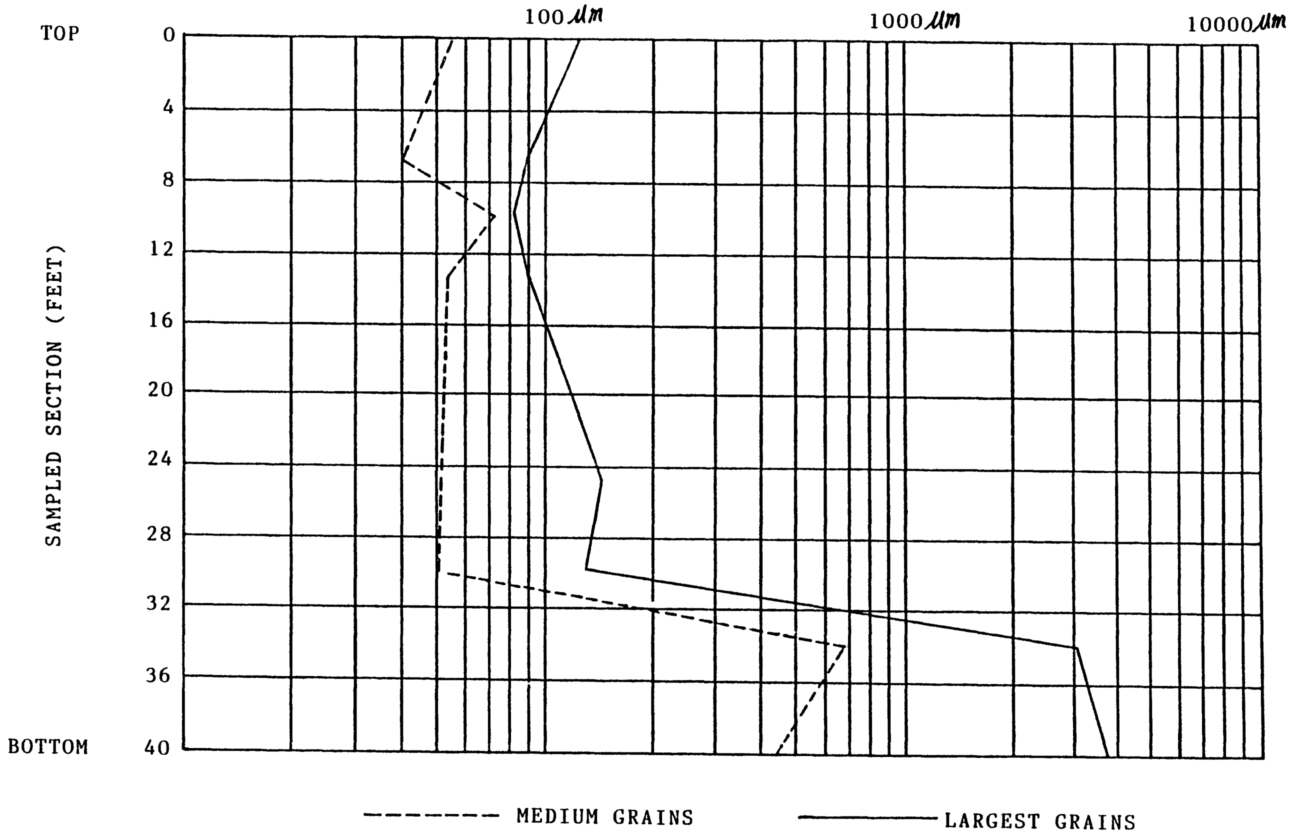


Figure 15. Vertical Variation of Grain Size in the Sampled Section of the Harding Sandstone (Mixing Bowl Area)

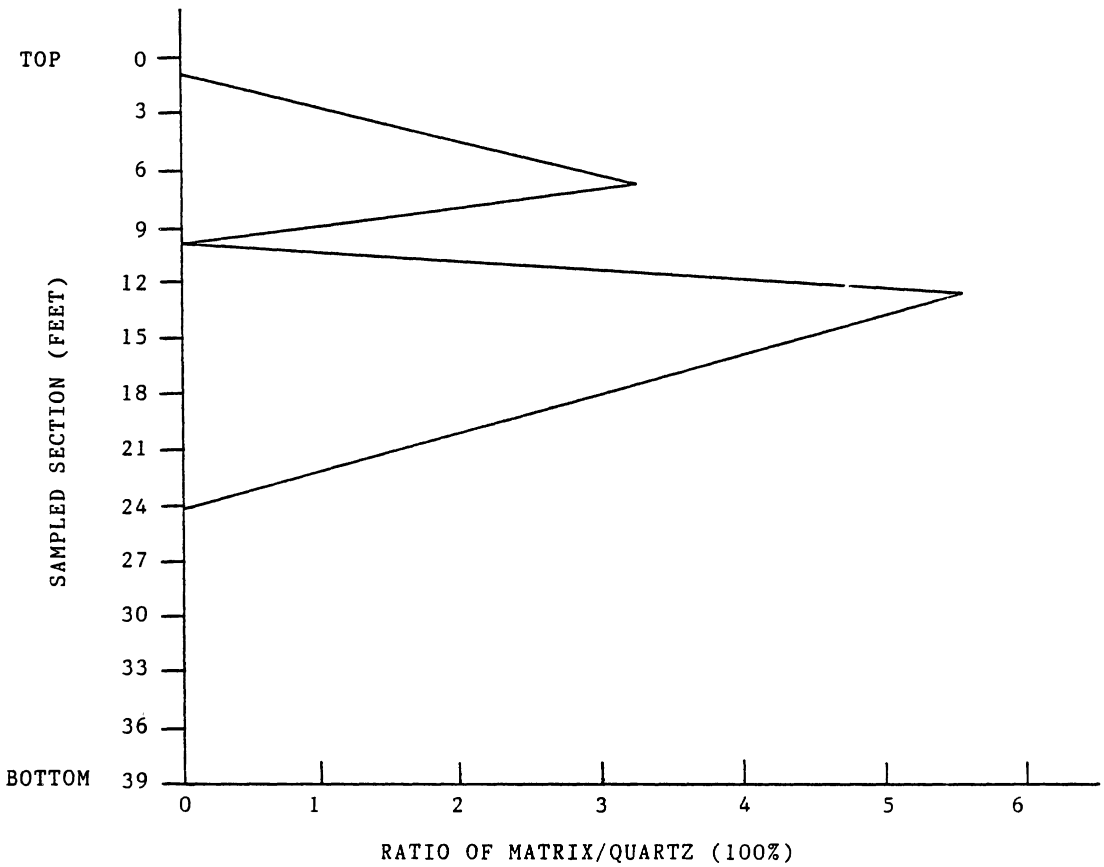


Figure 16. Vertical Variation of Matrix/(Monocrystalline Quartz + Polycrystalline Quartz) in the Sampled Section of the Harding Sandstone (Mixing Bowl Area)

recorded by matrix content (Figure 17). The permeable beds with less matrix have more carbonate cements than the non-permeable beds. It is interesting to note that the more carbonates there are, the less secondary silica there is. This contrary relationship probably records a variation in the pH of ground water between acidic conditions which favor the formation of secondary silica and alkalinic conditions which favor the formation of carbonates.

Diagenesis. Samples in the Twin Mountains and Phantom Canyon areas show the following diagenetic sequence (Figure 18):

1. The dissolution of matrix. A significant amount of matrix was removed in top samples HNI-1 and HNI-2 and bottom samples HNI-9, HNI-13, and HNI-14. However, only a small portion of matrix was dissolved in the middle samples HNI-3, HNI-4, HNI-5, HNI-7, and HNI-8. The greater dissolution of the matrix in the bottom and top samples suggests that alkaline solutions migrated from both the top and bottom of the Harding Formation toward the middle;

2. The precipitation of secondary silica and quartz overgrowths. The silica source was possibly from the dissolution of matrix;

3. The precipitation of carbonates and their replacement of quartz grains and secondary silica. In sample HNI-1 nearly all the secondary silica was replaced by calcite. During the precipitation of some of the carbonates, a reducing environment was developed and the ferric iron, possibly

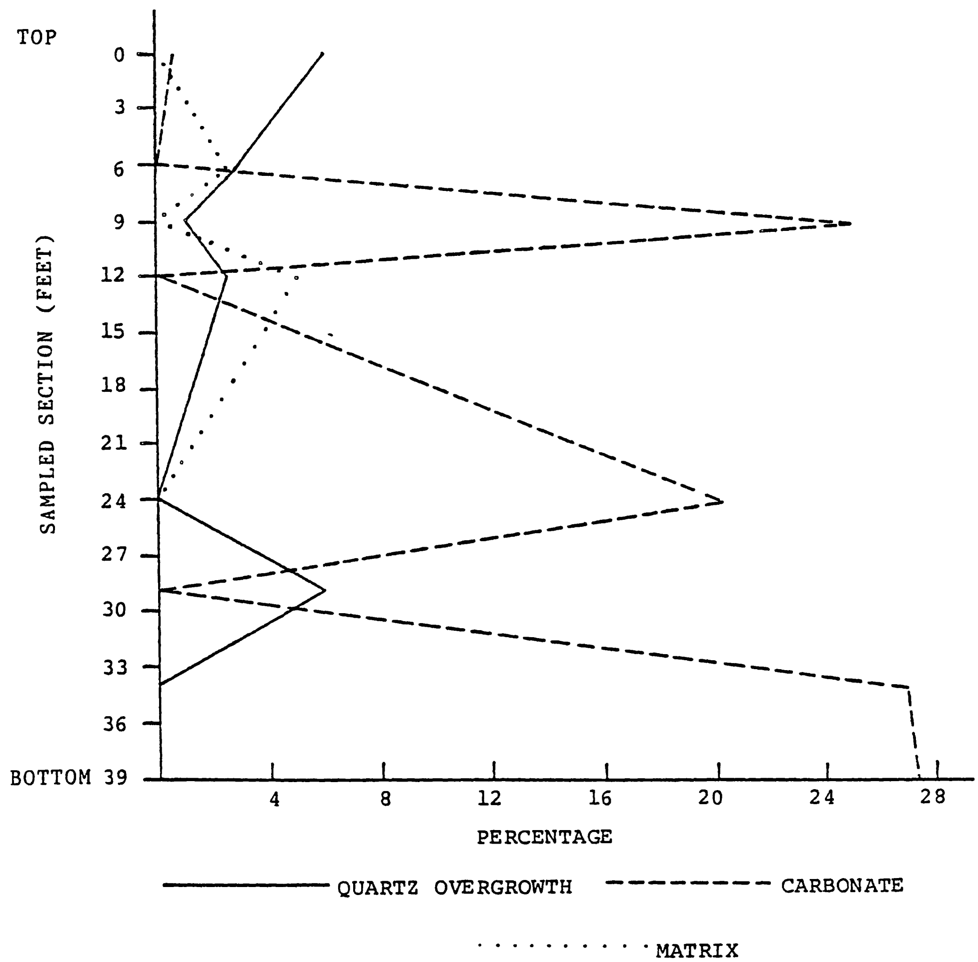


Figure 17. Vertical Variations of Secondary Minerals and Matrix in the Sampled Section of the Harding Sandstone (Mixing Bowl Area)

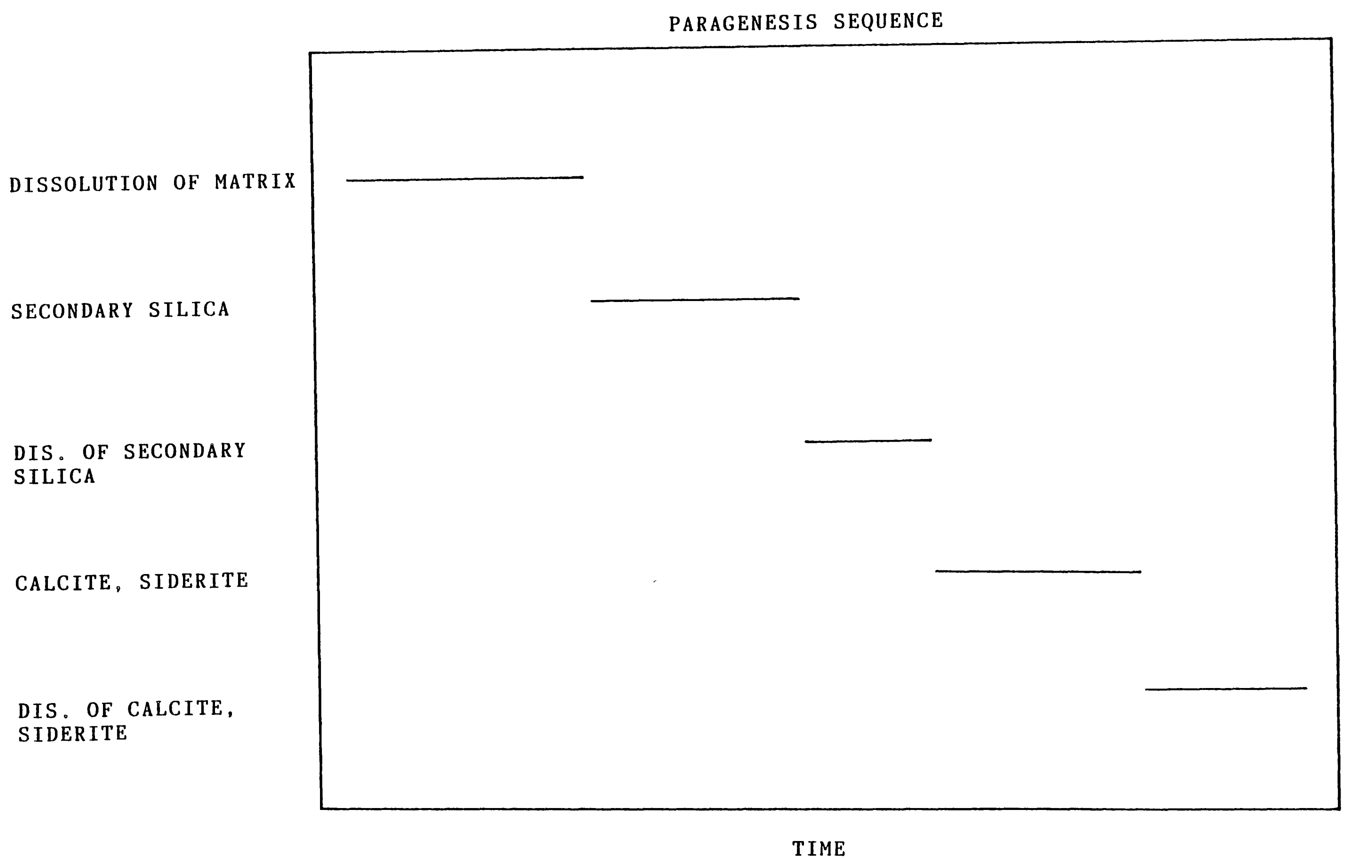


Figure 18. Diagenesis of the Harding Sandstone (Twin Mountains Area)

from hematite, was reduced to ferrous iron. Therefore, siderite was formed by the combination of the ferrous iron and carbonate anion, the latter was possibly derived from the dissolution of the Manitou Limestone fragments.

Samples in the Mixing Bowl area record a different diagenetic sequence than the samples in the Twin Mountains and Phantom Canyon areas (Figure 19). The first diagenetic solutions recorded were acidic and rich in silica, resulting in quartz overgrowths. Later solutions were alkaline, leading to the dissolution of the matrix and some other silicate minerals. Secondary silica, which cements the rocks heavily, was then precipitated in an acidic environment. When the formation water was alkaline again, the first generation of calcite and siderite formed, replacing some secondary silica. Thereafter, the rocks were fractured (sample HM-16), and the fractures filled by second-generation calcite. The dissolution of calcite and siderite was the last diagenetic event recorded.

Age. Various paleontological studies by Johnson (1945), Pollack (1951), and Sweet (1954) show that the basal conglomerate and lower sandstone are the Late Chazyan, and the remainder of the Harding is of the Black River age. The Harding Sandstone can be correlated with the lower part of the Simpson Group of Oklahoma.

Depositional Environment. The lithology, paleontology, sedimentary features, and petrographic characters of the conglomerate and sandstones suggest that they were deposited

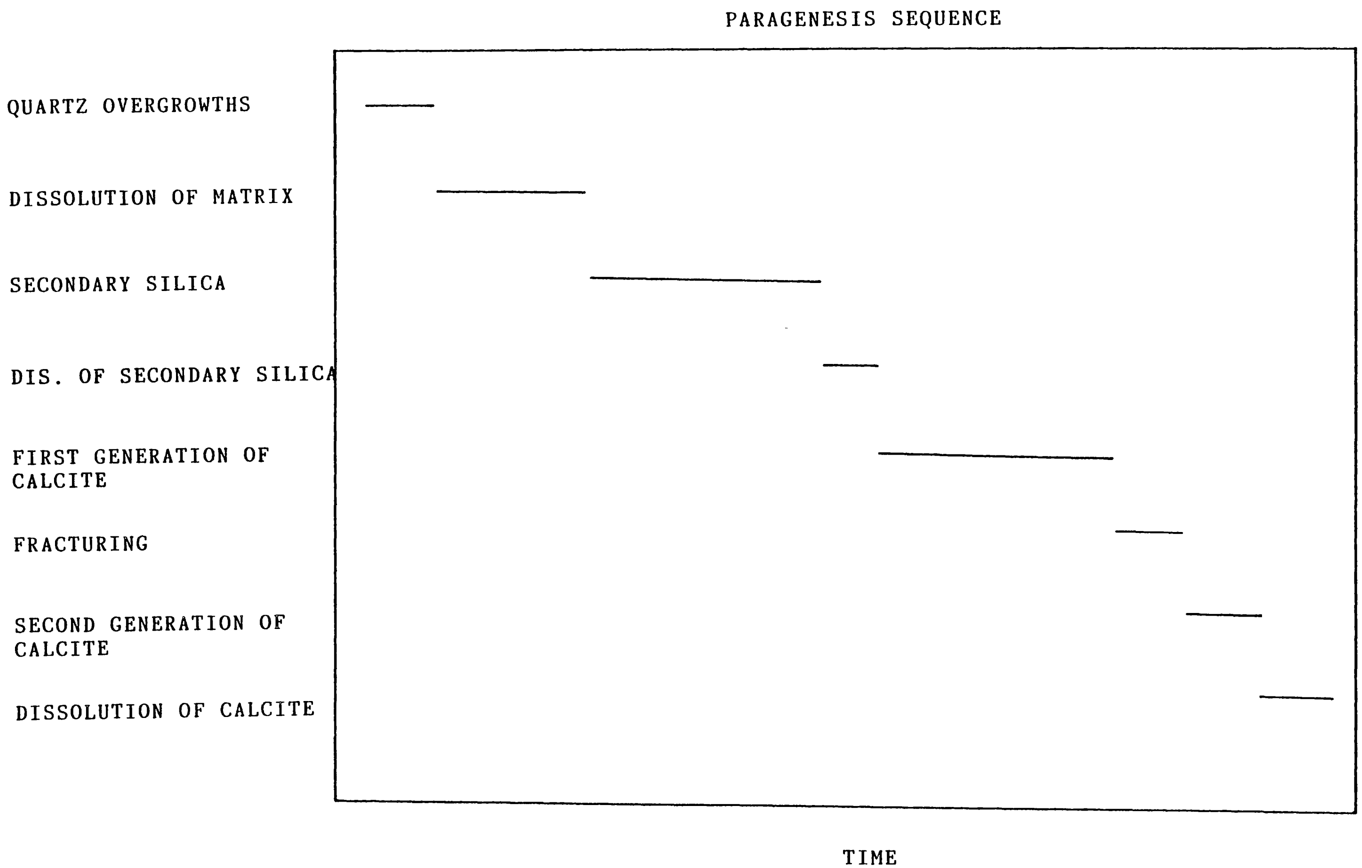


Figure 19. Diagenesis of the Harding Sandstone (Mixing Bowl Area)

in a turbulent shelf margin environment at a slow rate. Therefore, waves and currents had enough time to winnow out finer particles and round coarser grains. A considerable number of extremely round quartz grains were found during examination of the thin sections of the Harding. On the basis of the uniformity and maturity of mineral suite, fineness, and angularity of quartz grains,
Tolgay (1952) concluded that the conglomerate and sandstone was deposited in an infraneritic environment on a stable platform. The two layers of shale indicate a reduction in the rate of influx of coarser detrital sediments and/or lower energy conditions. Gerhard (1967) analyzed isopach maps of the Manitou and Harding, and came to the conclusion that the configuration of the distribution of their thicknesses is similar and that the two formations were laid down in a similar environment. However, the studied area received clastic sediments during the Harding time, but it received carbonate sediments during the Manitou time.

Tolgay (1952) studied the median size and the sorting coefficient of the upper sandstone unit and found that they increase toward the northwest. This finding suggests that the shore line of the Harding sea was northwest of the Canon City embayment.

Fremont Formation

Outcrop and Stratigraphic Relationships. The exposure of the Fremont Formation is nearly continuous around the margin of the Canon City embayment. The only discontinuity is in the vicinity of Phantom Canyon where pre-Fountain erosion removed the formation.

In the western part of the Canon City area, the Williams Canyon Formation overlies the Fremont unconformably. In other places, except for a few locations where Williams Canyon is present, the Fremont is overlain disconformably by conglomerates of the Fountain Formation. Figure 20 shows the correlation of the Fremont in the Canon City area.

Lithology. The formation is composed of crystalline to fine granular dolomite. The lower portion of the formation, which is massive and generally reddish to pink, contains a prominent zone of coral, Receptaculites, and nautiloid cephalopods (Sweet, 1954). The upper portion, which contains two purple marker beds, is mainly tan to light buff, thinly bedded, sandy, argillaceous, and fine-grained. A continuous 4- to 8-foot bed of laminated, cherty, and dolomitic shale occurs in the middle. Sweet (1954) calls the upper portion the "Priest Canyon Member," the middle portion the "Cherty member," and the lower portion the "Massive Member." Monk (1954), after examining diagnostic fossils, distinguished the same three intervals.

Age. The fauna, studied by Johnson (1954), Sweet (1954), and Monk (1954), shows that the Fremont Formation is equivalent of the Bighorn Formation in Wyoming, the Montoya Formation in Western Texas and New Mexico, and the upper

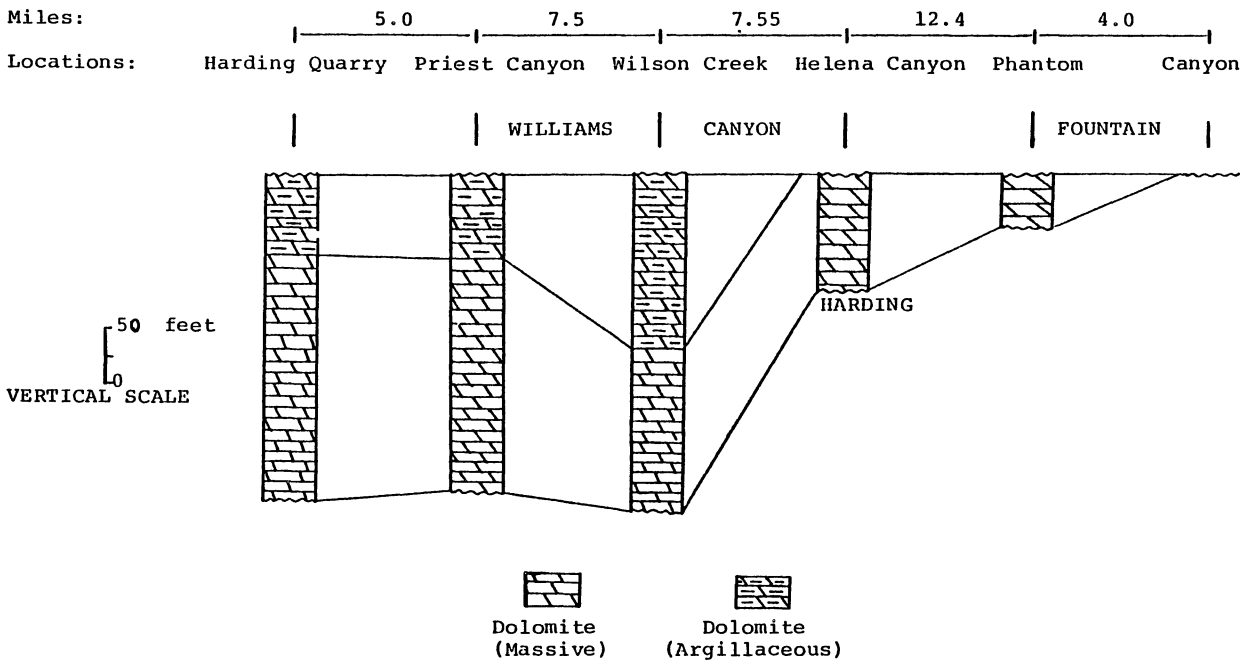


Figure 20. West-East Cross Section of the Fremont Dolomite in the Canon City Embayment (After Sweet, 1954)

part of the Viola Formation and Sylvan Shale of Oklahoma and Kansas.

Depositional Environment. The bedding, bioturbation, and the fauna of the Fremont suggest that the formation was deposited in a littoral and photic zone of a warm shallow sea (Monk, 1954). The presence of sand grains indicates the input of detritus during the carbonate precipitation.

Williams Canyon Formation

Outcrop and Stratigraphic Relationships. In the Canon City Embayment the Williams Canyon Formation occurs as erosional remnants beneath the Fountain Formation. According to Frederickson (1961), a continuous exposure extends from the north side of the Arkansas River northward to Wilson Creek along the western side of Shaw's Park. West of Red Canyon Park, Williams Canyon beds occur again in a small exposure and in several nearly horizontal remnants in the vicinity of the Kansas Geology Camp in east Garden Park near Millsap Creek. Northeast of Canon City, the Williams Canyon Formation was completely removed by erosion. Figure 21 shows the correlation of the formation in the Canon City area. This formation is unconformably overlain by the Fountain Formation throughout the entire Canon City area.

Lithology. Because the Williams Canyon Formation occurs as remnants, its thickness varies considerably between outcrops. Maher's (1953) investigation shows that the greatest thickness is in the Colorado Springs region, where one outcrop measures 53 feet.

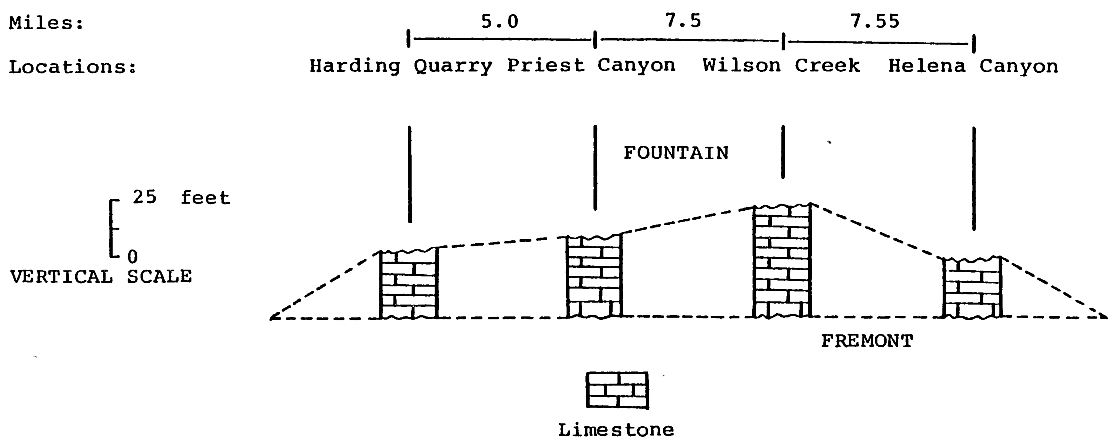


Figure 21. West-East Cross Section of the William Canyon Limestone in the Canon City Embayment (After Sweet, 1954)

This formation consists mainly of microcrystalline to finely crystalline light gray or purplish-gray limestone and dolomite; the thickness of individual beds ranges from 1 inch to 4 feet (Frederickson, 1961). Several thin, reddish-purple, contorted shale layers occur throughout the formation, and some carbonate beds are shaly. Quartz grains are present in all of the exposures, and the sand percentage increases northeastward. Robertson (1957) studied the quartz sand grains in detail and found that the grains are rounded and frosted. The frosted appearance is due both to the fine quartz overgrowths and pits on the surface. He thought that some of the to pits are caused by pressure and differential solution rather than abrasion.

<u>Age</u>. Because no significant fossils were found, it is difficult to determine the exact age of the Williams Canyon Limestone. Maher (1953) believed the formation is of a Mississippian age, but according to Lee (1915), the Early Mississippian fossils Maher found in the Garden Park area are actually associated with the basal part of the Fountain Formation. Based on the lithology and stratigraphic position of the formation, Brainerd et al. (1933) thought that it is of a Mississippian age. However, MacLachlan's (1961) subsurface study on the Paleozoic rocks in eastern Colorado suggested that the Williams Canyon is equivalent to part of the Chaffee Formation (upper Devonian).

<u>Depositional Environment</u>. The penecontemporaneous deformation of the shale layers, the flaggy bedding, and the

lithology all suggest that the formation was deposited on a shelf margin or a platform margin. Because the Williams Canyon limestone was left in patches by the pre-Fountain erosion, it is impossible to delineate the shore line's location or its distance from the Canon City area during the deposition of Williams Canyon.

CHAPTER V

LATE PALEOZOIC ERA

Geological Setting

Detailed mapping of the thickness and lithofacies of Pennsylvanian rocks in Southeastern Colorado shows that the present sites of Canon City, the Wet Mountains, and the Front Range of the Rockies were within a low landmass at the beginning of Pennsylvanian time. Later, the Morrowan sea from the Anadarko Basin encroached on the flank of the landmass, but failed to cover it. Near the close of Morrowan time, uplifting and faulting took place around the present site of the Front Range and Wet Mountains, giving rise to the Ancestral Rocky Mountains. The Wet Mountains and Twin Mountains faults may have begun or been rejuvenated at this time; Curtis (1958) believed that the size and extent of the Ancestral Rocky Mountains was nearly the same as that of the present Rockies. He explained the structures of the Ancestral Rockies with a block-faulting model, suggesting that the origin of the tectonic features is similar to the rift valleys of Africa. Lee (1923) studied the quantity of the debris from the Ancestral Rockies and concluded that they were as high as the present Rocky Mountains. According to Maher (1953), a trough, which extended from the Front Range area into the Hugoton embayment in Oklahoma, was formed with the creation of the Ancestral Rockies. The trough was covered by a sea at the end of Morrowan time. During the following Atokan time, the coarse arkosic clastics which make up the Fountain Formation were deposited along the flank of the mountains, and shale and limestone were laid down in the offshore marine waters. At the close of Atokan time, another event of uplifting and faulting occurred, leading to the invasion of the area southeast of Canon City during Des Moinesian time. The margin of the trough (the studied area) which was adjacent to the mountains still received coarse arkosics while the deep part of the trough received shale and limestone sediments. During the following Missourian time, the steep cliffs, which were created by the tectonic movement at the end of Morrowan age and Atoka time, were so worn down that only a small amount of coarse clastics was generated. By the end of Missourian time, a gentle upwarping took place in southeastern Colorado, leading to the withdrawal of the sea. During the following Virgilian time, the sea expanded, but failed to cover the Canon City area.

Petrology and Depositional Environment

Fountain Formation

Outcrop and Stratigraphic Relationships. The Fountain Formation is present throughout the Canon City area except along the Grape Creek ridge, where it is absent due to the onlap by conglomerates of the Ralston Creek Formation (Schulze, 1954). According to Frederickson, Delay, and Saylor (1956), a break exists between Fountain and Lykins deposition, and the Fountain Formation is disconformably overlain by the Ralston Creek Formation in the western part of the area, where the Lykins Formation was completely removed by erosion.

Lithology. The Fountain Formation is composed of two major lithologies: (1) a lower zone of arkosic conglomerates, conglomeratic sandstones, sandstones and siltstones, and (2) an upper zone of conglomeratic sandstones, siltstones, and nodular limestones. Generally speaking, these two major lithologic variations represent a fining upward sequence.

The lower zone is a series of highly cross-stratified arkosic sandstones and conglomerates. The rocks are red, friable, and poorly sorted. Some subrounded pebbles and cobbles are present. Festoon-type cross-bedding is associated with significant lateral and vertical variations.

The upper zone consists mainly of alternating purple to red, fine- to coarse-grained, cross-stratified, sandstones and conglomeratic sandstones interbedded with purple to red arkosic siltstone. The grain size, scale of cross-stratification, and thickness of beds become smaller toward the top of the Fountain. Some prominent lenticular limestone beds at the top portion are associated with silicic sedimentary rocks. The limestone is dense, finely crystalline, and nonfossiliferous. Stratigraphically, coarse-grained conglomerates are absent above the limestone (Orr, 1974).

Thin-Section Description. Three thin sections were made from the rock samples collected in the Skyline Drive area. The main components of the Fountain Formation are the metamorphic rock fragments (3-45%), polycrystalline quartz (1-28.5%), and monocrystalline quartz (5-27%). Biotite and muscovite are present in small amounts and calcite in trace amounts. The polycrystalline quartz grains are generally composed of three or four single quartz crystals. Much of the biotite in the rock fragments was decomposed to some extent, giving rise to a certain amount of hematite. Also, some biotite was bent by strong mechanical compaction. The formation is poorly sorted, immature, and loosely cemented.

Diagenesis. Figure 22 shows the following diagenetic events: after being buried, the arkosic clastics were intensively compacted, leading to the bending of mica (Figure 23) and fracturing of feldspar and quartz. Then, the dissolution of the matrix occurred in an alkaline environment (Figure 24). At the same time or possible later, the first generation of calcite was precipitated. Thereafter, a small amount of hematite formed in an oxidizing environment before the precipitation of second-generation calcite. When the environment became acidic, kaolinite formed, and both the first and second generations of calcite were dissolved to some extent. The last event was the precipitation of microcrystalline silica.

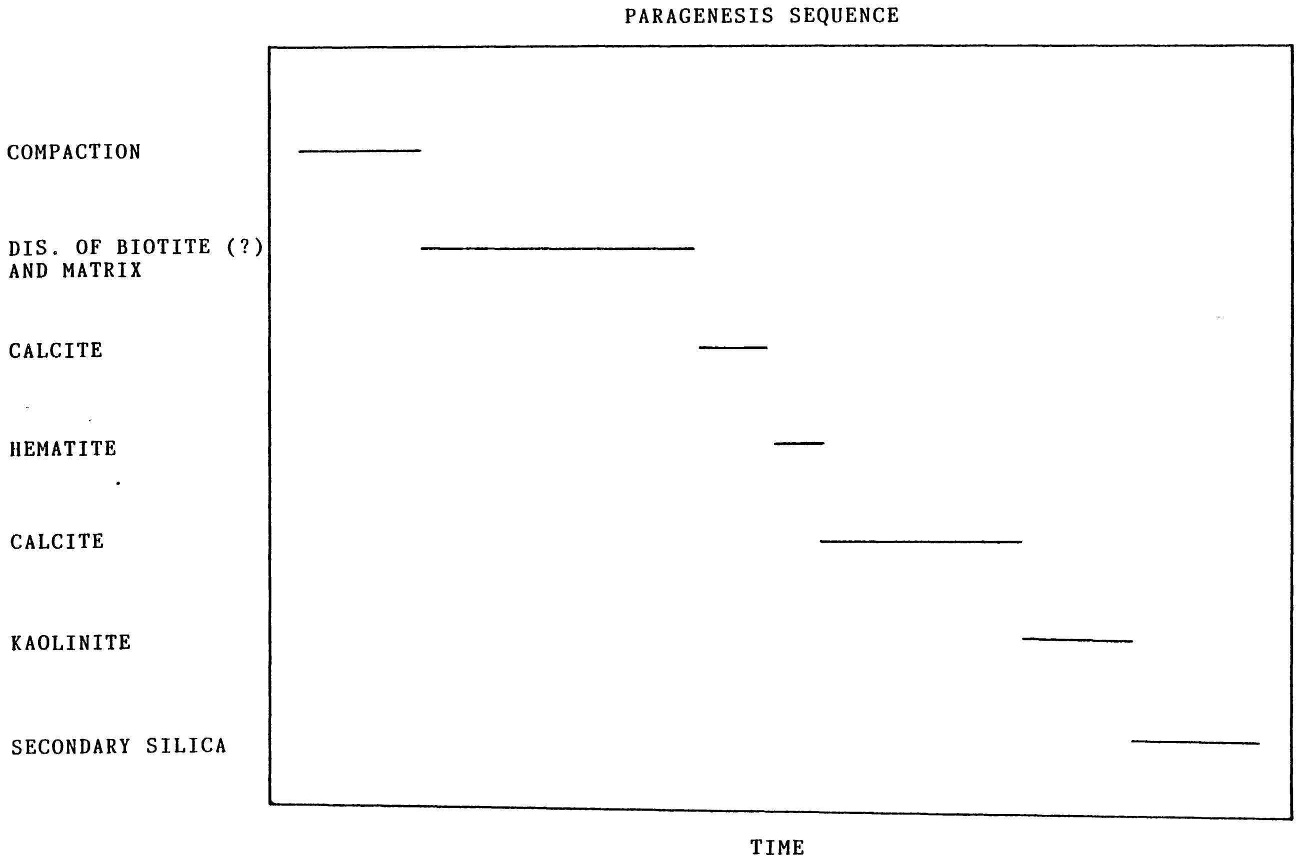


Figure 22. Diagenesis of the Fountain Formation (Skyline Drive Area)

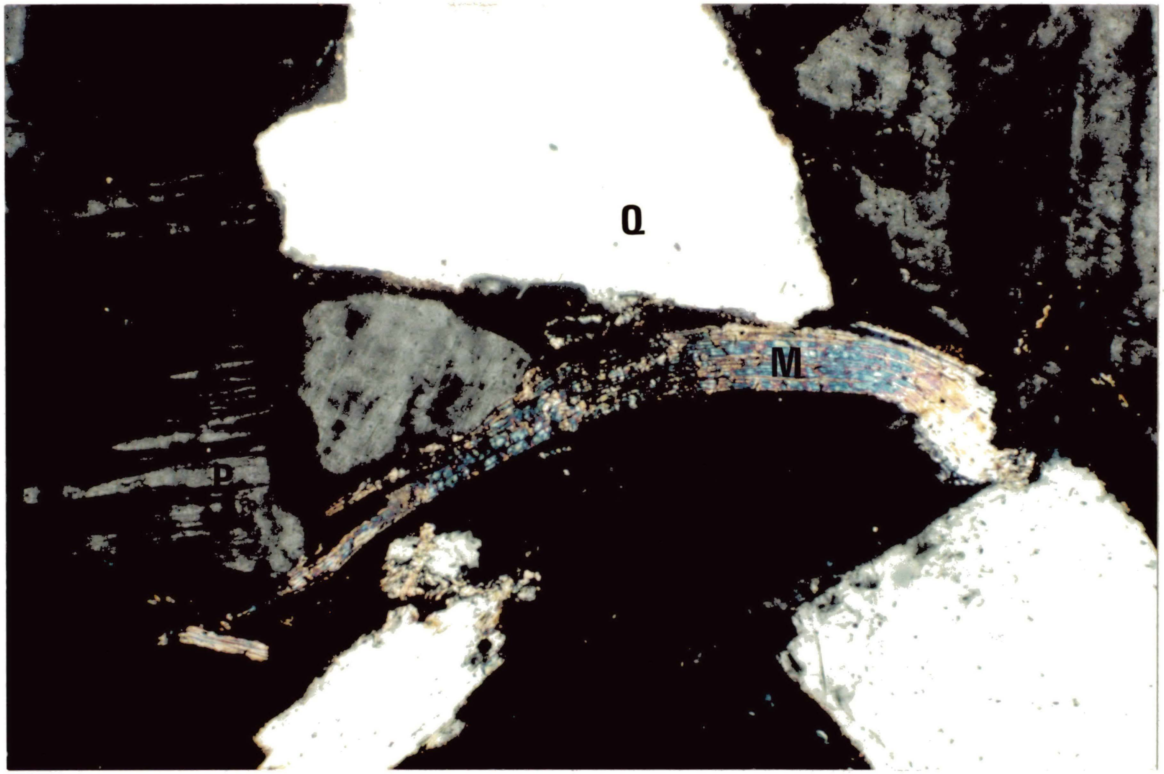


Figure 23. Quartz (Q), Plagioclase (P), and Bent Mica (M) in the Fountain Formation, 10 x 10 (+) (Skyline Drive Area)

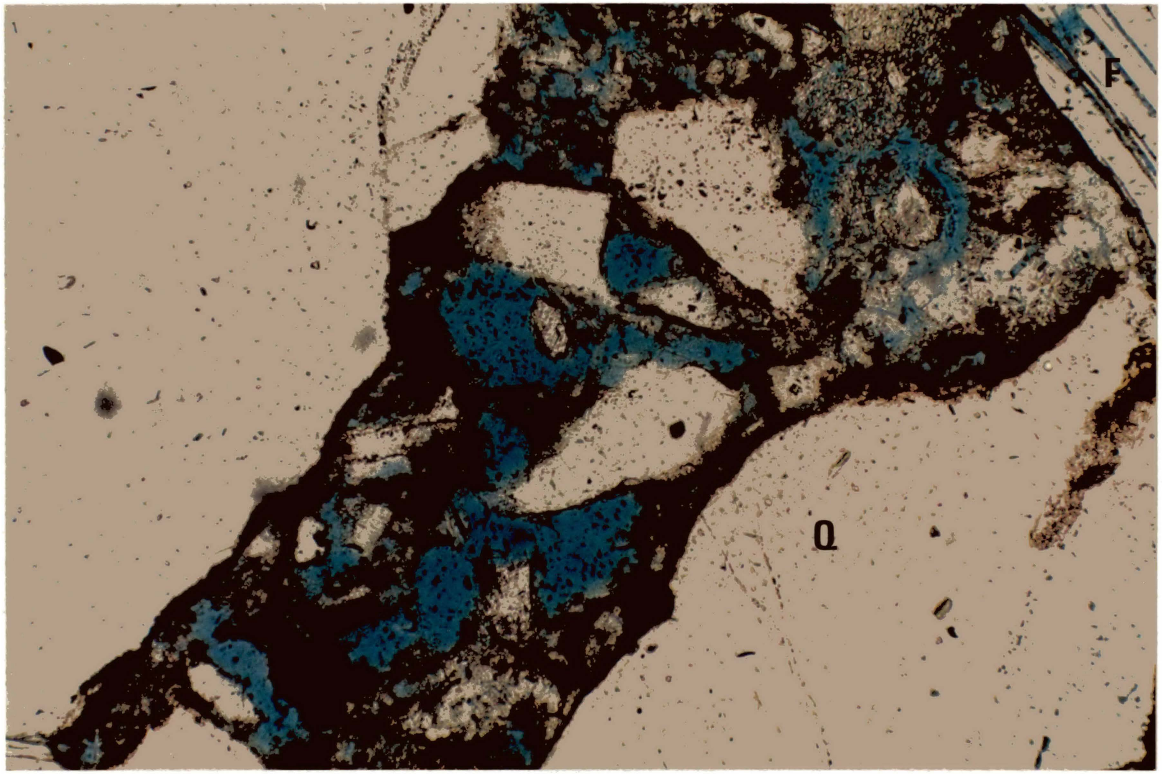


Figure 24. Dissolution of Matrix in the Fountain Formation, Q: Quartz, F: Feldspar (Skyline Drive Area)

Age. Few significant fossils have been found in the Fountain Formation, making it difficult to determine the Fountain's exact age. Poorly preserved Upper Pennsylvanian (Stephanian) plants have been found by Cloyd and Donovan (Donovan, personal communication, 1986). However, the subsurface correlations by Maher (1953) and Mitchell (1954) indicate that the age of the formation ranges from Morrowan to Virgilian times (complete Pennsylvanian).

Depositional Environment. According to Hubert (1960), the deposition of the Fountain Formation was preceded by a long period of erosion, and extensive weathering generated a pre-Fountain lateritic regolith. Near the end of Morrowan and Atokan time, major uplifting accompanied by faulting took place around the positions of the present Front Range and Wet Mountains, creating the Ancestral Rocky Mountains. During the following Atokan, Des Moinesian, and Missourian ages, the lower part of the Fountain, which is composed mainly of coarse arkosic clastics and conglomerates, was laid down along fault scarps and mountain flanks. The festoon-type cross-bedding, poor sorting, size of the large fragments, and rapid lateral variations of lithology suggest that the lower part was deposited in the fan-head and mid-fan subenvironments of a series of coalescing alluvial fans (Howard, 1966). The conglomerates and coarse sandstones possibly represent debris and/or fluvial channel deposits while the finer-grained materials represent overbank deposits.

During the Missourian and Virgilian times, when the upper Fountain was formed, the steep gradients created by the tectonic movement at the end of Morrowan and Atokan time were possibly so worn down that fewer clastics of coarse size were produced and the slopes of the fans were much smaller, resulting in the development of a much lower energy environment. The fluvial streams on the low gradient floodplain deposited fine sediments, and frequent overbank flooding and/or sheet floods in a distal fan subenvironment laid down siltstones. Near the end of the deposition of the Fountain Formation, some fresh-water lakes developed and deposited lenticular limestone beds. An alternative explanation is that the limestones are of pedogenic origin and formed as calcretes (caliche) (Donovan, personal communication, 1986).

According to Maher (1953), seas invaded the Canon City embayment during Des Moinesian, Missourian, and Virgilian ages. However, no evidences of these invasions were found in the studied area.

Lykins Formation

Outcrop and Stratigraphic Relationships. The Lykins Formation crops out in the Canon city embayment with the exception of the following three areas: (1) Garden Park, where the formation was completely removed by erosion; (2) the area between Eight Mile Creek and Oil Creek, where the exposure is poor due to post-Lykins, pre-Ralston erosion;

and (3) the area south of the Arkansas River, where the Lykins is absent due to the combined factors of truncation and overlap (Frederickson et al., 1956).

The Lykins Formation is unconformably overlain by the Ralston Creek Formation.

Lithology. The Lykins Formation is composed primarily of reddish-orange siltstones and shales. Near the base of the formation, thin gypsum layers occur in a few locations (Frederickson et al., 1956). Two persistent limestones are present in the upper part of the Lykins which Heaton (1933) named the "crinkled" limestones.

The formation is divided into the following five units from bottom to top: lower siltstone, lower crinkly laminated dolomitic limestone, middle siltstone, upper crinkly laminated limestone, and upper siltstone. The lower siltstone is orangish-red to reddish-brown, thinly bedded, shaly, and slightly calcareous. The lower crinkly laminated limestone is gray, finely crystalline, silty, and gypsiferous; its top and bottom surfaces are irregular, and its convex features are algal structures (laterally-linked stromatolites). The middle siltstone is similar to the lower siltstone lithologically. The upper crinkly laminated limestone is white to gray and moderately crystalline. The upper siltstone consists of interbedded red to pink mottled gray, thinly bedded, and gypsiferous siltstones.

The thickness of the formation varies from 0 to 190 feet in the embayment.

Thin-Section Description. Three thin sections were made from the rock samples collected from the upper Beaver Creek area. In the Lykins Formation, the limestone is silty, finely to moderately crystalline, and has a chemical origin. The siltstone is composed primarily of monocrystalline quartz (43-93.5%) polycrystalline quartz (1-2%), plagioclase and carbonate rock fragments. The polycrystalline quartz grains are composed of two or three single quartz grains and are possibly from both granite and metamorphic rocks. The presence of carbonate rock fragments indicates that the source rocks of the Lykins Formation probably include Ordovician limestone and dolomite. Figure 25 shows the variations of the detrital components. It is evident that at the end of the deposition of the Lykins Formation, tectonic movement became stronger than before, because more polycrystalline quartz and less monocrystalline quartz was generated. The Lykins Siltstone is cemented by both calcite and siderite. Figure 26 shows the variations of carbonates and matrix. The patterns of the variations indicate that the formation of carbonates (calcite and siderite) was controlled by permeability of the rock which was, in turn, controlled by the amount of matrix.

Diagenesis. The first diagenetic event is the formation of the quartz overgrowths, at which time the formation solution was acidic and rich in silica. Then some matrix was dissolved, giving out some ferrous iron. Later, siderite was precipitated by the combination of the ferrous

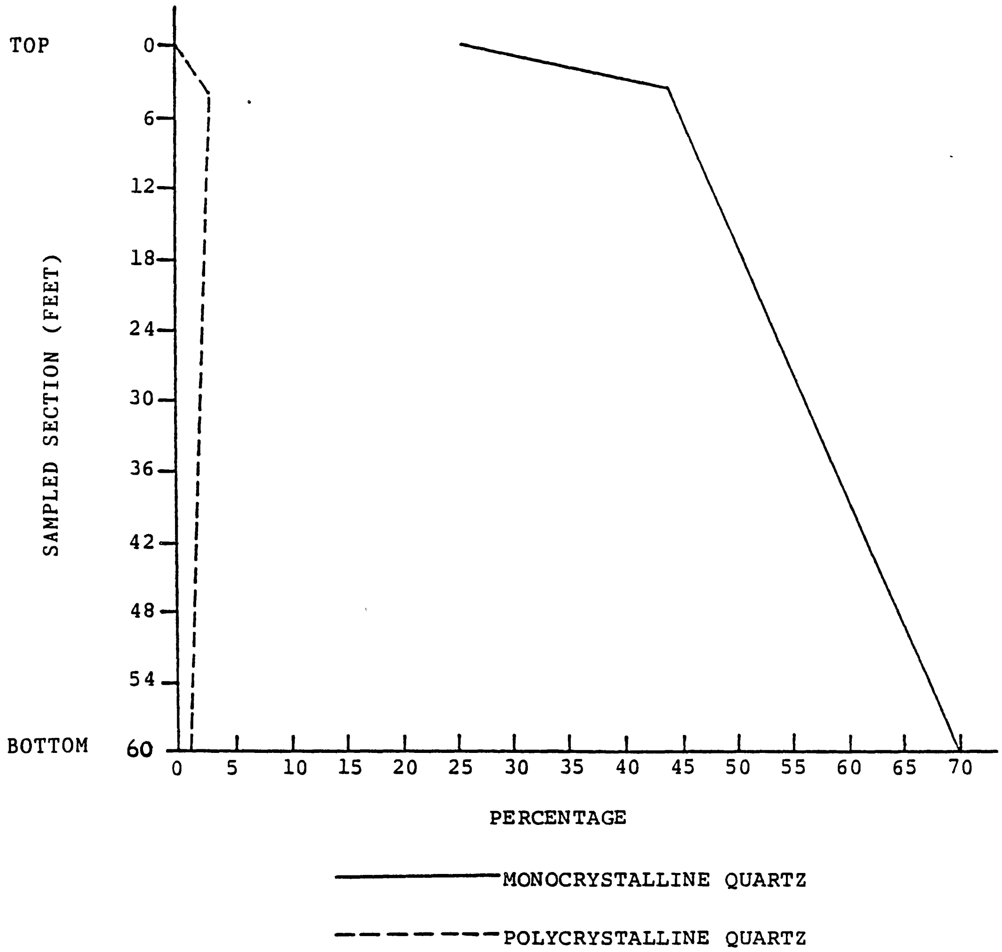


Figure 25. Vertical Variations of Monocrystalline Quartz and Polycrystalline Quartz in the Sampled Section of the Lykins Formation (Upper Beaver Creek Area)

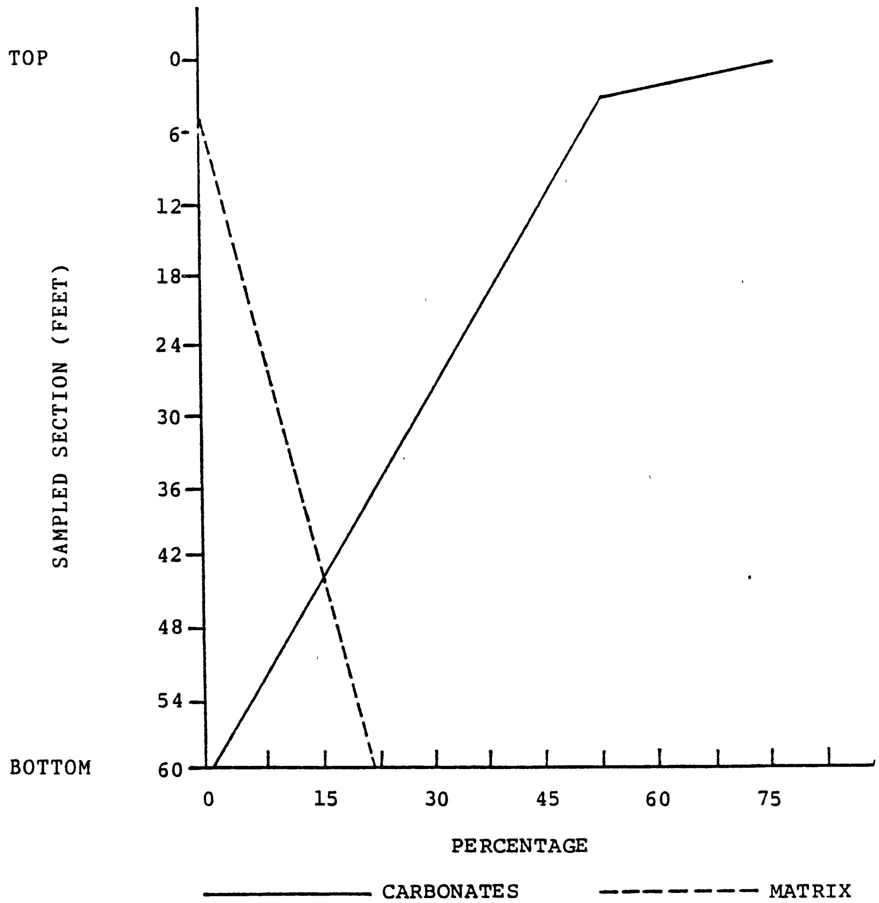


Figure 26. Vertical Variations of Carbonates and Matrix in the Sampled Section of the Lykins Formation (Upper Beaver Creek Area)

cations and carbonate anions. At the same time, secondary calcite formed. Last, the dissolution of calcite and siderite created some secondary porosity. Figure 27 shows the above sequence.

Age. Heaton (1933) and Dorrell (1940) assigned the sediments above the upper crinkled limestone to the Triassic and the beds below the top of the crinkled limestone to the Permian. Girty's and Butters' (1912) paleontological study also indicated a Permian age for fossils from the crinkled limestones of the Lykins Formation. However, Maher (1953) and Mitchell (1954) put the Permian-Triassic boundary higher, within the upper siltstone.

Heaton (1933) correlated the crinkled limestone with the San Andreas Limestone in New Mexico, and Maher (1953) correlated the upper and lower limestone units with the Day Creek Dolomite and Blaine Formation, respectively, of Western Kansas.

Depositional Environment. The interbedded limestones with terrigenous clastic sediments, crinkle laminations, scarcity of marine fossils, and the association of gypsum suggest a playa or lacustrine environment. The siltstones were deposited when there was a great influx of detrital materials, and the limestones and gypsum were precipitated when the input of detritus was small and water was clear. The crinkled laminations and the circular and convex structures might be generated by some form of algae. The lakes

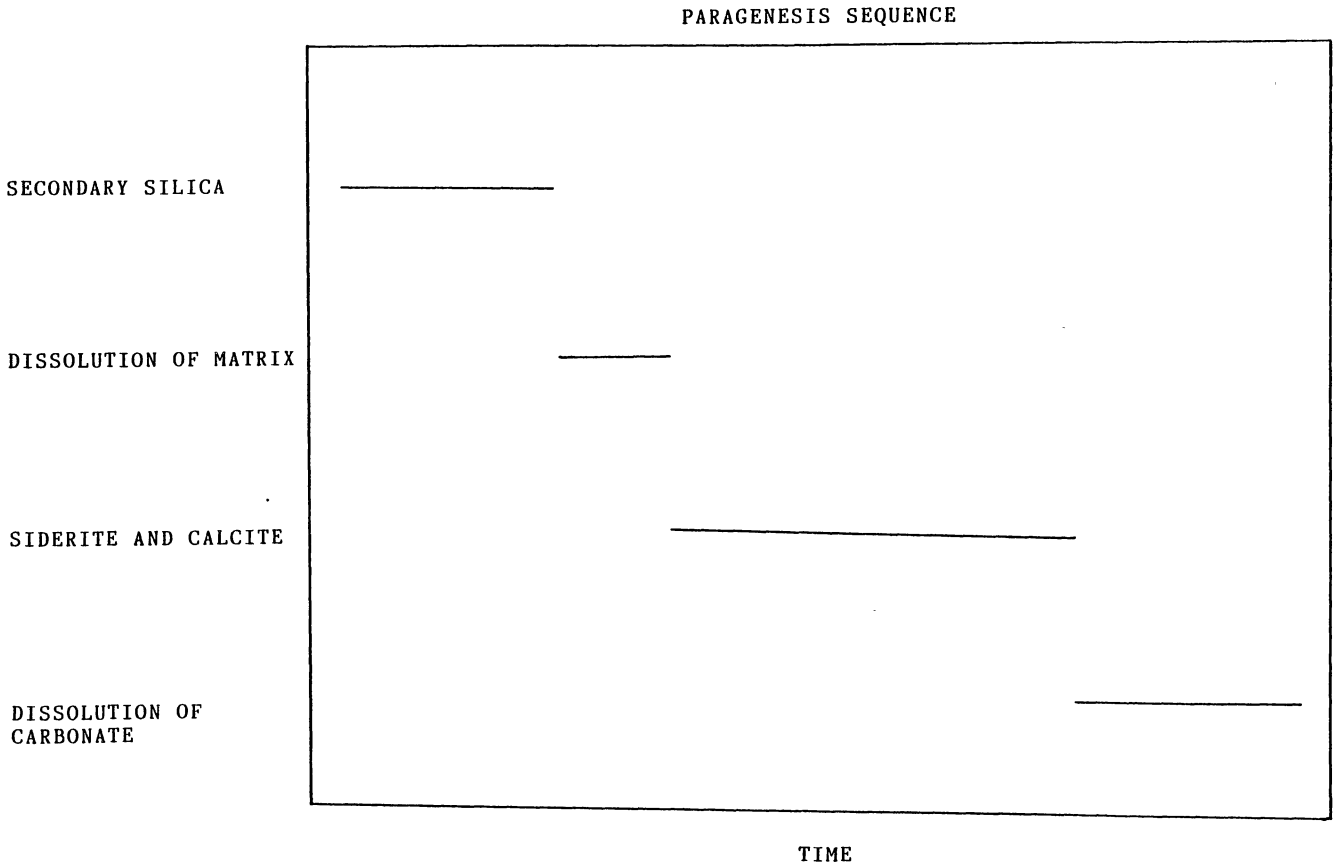


Figure 27. Diagenesis of the Lykins Formation (Upper Beaver Creek Area)

developed possibly on a flat coastal plain or a flat alluvial fan. Figure 28 shows that the relief before the Lykins was only about 150 feet.

The gypsiferous beds suggest an arid climate, shallow water, restricted circulation and high evaporation rate; the small grain sizes and absence of cross-bedding indicate a low energy environment.

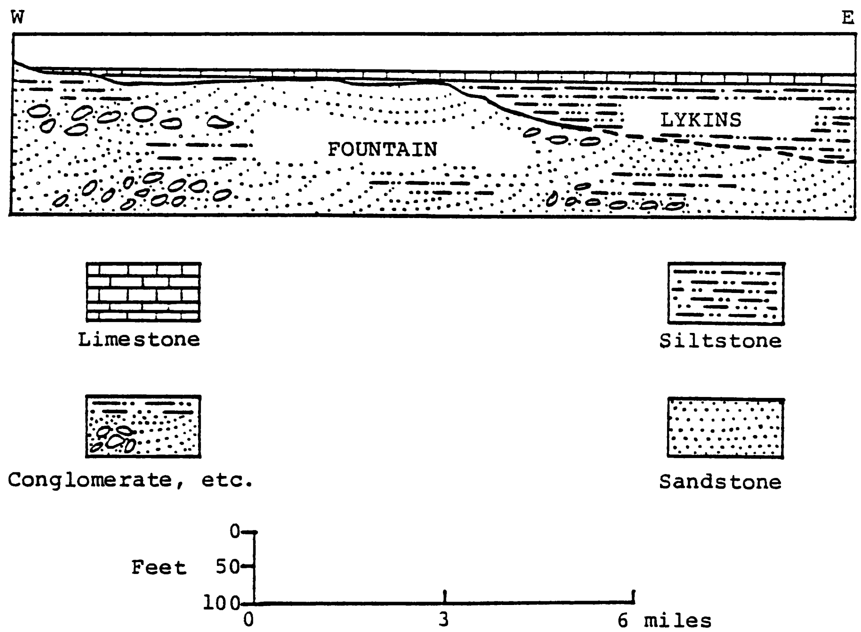


Figure 28. West-East Cross Section of the Interval Between the Lower Crinkled Limestone Unit and the Base of the Lykins in the Canon City Embayment (After Frederickson, Delay, and Saylor, 1956)

CHAPTER VI

MESOZOIC ERA

Geological Setting

During the Triassic time, the Ancestral Rocky Mountains region was a positive landmass; however, the positive landmass had a low relief and furnished only fine-grained detritus from the Lykins Formation to the northwestern part of Colorado, which was covered by sea. This erosion of the Lykins Formation went on until Late Jurassic when the Ralston Creek Formation was deposited in a continental-inland sebkha. According to Frederickson et al. (1956), a minor uplift in the area west of Canon City, possibly in the Wet Mountains area, was coincident with the gentle subsidence to the east of the studied area. The subsiding process created a big lake which dried up later on in an arid climate. However, during the following Morrison time, a warm moist climate replaced the previous arid one, resulting in the development of fresh water streams, lakes, and swamps in which variegated marls, shales, limestones, and sandstones were deposited. This formation contains fresh-water shells, the remains of land mammals, and the bones of gigantic dinosaurs that lived in the swamps and streams and on the shores of the shallow temporary lakes. According to Haum and Kent (1965), a great portion of the coarse clastic sediments was delivered by the streams running from the mountains further to the west in the present state of Nevada.

If the Morrison is confined entirely to the Jurassic, then the drying-up of the Morrison lakes was succeeded by land situations in the Ancestral Rocky Mountains region throughout the early part of the Early Cretaceous time. During the early Early Cretaceous, the base-leveling of the Ancestral Rockies was completed. A later subsidence at the close of late Early Cretaceous, possibly together with a rise of the sea level, had progressed far enough to allow the sea water from the Gulf of Mexico and the Arctic to invade the Western Interior as far as the present mountain front and possibly much farther. Thereafter, the Dakota Group was deposited along the sea coast.

Continued subsidence, possibly accompanied by the continued rise of sea level, lead to the formation of the broad Western Interior basic during Upper Cretaceous time. In the basin, sands, muds, and carbonate ooze were deposited and later lithified into the marine formations of Upper Cretaceous age. Along the western margin of the basin, beds of peat eventually formed the upper Cretaceous coals. Molenaar (1983) recognized five major transgressive-regressive cycles in the upper Cretaceous time. The initial transgression (Greenhorn) was the most extensive throughout the Rocky

Mountains region. Each succeeding cycle failed to transgress as far landward as the preceding one, suggesting an overall filling of the Western Interior. The withdrawal of sea water was accompanied by a differential movement of the rocks (Laramide Revolution). The revolution lifted the Rocky Mountains region, and the resulting erosion created the post-Vermejo unconformity.

In the Canon City embayment, open sea conditions prevailed until after the middle of the Montana Epoch (81 m.y. to 65 m.y.) By this time, the filling of the interior with sediments derived from the west and northwest had proceeded to such an extent that a deltaic environment was established. The upper portion of the Pierre Shale was deposited in the pro-delta, the Trinidad Sandstone as delta-front, and the Vermejo Formation on the delta plain.

Petrology and Depositional Environment

Ralston Creek Formation

Outcrop and Stratigraphic Relationships. The Ralston Creek Formation crops out continuously from west to east in the embayment. In the eastern part of the studied area, the contact between this formation with the overlying Morrison Formation is conformable; the two formations are difficult to be differentiated (Brady, 1969). Frederickson et al. (1956) and Brady (1969) suggested that the absence of gypsum, color change from pastels to drabber colors, a great increase in carbonate content, and the occurrence of the well-cemented sandstones identify the Morrison sediments. However, in the central and western parts where the conglomeratic Ralston Creek beds are overlapped by the Morrison beds of fine sandstone and claystone, the two formations are in disconformable contact.

Lithology. The formation consists of interbedded gypsum, limestone, siltstone and sandstone. Gypsum beds occupy a large portion of the sequence and contain abundant detrital sediments. The gypsum is white, orange-red, weathered gray and finely to coarsely crystalline, and it occurs in both thinly laminated and nodular beds which are usually jointed. The siltstone is tan to brown, light green to gray, and thinly bedded. Sandstone units occur primarily in the lower Ralston Creek Formation. They are gray, olive-green to tan, moderately sorted and laminated. Some detrital grains reach the conglomerate size, and they are angular to sub-angular.

Thin-Section Description The sandstone samples from a gypsum quarry in the Indian Spring area are composed primarily of monocrystalline quartz (37.5-71%), polycrystalline quartz (0-15.5%), and feldspar (1-13%). The bottom sample contains a high amount (15.5%) of polycrystalline quartz derived from metamorphic rocks, indicating that the Indian Spring area is near the source rocks. A few grains of feldspar were observed; microcline and albite twins are common. The rock is cemented by gypsum (Figures 29 and 30), calcite, and quartz overgrowths. However, a significant

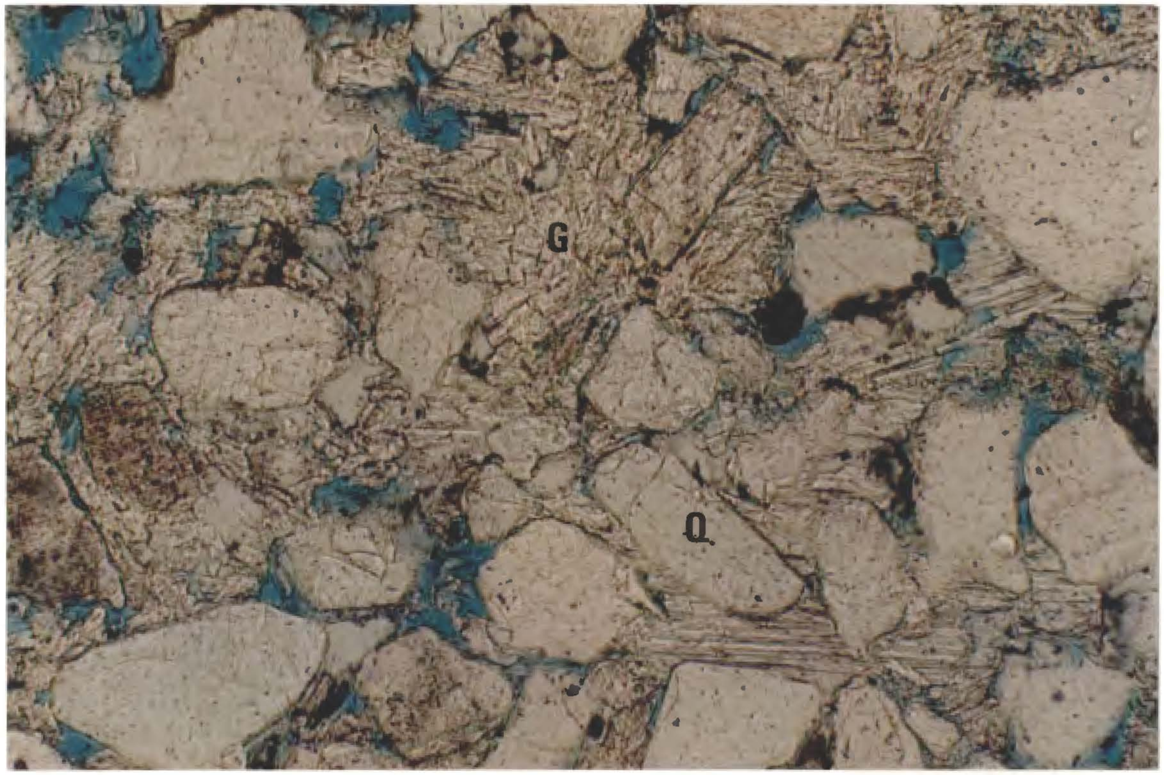


Figure 29. Gypsum (G) and Quartz (Q) of the Ralston Creek Formation, 10 x 10 (-) (Indian Spring Area)

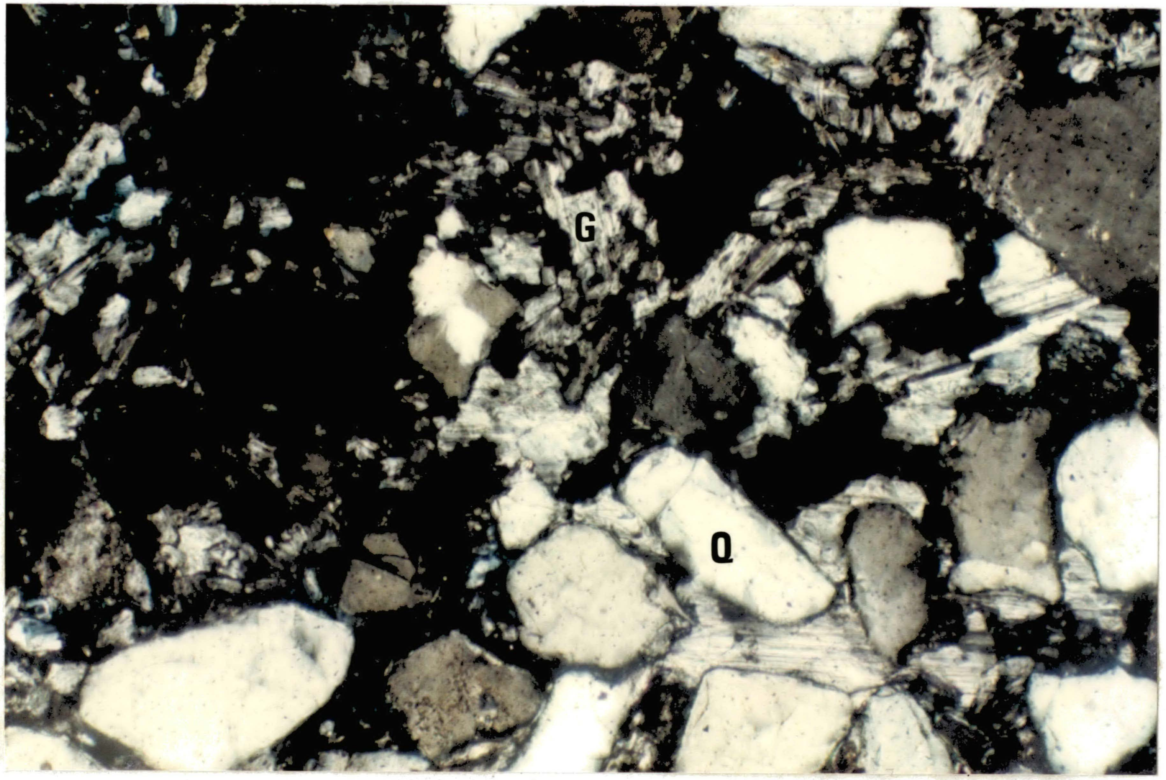


Figure 30. Gypsum (G) and Quartz (Q) of the Ralston Creek Formation, 10 x 10 (+) (Indian Spring Area)

amount of gypsum was removed by dissolution processes, leading to 12-15.5% secondary porosity. One sample from the middle part of the Ralston Creek in Skyline Drive is heavily cemented by calcite (47%) and has no gypsum. The absence of gypsum indicates that the Skyline Drive area was in a different sedimentary environment than the Indian Spring area, where a large amount of gypsum precipitated during the deposition of the Ralston Creek Formation.

Diagenesis. The diagenetic sequence of the samples is shown in Figure 31. The rocks were originally cemented by calcite, which replaced a small amount of quartz and feldspar. Later on, formation water became acidic and the calcite was dissolved. When the concentration of $SO^{4=}$ became very high, the following chemical reaction took place:

$$CaCO_3 + SO_4^{=} \rightleftharpoons CaSo_4 + CO_3^{=}$$

That is, calcite was replaced by gypsum. Then, the dissolution of gypsum created a significant amount of porosity (Figures 29 and 30).

The sample from the Skyline Drive area experienced the following three diagenetic stages: (1) the dissolution of feldspar in an alkaline environment; (2) the precipitation of calcite; and (3) the dissolution of the calcite and the generation of secondary porosity.

Age. Fossils reported in the formation are fossil fish (Schoewe 1930; Dunkle 1942), a single gymnosperm, and pelecypods (Imlay, 1952). These fossils date from the formation as Jurassic.

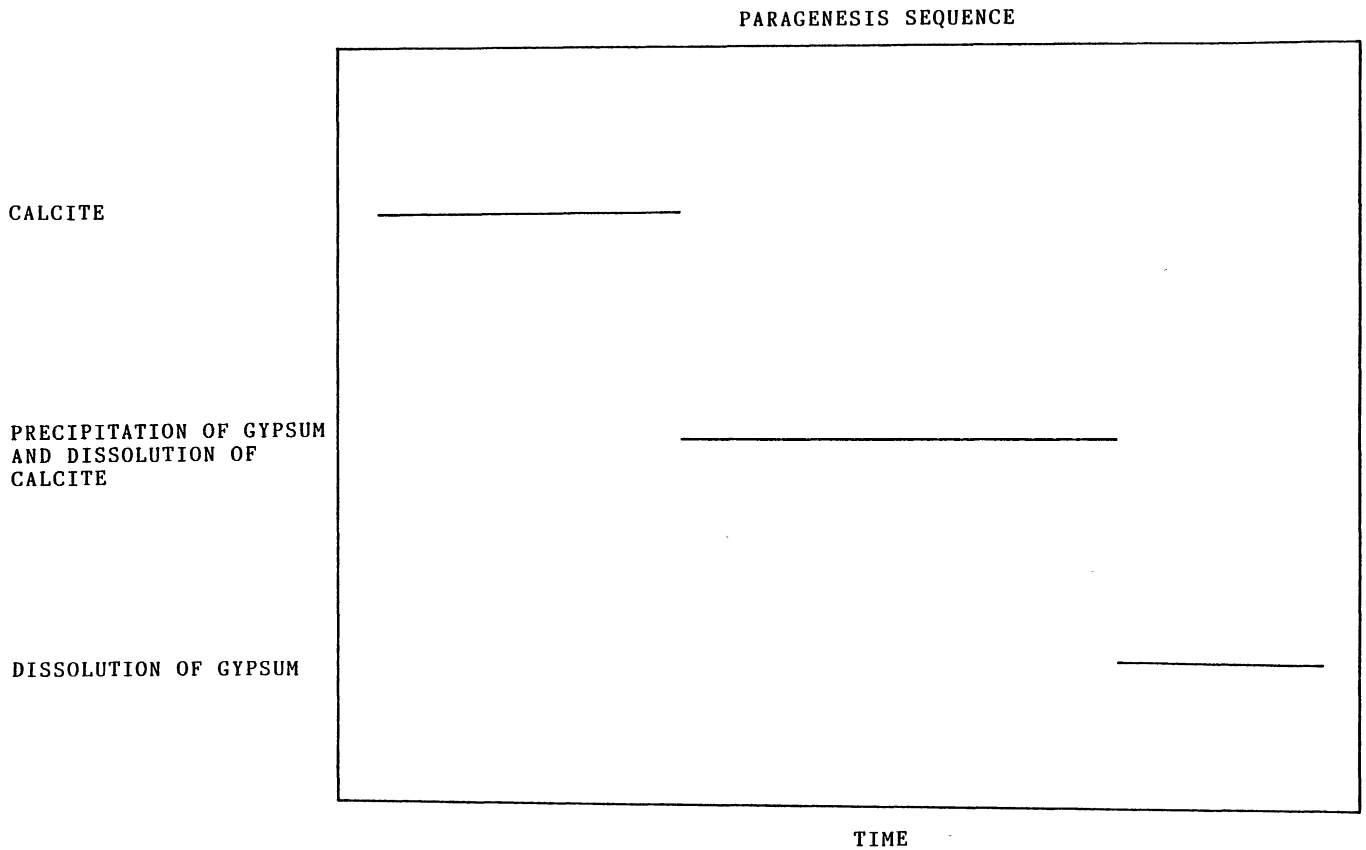


Figure 31. Diagenesis of the Ralston Creek Formation (Indian Spring Area)

Imlay (1952) correlated the Ralston with the Swift Formation of Montana, the Summerville Formation or the Curtis Formation of northern Colorado. He concluded that the formation is of Late Jurassic age.

Depositional Environment. The erosion of the Lykins and the Fountain Formation began in the Late Permian Period and continued through the Triassic and the Early-Middle Jurassic time. Until the deposition of the Ralston Creek Formation, the relief of the land surface was less than 100 feet (Figure 32). The two crinkled limestones in the Lykins Formation were resistant to erosion and controlled the development of the pre-Ralston topography, and erosion progressed rapidly once the limestone layers were cut through.

Frederickson et al. (1956) determined three facies of the Ralston Creek Formation (Figure 33). The first is the conglomerate facies in the western part of the studied area from the Six-Mile Park to the Grape Creek area. The conglomerate was deposited in a contential environment, and during the deposition the northward-trending stream divide near Garden Park served as a barrier to trap the coarse sediments which were derived from the site of the present Wet Mountains. The second is the sandstone facies which is interpreted as beach deposits. The third is the gypsum-shale facies of a sebkha origin. These three facies are intergradational from west to east. However, the very thinly-laminated limestone and gypsum, and the absence of traction sedimentary structures suggest that the sandstone

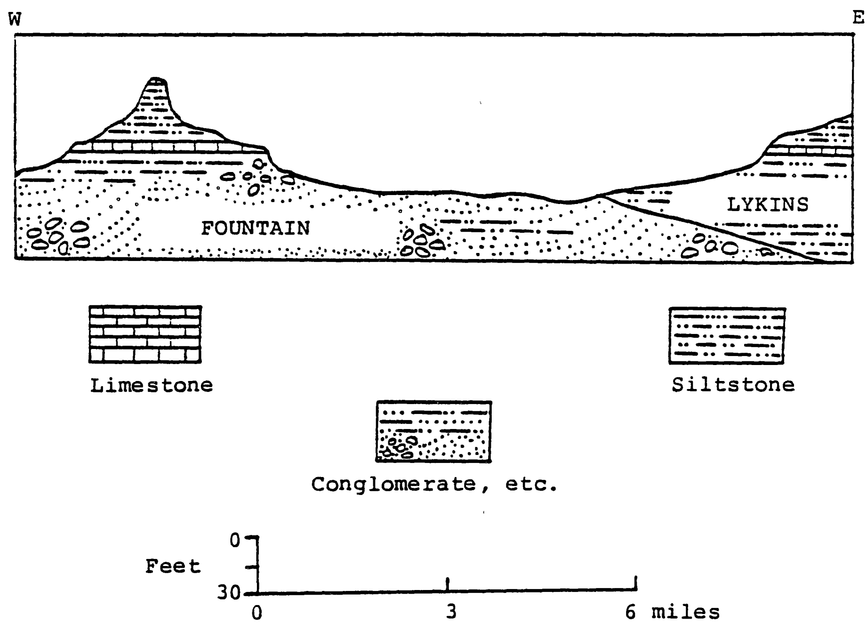


Figure 32. West-East Cross Section of the Interval Between the Lower Crinkled Limestone and the Base of the Ralston Creek in the Canon City Embayment (After Frederickson, Delay, and Saylor, 1956)

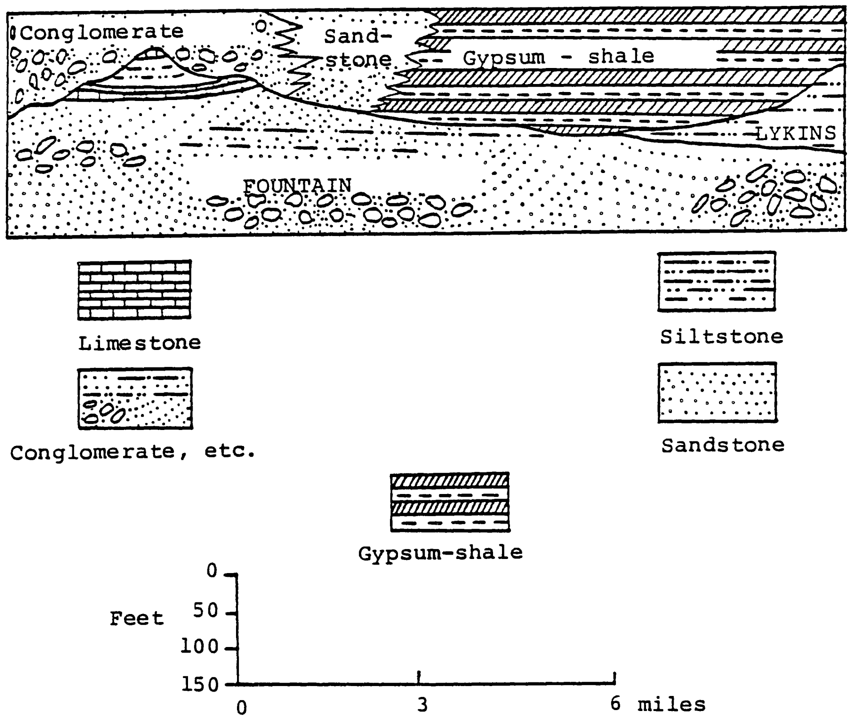


Figure 33. West-East Cross Section of the Ralston Creek Formation. The Variation of the Formation's Thickness is Controlled by the Pre-Ralston Topography (After Frederickson, Delay, and Saylor, 1956)

facies is a lake beach deposit and the gypsum-shale facies a shallow-water lacustrine deposit. The alternating sequence of gypsum and fine clastic beds reflects repeated evaporate deposition followed by an input of detrital materials. The cracking of the gypsum indicates the periodic drying of the lakes and an arid to semiarid climate.

Morrison Formation

Outcrop and Stratigraphic Relationship. In the Canon City embayment, the Morrison Formation crops out along hog-back and escarpment slopes. Mostly, it rests on older sedimentary rocks. However, it nonconformably lies on the Pikes Peak Granite and Idaho Spring Formation in the northern part of the studied area, and is disconformably overlain by the Lytle Sandstone of the Early Cretaceous age.

Lithology. The Morrison Formation is composed of interstratified claystones (71%), sandstones (20%), siltstones (7%), and carbonate rocks (2%) (Brady, 1969). The claystones are greenish, purple, and poorly indurated; the sandstones are greenish-gray, very fine- to medium-grained, cross-stratified, and lenticular; siltstones are greenish-gray to red, thin-bedded, and fissile; the limestones are gray to black, finely crystalline, dense, thinly laminated, and thin-bedded. The thickness of the Morrison Formation is about 350 feet.

Thin-Section Description. The samples from the OK Corral area, which are within a cross section recording the fining upward sequence of a fluvial channel, are classified into medium-fine grained quartzarenite and conglomerate litharenite. The quartzarenite contains a trace amount of feldspar and polycrystalline quartz. Both microcline and albite twins are common. One polycrystalline quartz grain is usually composed of three of four monocrystalline quartz crystals, and the contact between the crystals is skull-seam-shaped, indicating a metamorphic origin. The litharenite is comprised of chert and the fragments of quartzarenite and metamorphic rocks. Conglomerate is generally subrounded-rounded, and makes up about 60% of the litharenite. Both the quartzarenite and litharenite are cemented by microquartz and chalcedony. As shown in Figure 34, the variation of chert is not obvious; however, the variations of polycrystalline and monocrystalline quartz are dramatic and the two curves are contrary to each other in their amounts. This relationship might be caused by the change of the environmental energy. When the energy was high, the amount of unstable polycrystalline quartz became less and stable monocrystalline quartz became more, and vice versa. Figure 35 shows the variations of secondary minerals and matrix. Obviously the distribution of secondary minerals is controlled by stratification of the rocks. More permeable layers (with less matrix) have more secondary minerals than less permeable layers. Figure 36 suggests that there were three times when the water-flow of the

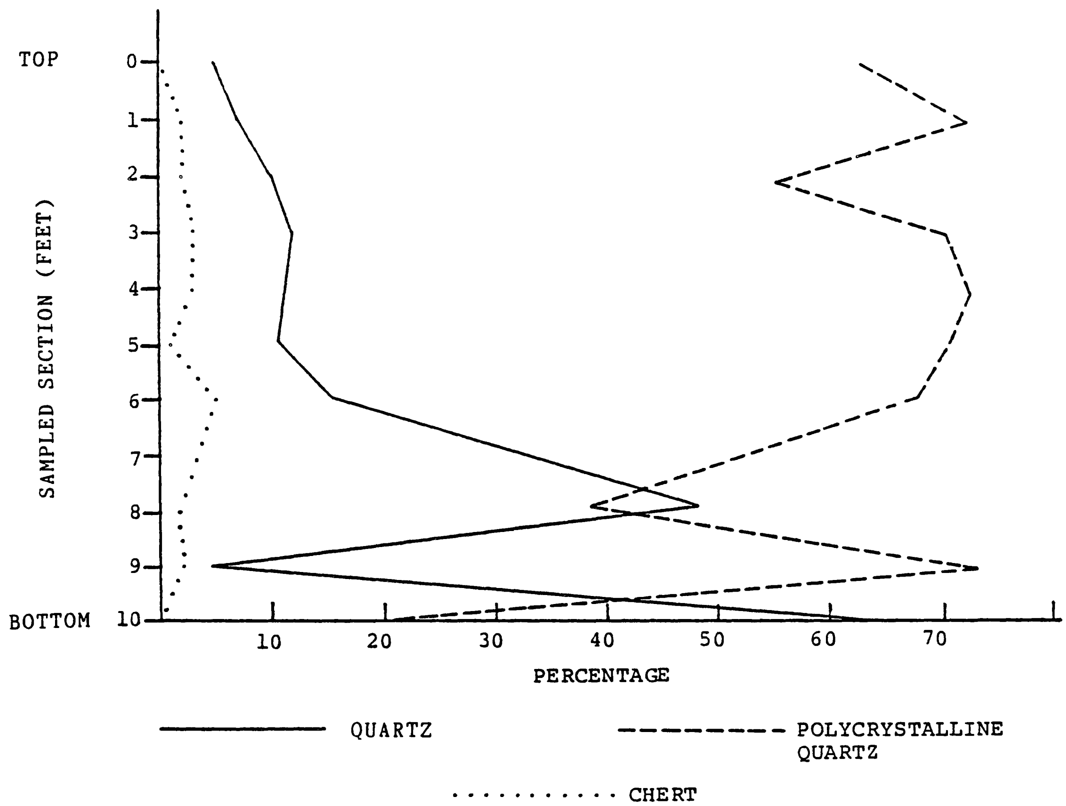


Figure 34. Vertical Variations of Detrital Components in the Sampled Section of a Meander (Morrison Formation, OK Corral Area)

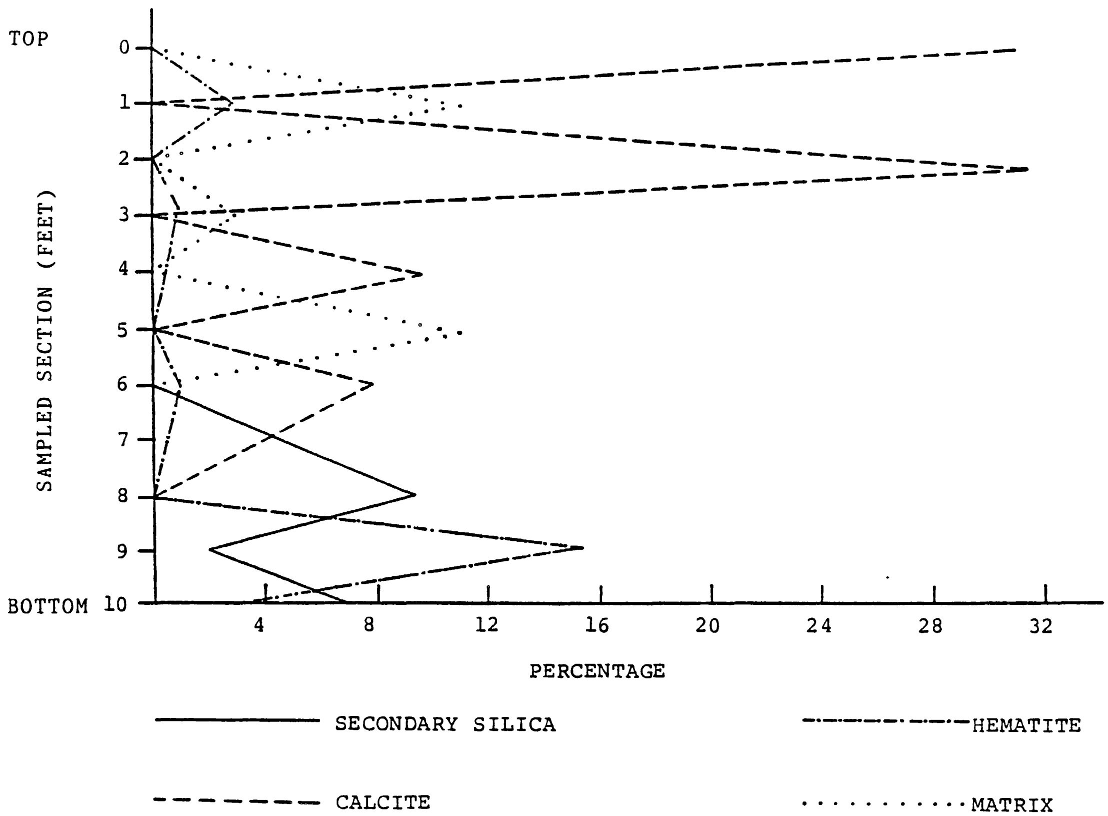


Figure 35. Vertical Variations of Secondary Minerals, Hematite, and Matrix in the Sampled Section of the Meander (Morrison Formation, OK Corral Area)

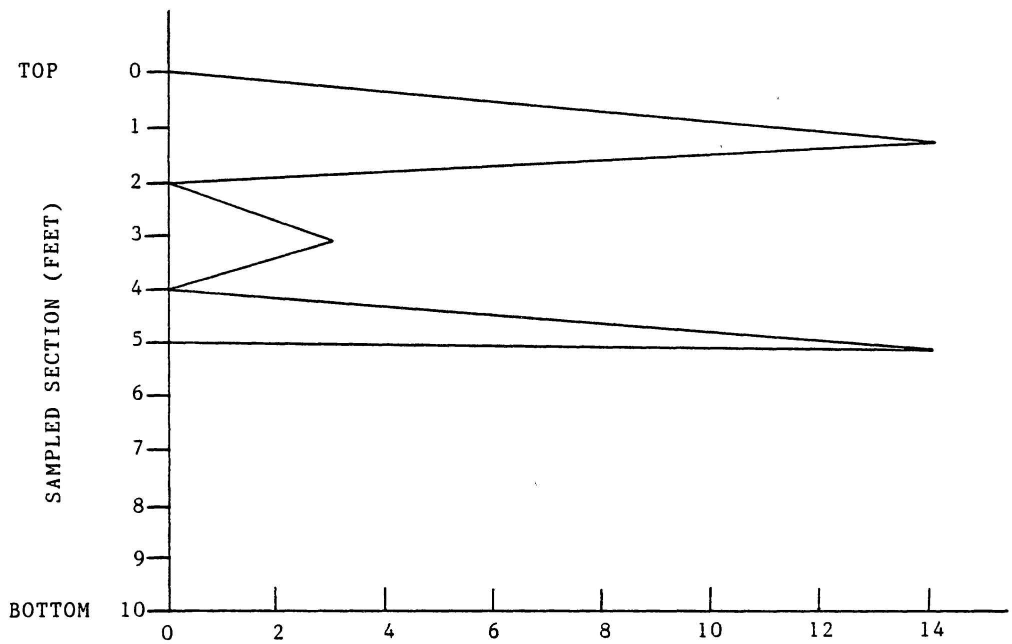


Figure 36. Vertical Variation of Matrix/(Monocrystalline Quartz + Polycrystalline Quartz + Chert) in the Sampled Section of the Meander (Morrison Formation, OK Corral Area)

meander was quiet. Figure 37 shows that the grain size variation follows a fining upward pattern, typical of channel deposition.

The samples from the Oklahoma Geological Camp area are classified as fine-medium sandstone quartzarenite. The percentage of monocrystalline quartz ranges from 0-90%; chert, polycrystalline quartz, feldspar (microcline and plagioclase), and carbonate rock fragments are in trace amounts. Polycrystalline quartz is from gneiss and/or granite. Both microcline and albite twins are common. Cementing minerals, which heavily cement the quartzarenite, are mainly secondary silica (1-36%) and calcite (0-75%). Figure 38 shows the variations of detrital fragments; quartz content increases upward while chert and polycrystalline quartz decreases. Figure 39 shows the variations of secondary minerals; calcite and secondary silica are counteractive against each other in their amounts. This situations was possibly caused by a variation of pH values which control the precipitation and dissolution of the two minerals. Figure 40 shows the variation of matrix/(quartz + chert). The interval with the ratio of 6% may be an overbank deposit (the top six feet of the sampled section). A fining upward sequence is shown in Figure 41.

Diagenesis. Figure 42 shows the diagenetic sequence of the samples from the OK Corral area. The first diagenetic event is the precipitation of quartz overgrowths. When the pH value of the formation solution became high, matrix and

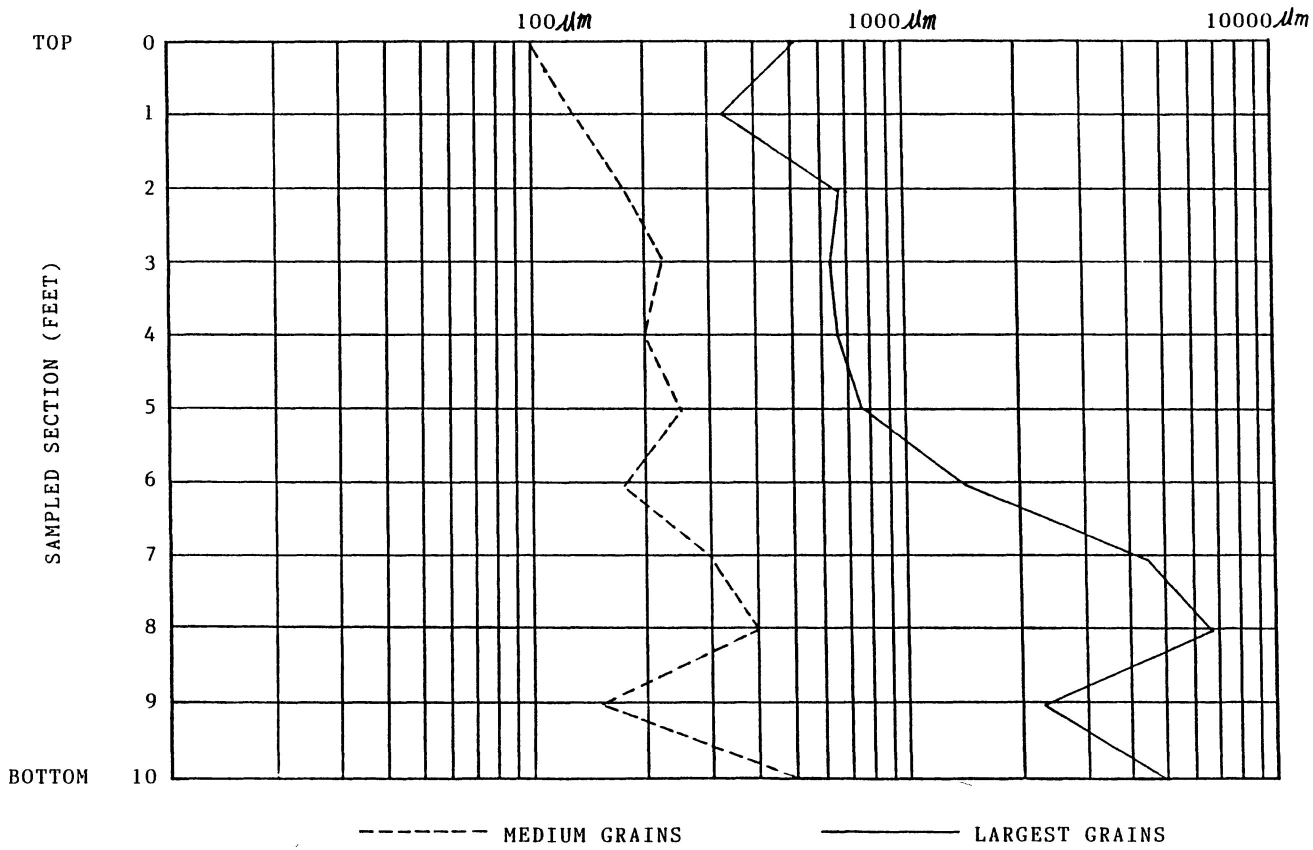


Figure 37. Vertical Variation of Grain Size in the Sampled Section of the Meander (Morrison Formation, OK Corral Area)

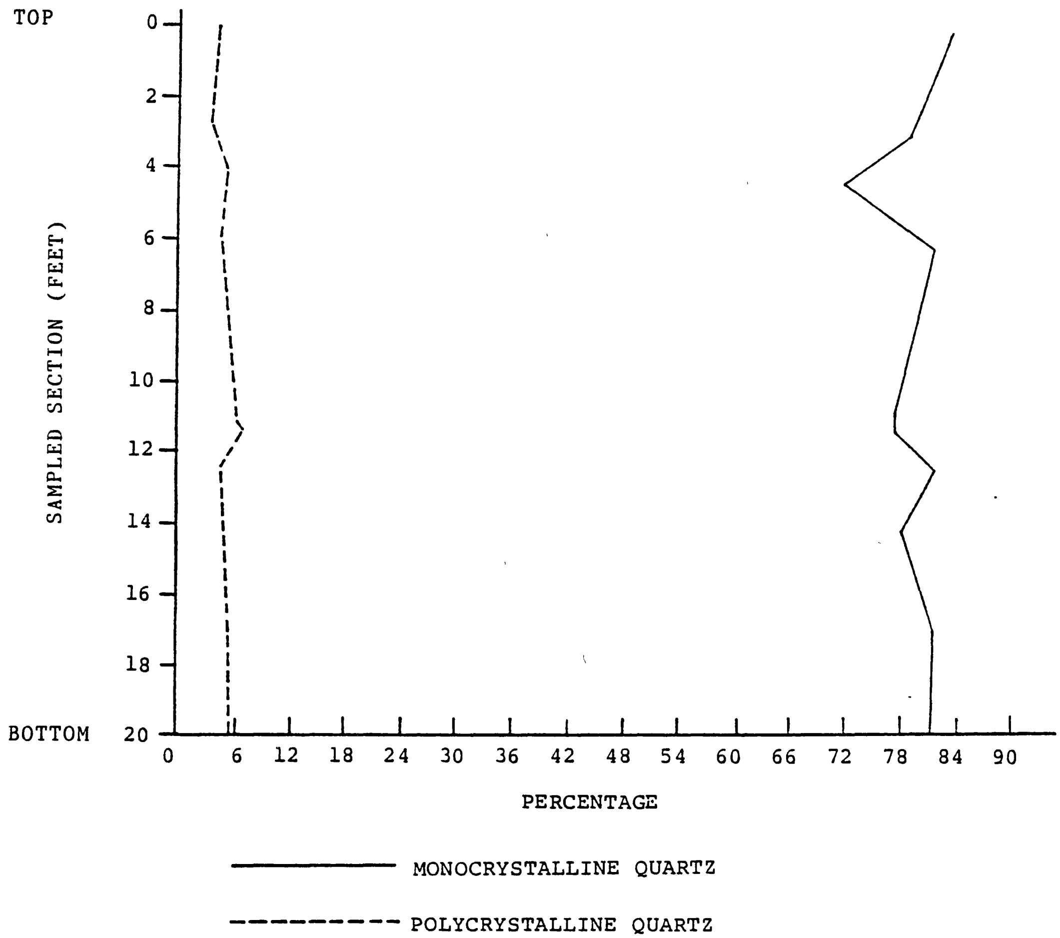


Figure 38. Vertical Variations of Detrital Components in the Sampled Section of the Morrison Formation (OK Geology Camp Area)

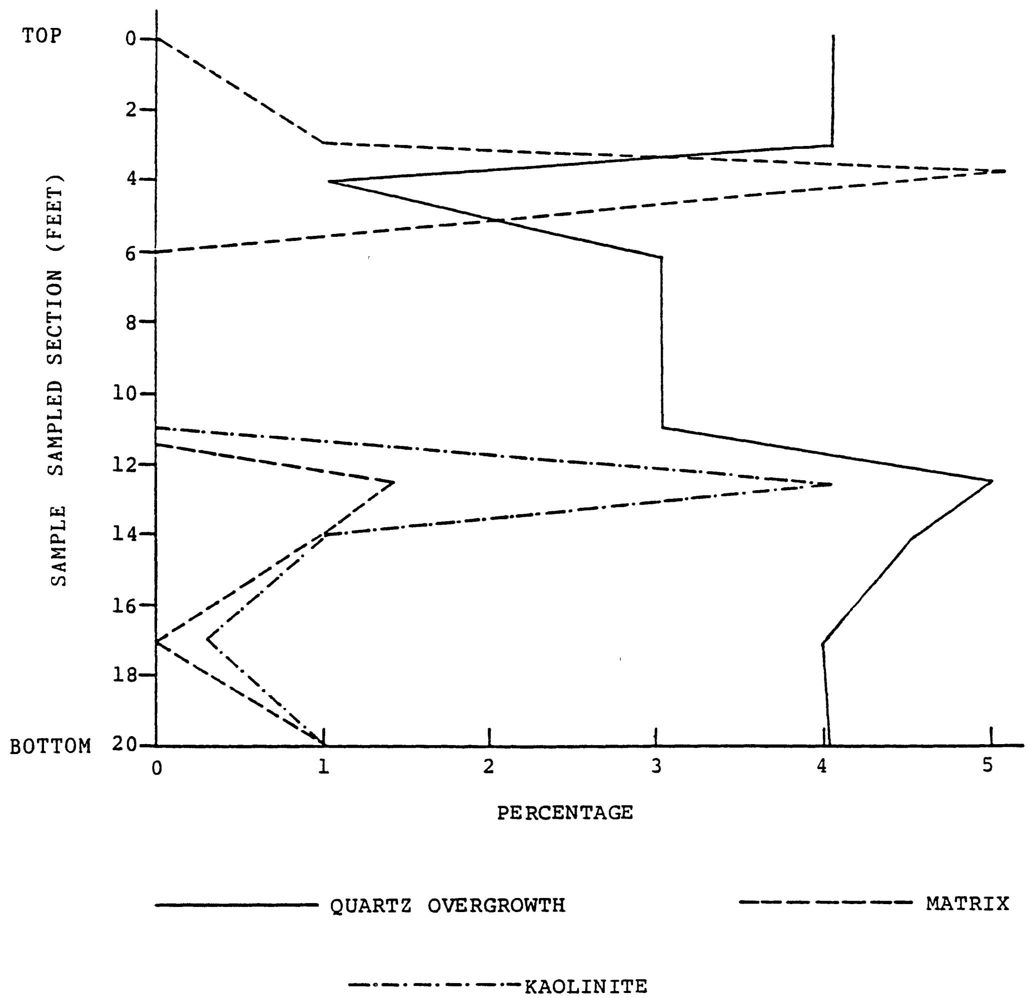


Figure 39. Vertical Variations of Secondary Minerals and Matrix in the Sampled Section of the Morrison Formation (OK Geology Camp Area)

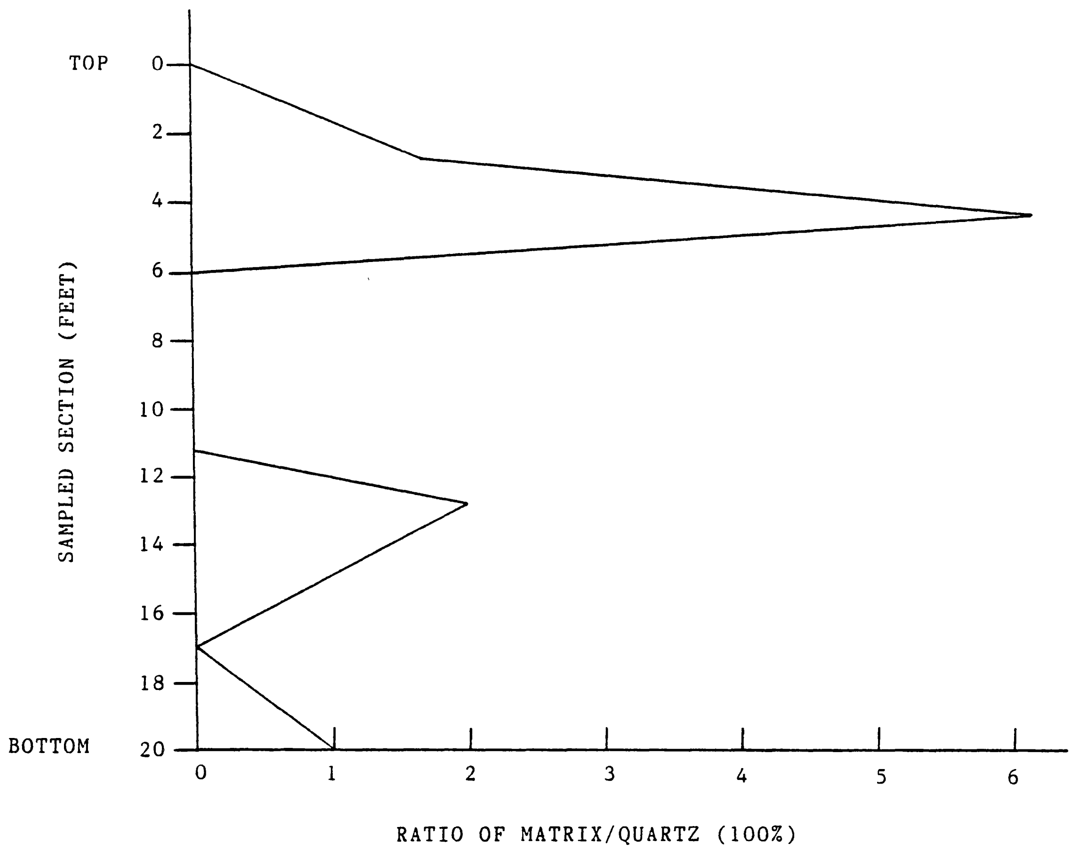


Figure 40. Vertical Variation of Matrix/(Monocrystalline Quartz + Polycrystalline Quartz) in the Sampled Section of the Morrison Formation (OK Geology Camp Area)

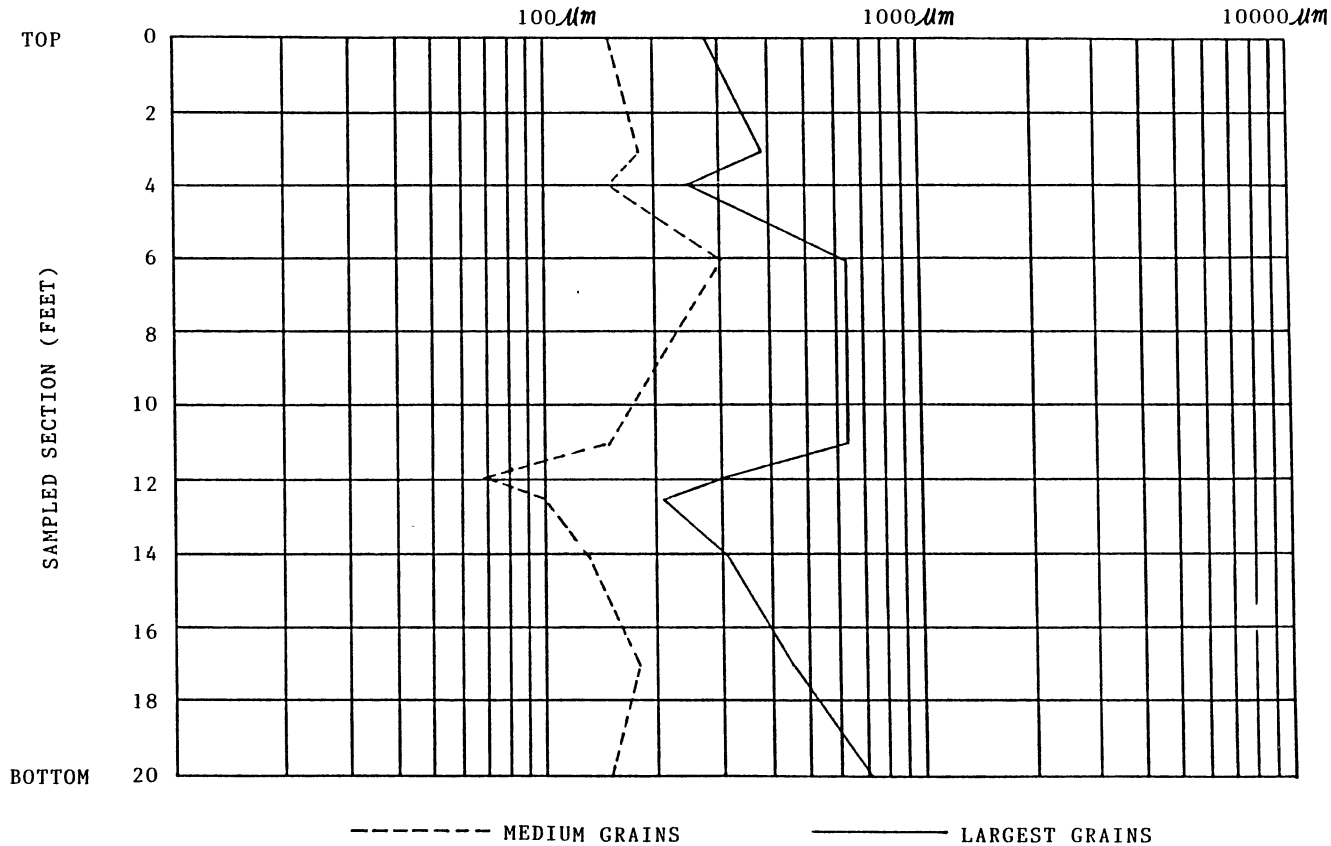


Figure 41. Vertical Variation of Grain Size in the Sampled Section of the Morrison Formation (OK Geology Camp Area)

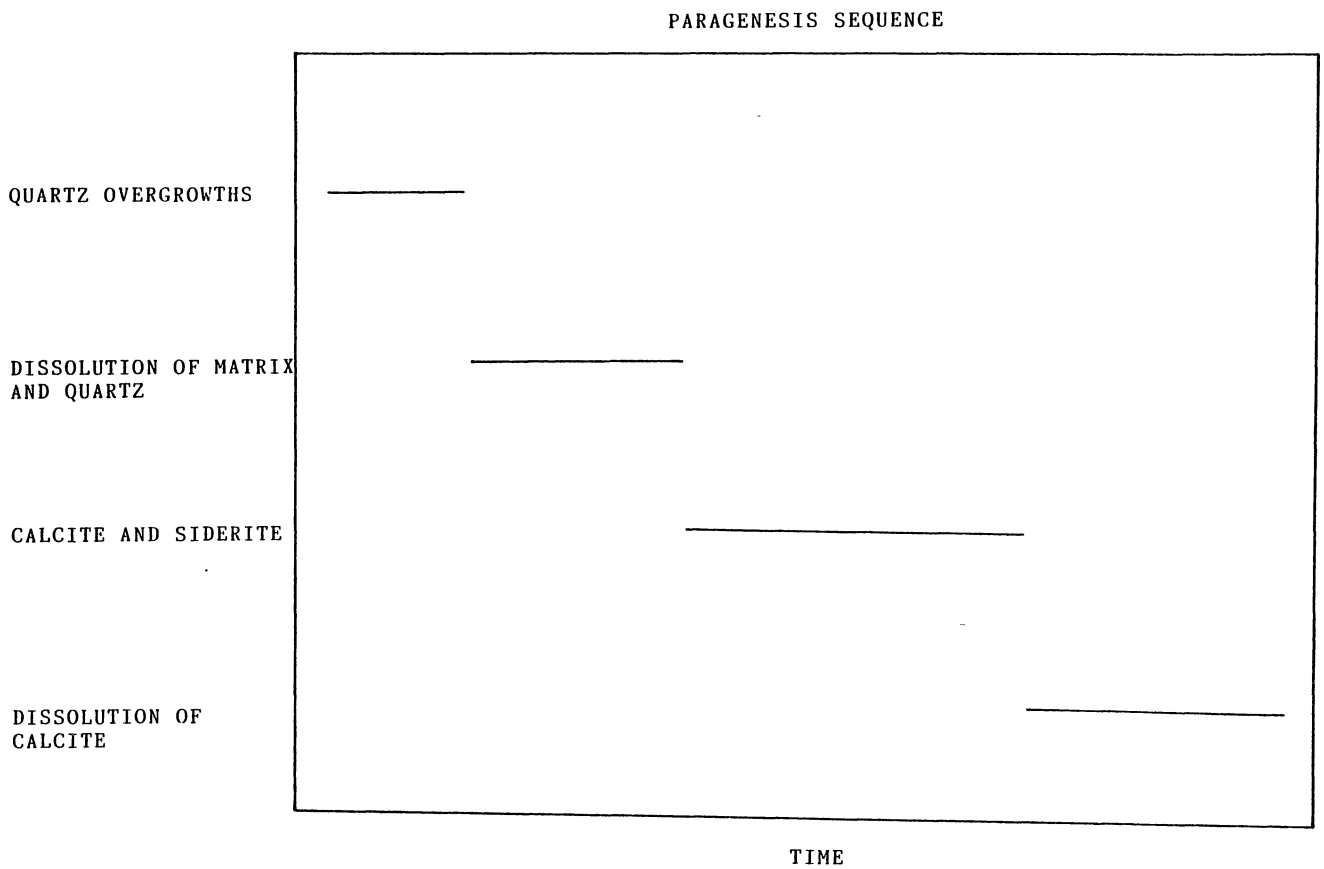


Figure 42. Diagenesis of the Sandstone in the Meander (Morrison Formation, OK Corral Area)

some quartz were dissolved. Following the dissolution of these silicates, calcite, and siderite were precipitated. There are possibly two generations of calcite, the first fibrous and the second blocky. Later, the dissolution of the carbonates created some secondary porosity.

The diagenetic sequence of the samples from the Oklahoma Geology Camp area is shown in Figure 43. After deposition, the rocks were subjected to an alkaline environment ($pH > 8.3$), leading to the dissolution of the matrix. Then, quartz overgrowth was precipitated when the pH value dropped below 8.3. Later, chalcedony, which is brown and perpendicular to the edges of quartz grains, formed in an impure silica-rich solution (Figure 44). The following microquartz (Figure 44), whose precipitation took place in a pure solution, is clear and pore-filling. Still later, the pH value went high again, leading to the precipitation of calcite and dolomite which replaced some quartz grains and secondary silica. The dolomite has zonations. Thereafter, the dissolution of the carbonates created some secondary porosity.

Age. On the basis of paleontology studies in the Garden Park area, Hatcher (1901) assigned a Late Jurassic age to the Morrison Formation. This assignment is agreeable with Cramer's (1962) pollen dating of the underlying Ralston Creek Formation as Middle and Upper Jurassic, and the transitional contact between the two formations.

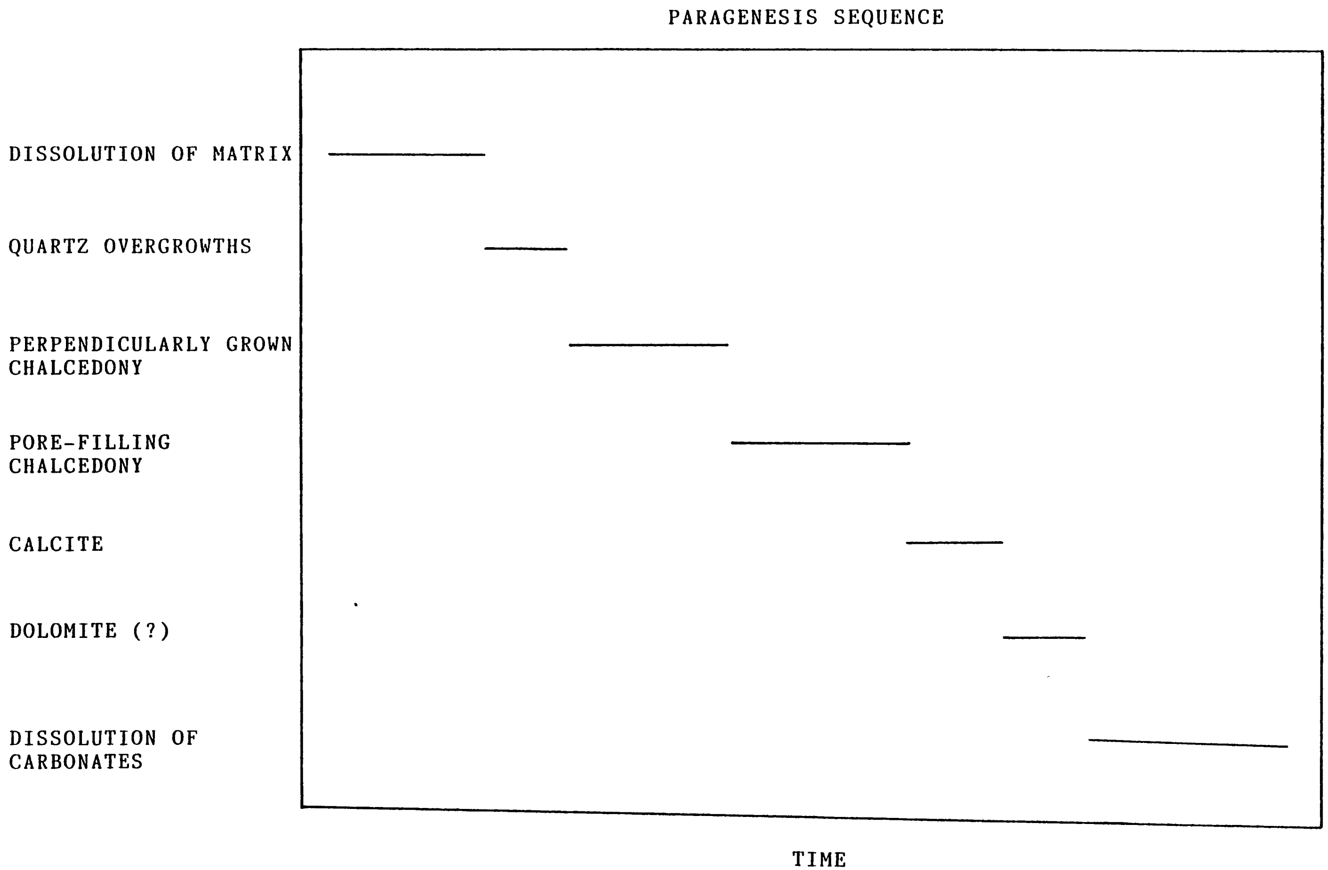


Figure 43. Diagenesis of the Morrison Formation (OK Geology Camp Area)

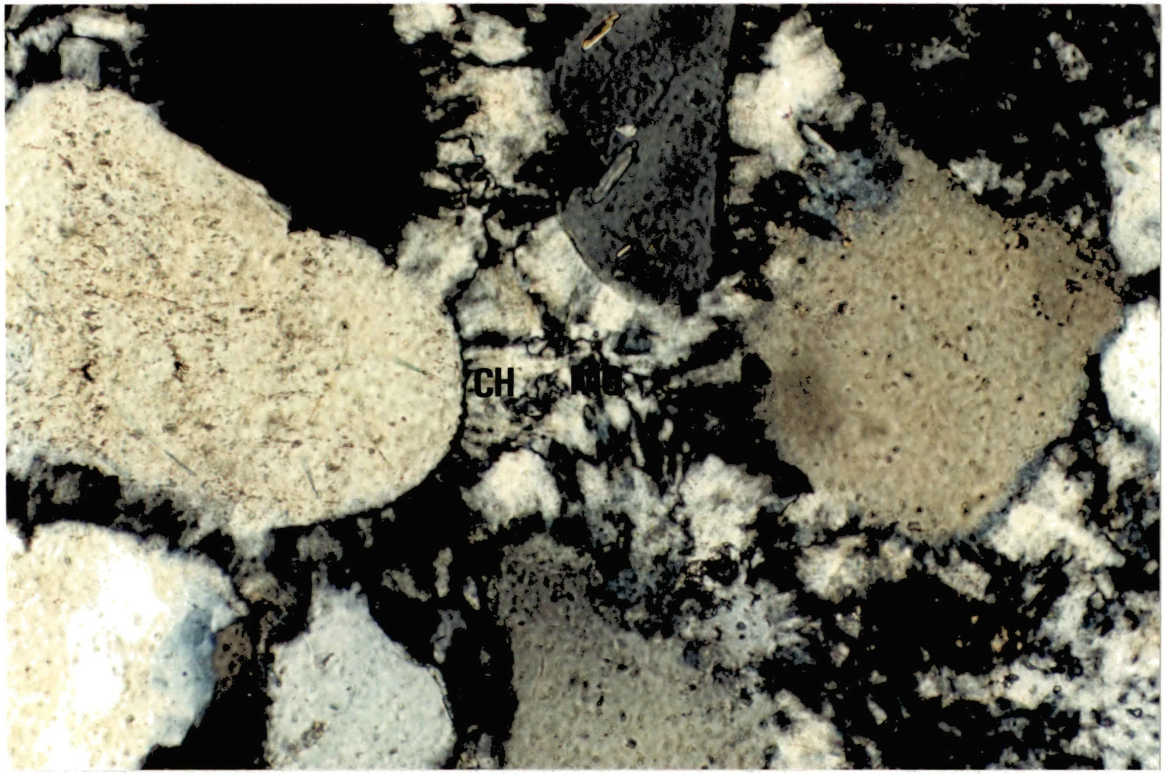


Figure 44. Well-Developed Chalcedony (CH) and Microquartz (MQ) in the Morrison Formation (OK Geology Camp Area)

<u>Depositional Environment (Sandstone Facies)</u>. The dinosaur bones and non-marine mollusk fragments indicate a non-marine environment. Sedimentary structures, such as the cross-bedding and climbing ripple marks as well as lateral and vertical stratification properties, suggest a channel deposit. However, even in a non-marine environment, channel deposition can be found in several circumstances; for instance, bars and beaches associated with large lakes, deltas, and alluvial fans, and flood plains. The disconformable contact of the Morrison Formation with the underlying Ralston Creek Formation suggests an alluvial origin. This is because alluvial channels erode downward into the lower sediments while bars and beaches are generally built on the underlying sediments. A distributary channel and its associated delta plain deposit sandstones and siltstones with the same pattern as that of the Morrison sediments; however, under the sandstones and siltstones there should be siltstones and clays which are formed in the delta front of brackish to marine conditions and the prodelta of a marine condition. Since these siltstones and clays are not present in the embayment, it is improbable that the Morrison is of a large delta origin. According to Leeder (1983), a floodplain is composed of coarse sediments of bed load deposition and associated with fine sediments of overbanking flow deposition. The coarse sediments (sandstones and conglomerates) occur in the fine sediments (siltstones and claystones) as the deposits of channel lags, point bars, and channel bars.

Thus, the occurrence of the bedded and lenticular channel sandstones in association with claystones in the Morrison Formation can best be explained by deposition in a flood-plain environment.

Leeder (1983) also discussed the patterns of deposits formed by braided and meandering streams. Because braided streams move freely over the floodplains, little or no argillaceous sediments of overbanking origin can be preserved. Therefore, coarse materials of bed load deposition dominate sediments of braided stream origin. On the other hand, a meandering channel occupies only a small portion of its alluvial plain at a given time. It lies within a meander belt which consists of active channels, abandoned channels, and near channel subenvironments. The meander belt shifts its position on the alluvial plain, with the shifting frequency depending on channel sinuosity. A highly sinuous channel is generally stabilized by clay and fine sediments generated by frequent channel cut-offs with sediments being deposited in the vicinity of the meander belt. Therefore, an unstable "alluvial ridge" rises on the alluvial plain. The branching of a channel bank and/or the sudden shifting of the meander belt to a new position on the flood plain relieves the above unstable condition. In this case, the bodies of the coarse sediments, which are limited in the narrow, linear meander belt, are separated by different clay and fine sediments deposited from suspended loads laterally and vertically. Low sinuosity streams have fewer cut-offs and thus fewer clay plugs, making their positions unstable. They sweep across floodplains more continuously and frequently than the meandering streams with high sinuosity. Allen's (1965) study shows that the deposits laid down by low sinuosity streams consist of coarse sediments which extend across the width of the alluvial plain as wedge-shaped sheets bounded by scour surfaces.

The sandstone bodies in the Morrison Formation throughout the embayment are not continuous. This fact is contrary to the distribution patterns of sandstone bodies laid down by braided and low sinuosity rivers. The predominance of claystones (71%) suggests that vertical accretion by overbanking flow played the most important role in the deposition of the Morrison Formation (Brady, 1969). Therefore, the Morrison was most probably deposited by high sinuosity streams. The presence of many small channel sandstone bodies suggests that the streams were very small. For example, one meandering channel, whose outline is still obvious in the OK Corral area, was about 25 feet wide.

Carbonate Facies. The limestone beds in the basal Morrison Formation are believed to be of fresh water lake origin, because the lateral extension of each bed is limited. The lakes were very small and possibly temporary in a low-land environment. With continued progradations of fluvial deposits from the west, the lake deposits were replaced by silts and clays of overbank origin.

The micritic texture and the scarcity of skeletal fossils indicate that the limestones were generated by chemical precipitation as a result of temperature variation and evaporation. Around the Skyline Drive area, limestone beds interstratify with siltstone or claystone beds. This interstratification suggests that lakes developed during dry seasons were filled with sediments laid down by overbanking flows during the wet seasons.

According to Brady (1967), dolomite in the formation is of both penecontemporaneous and secondary origin. The dolomite at the base of one section measured by Brady (1967) is thin, nonfossiliferous, and finely crystalline (<0.0039 mm). The above features of dolomite fit the criteria of penecontemporaneous (primary) dolomite proposed by Deffeyes, Lucia, and Weyl (1965). However, another dolomite sample, whose crystals are larger than 0.016 mm, is considered secondary on the basis of Folk's (1959) statement that dolomite crystals with grain sizes larger than 0.006 mm are of secondary origin.

Dakota Group

Outcrop and Stratigraphical Relationships. The Dakota Group is divided into three members: Lytle Sandstone, Glencairn Shale, and Dakota Sandstone. These members crop out nearly continuously along the margin of the embayment and generally form cliffs. Locally, they are off-set by faulting.

The Dakota Group is conformably overlain by the Graneros Shale member of the Benton Group.

Lithology. The lower, basal portion of the Lytle member consists of yellow to brown, fine- to medium-grained, algillaceous to quartzose and cross-bedded, lenticular and conglomeratic sandstone. The upper part of the unit is moderately- to well-sorted, however, the basal portion is poorly sorted. Cross-bedding is commonly the trough type. The middle portion of the Lytle is composed of sandstone, sandy siltstone, and sandy claystone that interstratify vertically. The sandstone is the same as that of the lower portion (basal portion) of the member discussed above except for the presence of ripple laminations and the gradation from sandstone to siltstone. The top portion of the Lytle is comprised of medium- to very fine-grained, thinly bedded, ripple-laminated, and moderately friable to well-cemented sandstone which contains abundant wood fragments and clay chips. Tabular and trough cross-bedding are common.

The Glencairn Member consists of gray to olive-green, fine-grained, and very thinly-bedded sandstone interbedded with dark gray, thinly-bedded shale. The sandstone is ripple laminated, and heavily burrowed.

The yellow to brown Dakota Sandstone is divided into lower and upper portions. The lower portion consists of medium-grained quartzose sandstone. The sandstone contains red speckles of iron stain and is moderately- to well-sorted, cross-stratified, and lenticular. The upper portion is fine to very fine sandstone, thinly- to thickly-bedded, and cross- to ripple-laminated.

<u>Thin-Section Description</u>. Samples from the Lytle Sandstone of the group were collected near the Oklahoma Geology Camp area. Petrologic study under a polarizing microscope shows that monocrystalline quartz is the main component. The quartz grains are of straight to slightly undulose extinction, and their common inclusions are zircon, mica (mainly biotite), and vacuoles. The percentage of the polycrystalline quartz varies from 3 to 5.5%. Its grains are composed of 3-4 single quartz crystals, and the boundary between the crystals is irregular, indicating a possible metamorphic rock origin. Chert grains (1-2%) tend to be smaller than the quartz grains. Matrix is a mixture of different clay materials. Both the chert and matrix suffered a great deal from dissolution. Hematite (1-4%) occurs in patches. Other detrital clastics are plagioclase, zircon, tourmaline, and mica, which occur in trace amounts. The rocks are cemented by quartz overgrowths (3.5-5%) and microquartz. The quartz overgrowths have well-developed straight edges and apparent dust rims. Detrital grains are subrounded to rounded and well-sorted because they once were in a turbulent near-shore environment.

Figure 45 shows small variations of monocrystalline and polycrystalline quartz. Figure 46 shows the variations of secondary minerals and matrix. In the bottom portion, quartz overgrowths, kaolinite, and the matrix follow the same pattern because their dissolution is controlled by the same factor--pH value. Matrix disappears in the middle due

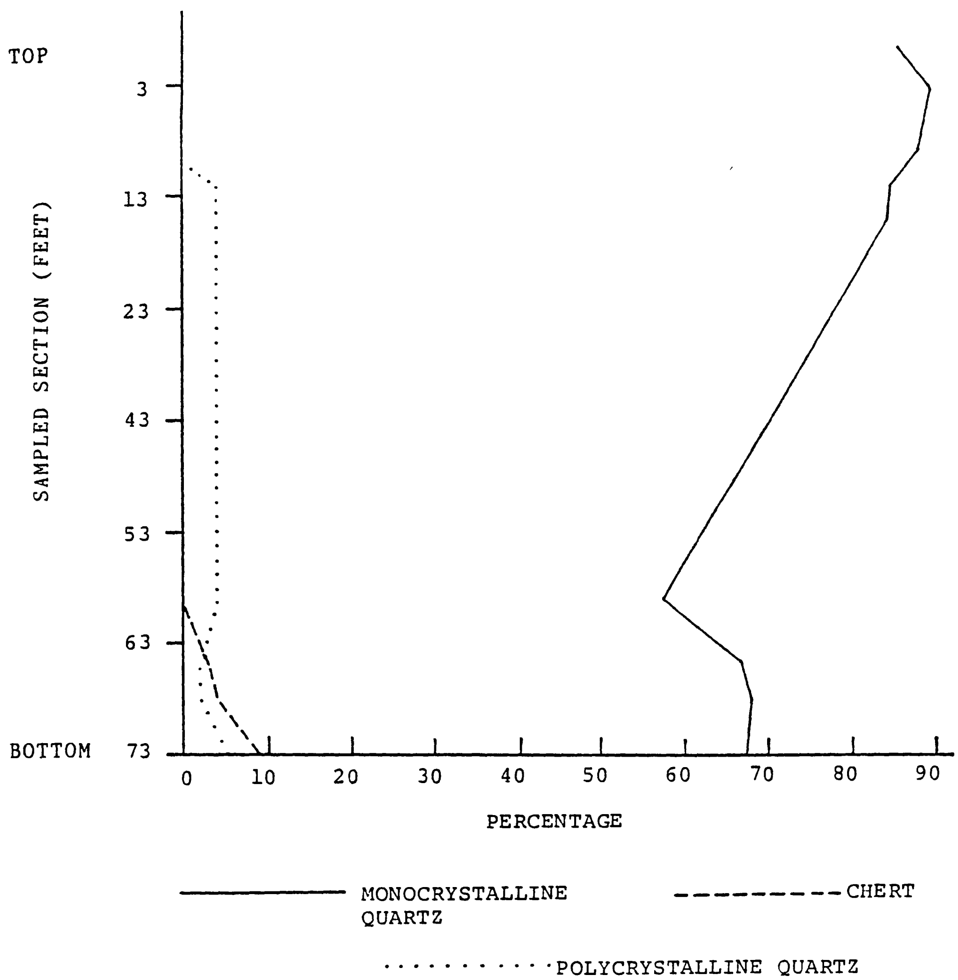


Figure 45. Vertical Variations of Detrital Components in the Sampled Section of the Lytle Sandstone (OK Geology Camp Area)

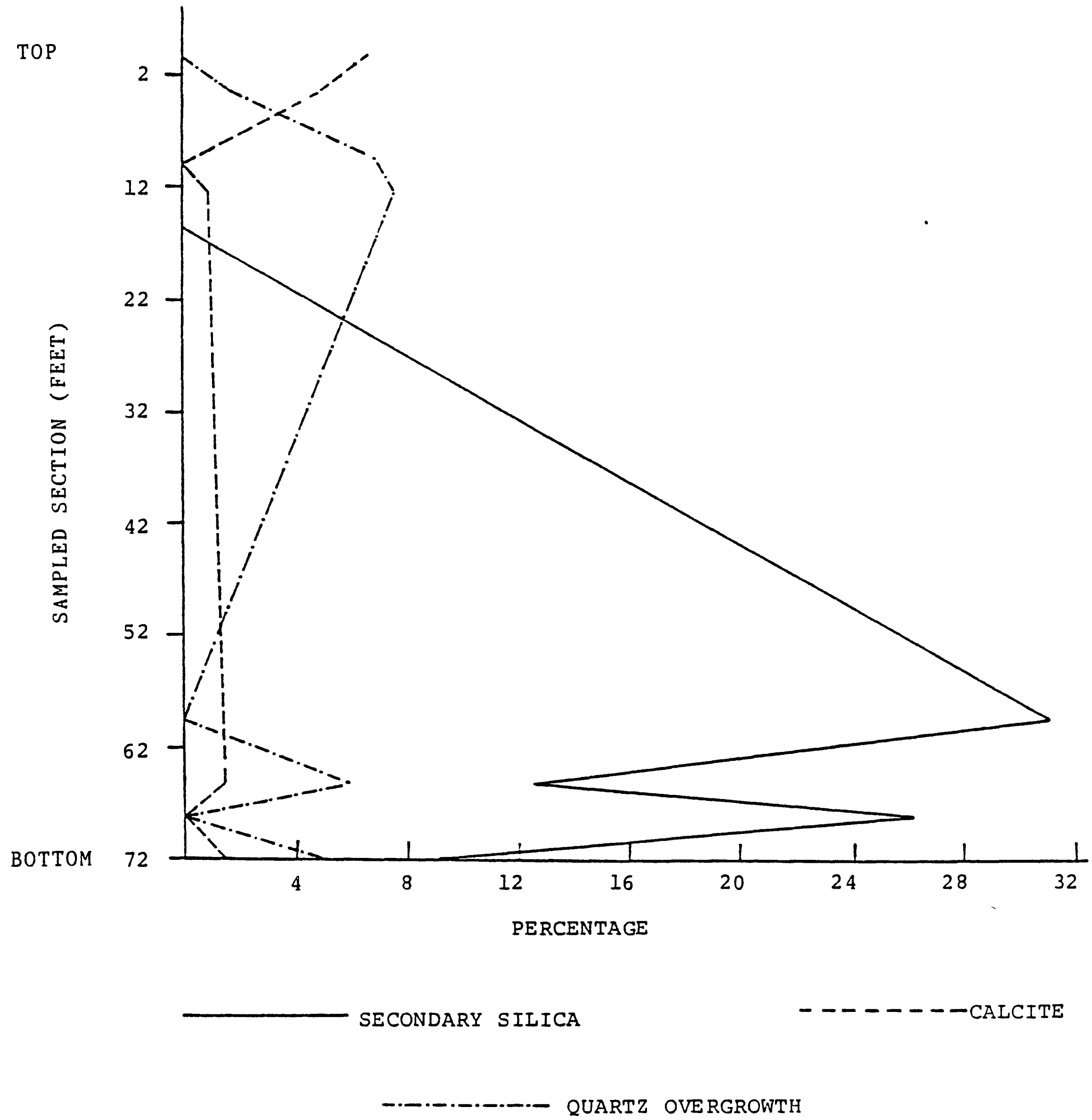


Figure 46. Vertical Variations of Secondary Minerals in the Sampled Section of the Lytle Sandstone (OK Geology Camp Area)

to dissolution. At the top, the curves of matrix and quartz overgrowths are counteractive against each other. This relationship occurs because more matrix makes it difficult for a silica-rich solution to pass through a bed. Figure 47 shows the variation of grain size.

Diagenesis. The first diagenetic event in the Lytle was the formation of quartz overgrowths (Figure 48) in an acidic environment; the quartz overgrowths are well-developed. Some silicate framework grains were dissolved when the pH value of the formation solution rose, giving rise to a significant amount of secondary porosity (Figure 49). Some pore spaces were oversized (Figure 50). Later, the precipitation of microcrystalline quartz occurred. The above diagenetic sequence is shown in Figure 51.

Age. Palynological evidence indicates that the Lytle and Glencairn are of Late Cretaceous age (Stoke, 1952. Young, 1960). Stose (1912) and Finlay (1916) dated the Dakota Sandstone Member as Early Late Cretaceous age; however, Waage's (1955) palynological investigation indicated that this member is the uppermost Early Cretaceous in age. According to Weimer (1970), the Glencairn Shale and the Dakota Sandstone can be correlated with the Kiowa Shale and the Cheyenne Sandstone, respectively, in Western Kansas.

Depositional Environments. Weimer (1969) divided the Dakota Group into three genetic units (Figure 52). Genetic unit one corresponds to the Lytle Sandstone which was laid down in a fluvial channel-floodplain environment. Dominant

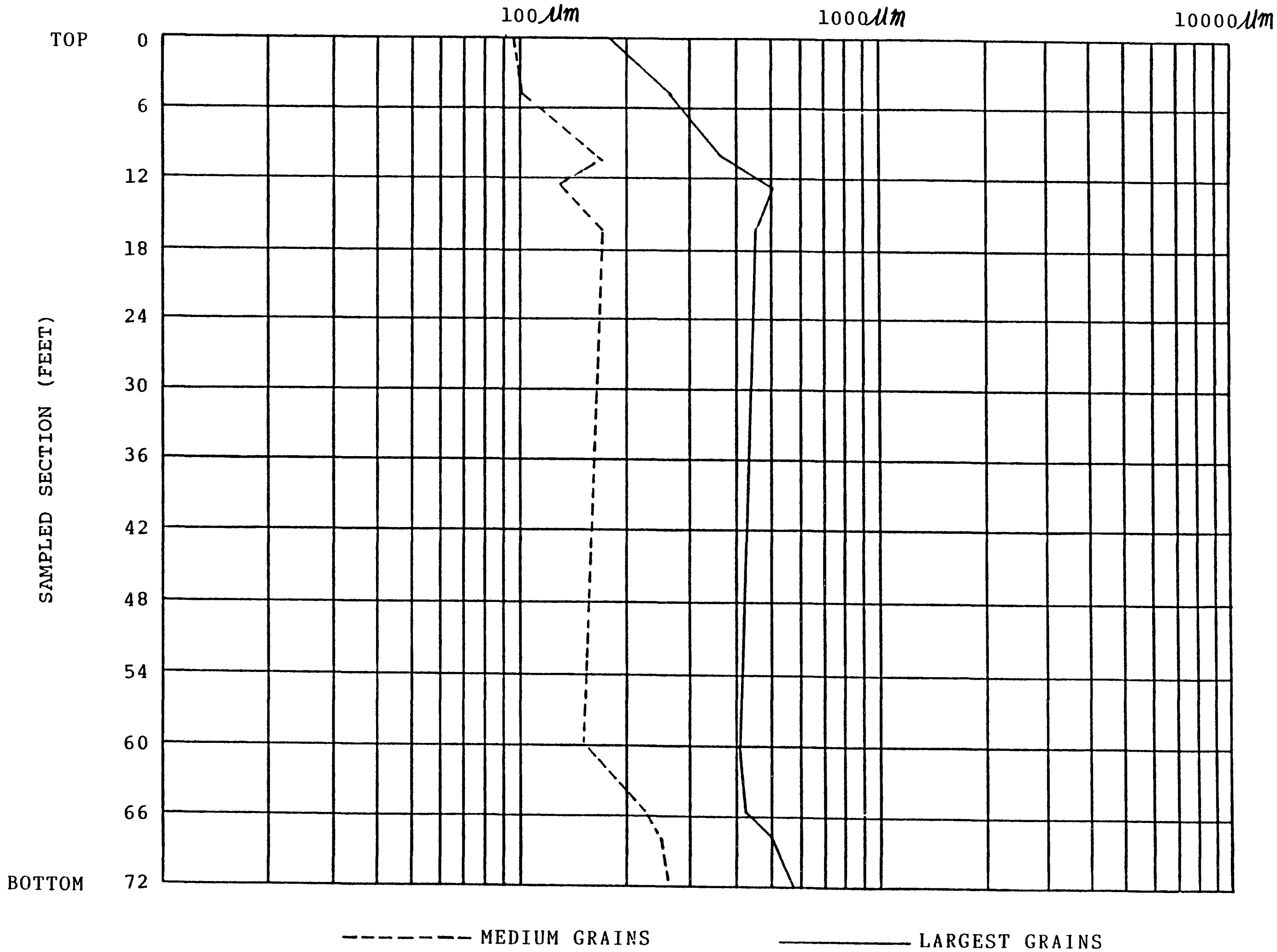


Figure 47. Vertical Variation of Grain Size in the Sampled Section of the Lytle Sandstone (OK Geology Camp Area)

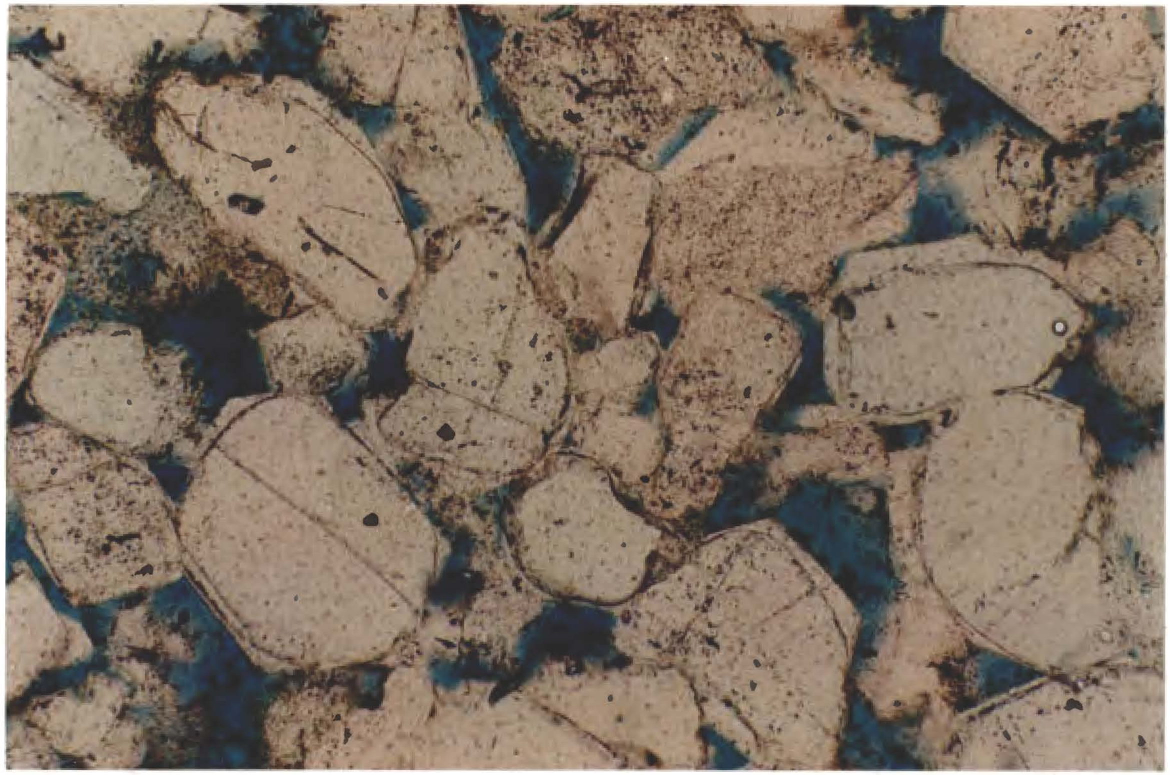


Figure 48. Well-Developed Quartz Overgrowths in the Lytle Sandstone, 10 x 10 (-) (OK Geology Camp Area)

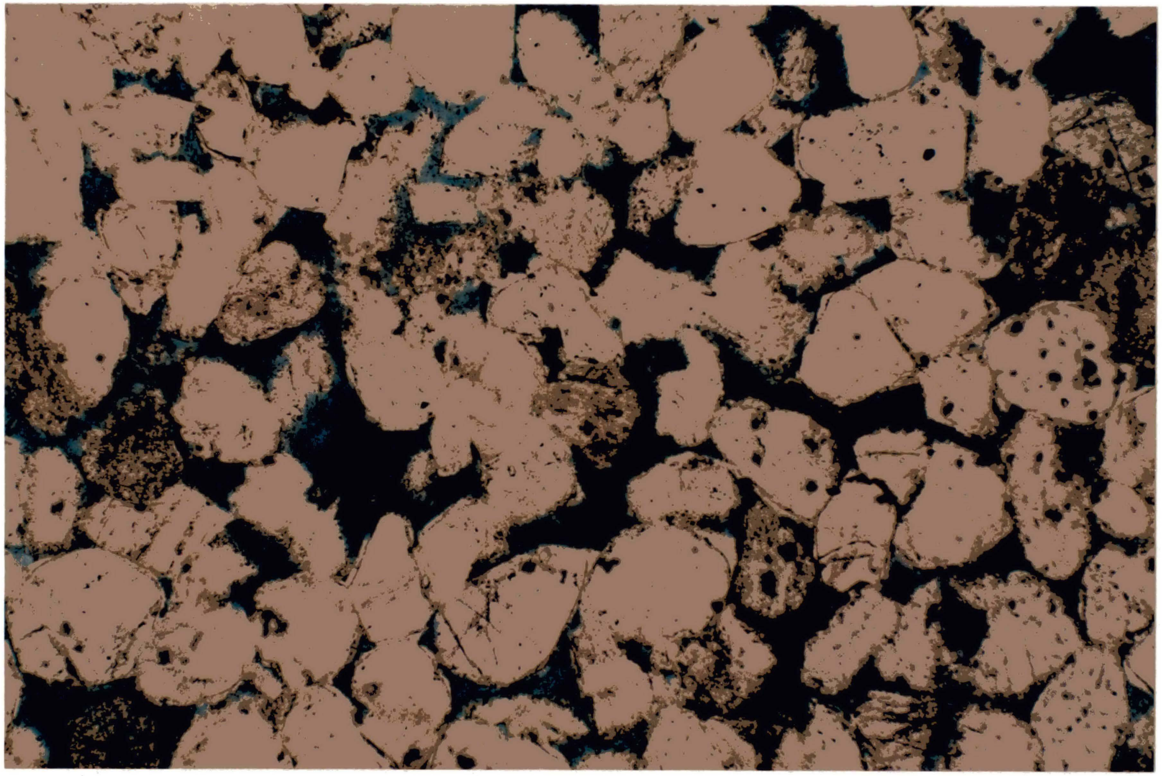


Figure 49. Well-Developed Secondary Porosity in the Lytle Sandstone, 10 x 10 (-) (OK Geology Camp Area)

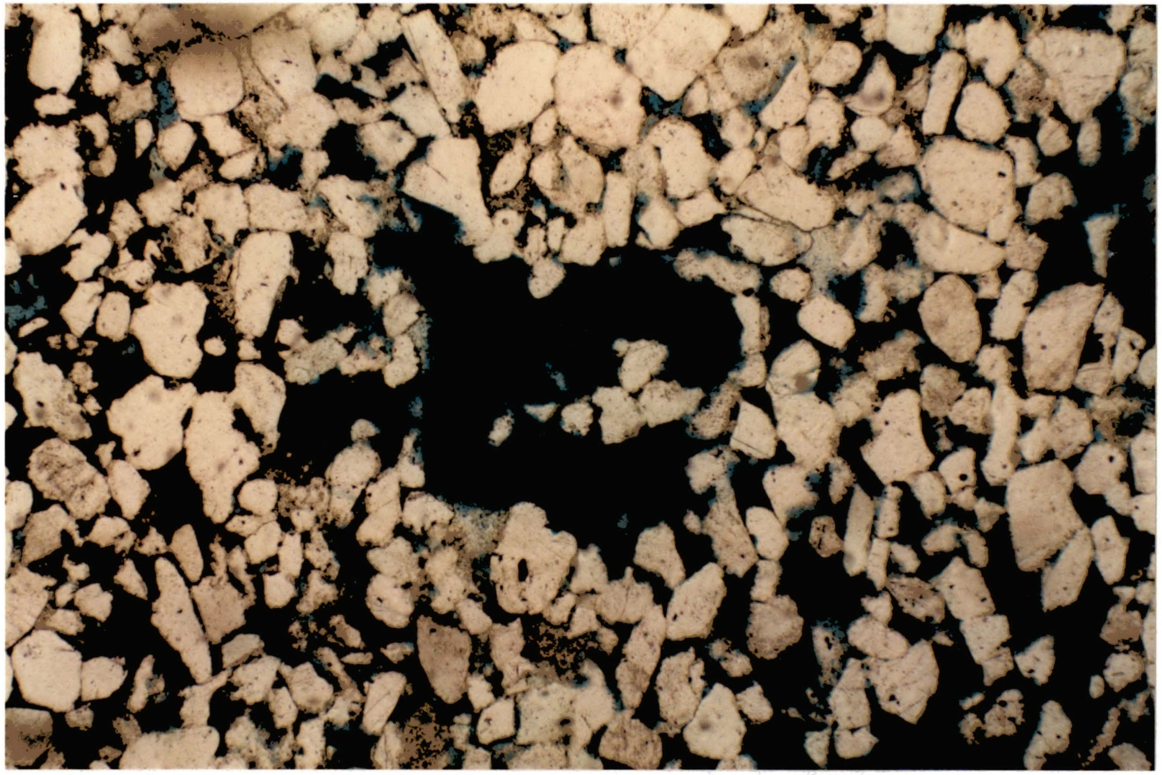


Figure 50. Oversized Secondary Porosity in the Lytle Sandstone, 10 x 4 (-) (OK Geology Camp Area)

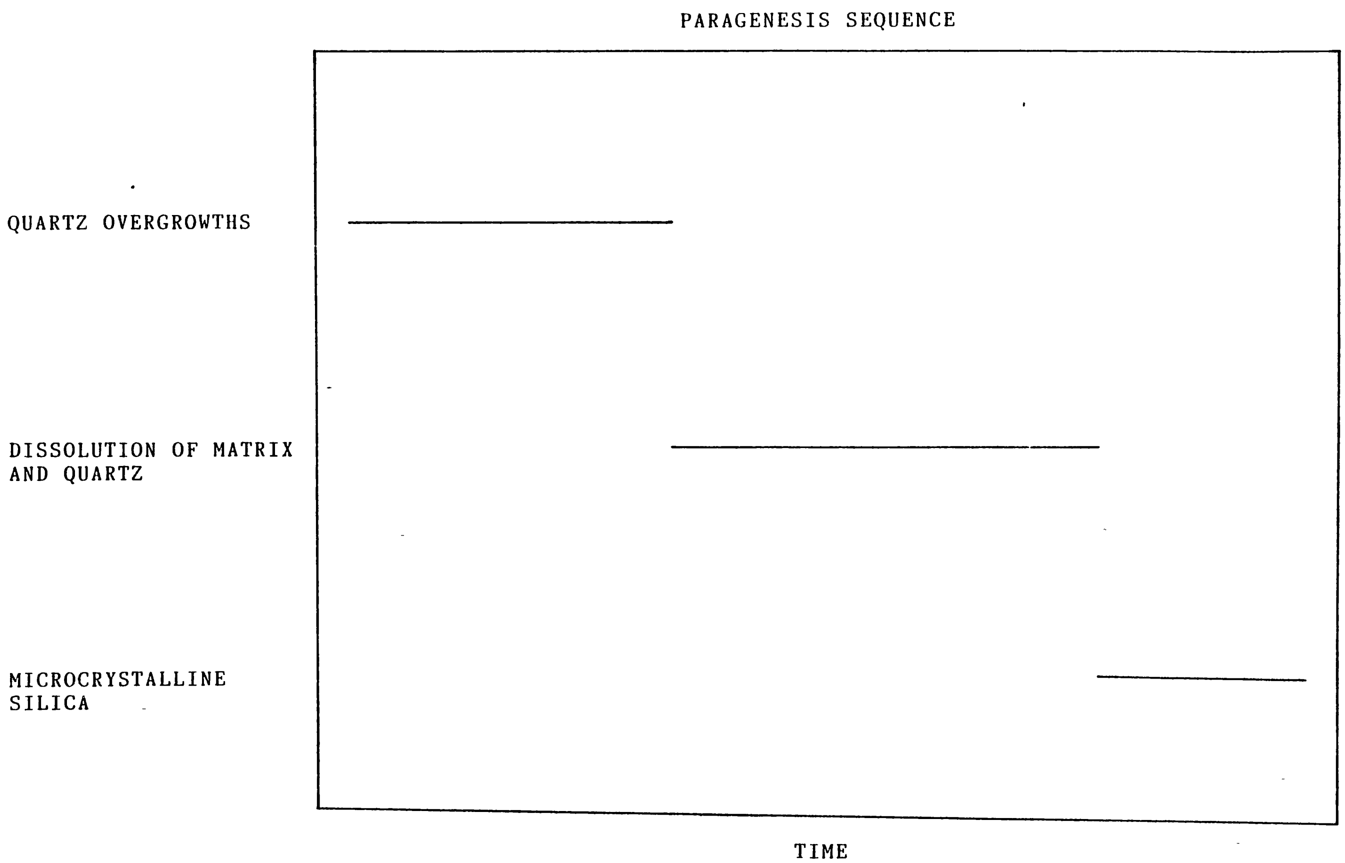


Figure 51. Diagenesis of the Lytle Sandstone (OK Geology Camp Area)

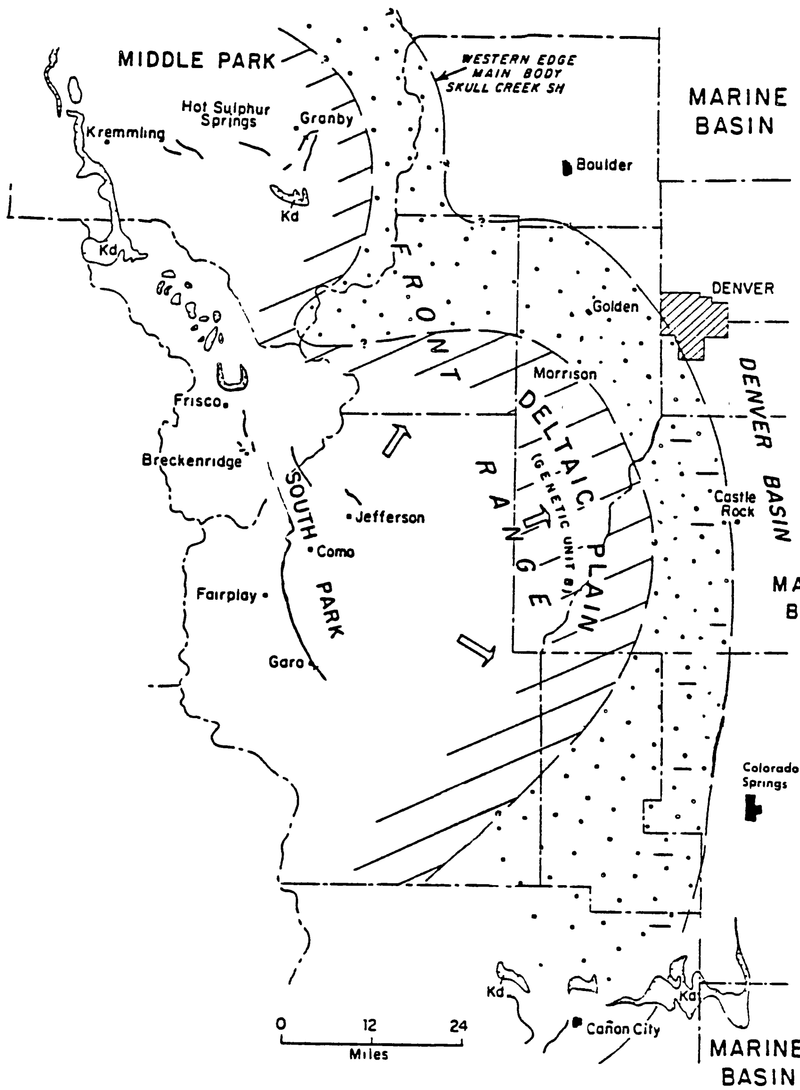


Figure 52. Depositional Environments of the Dakota Group. Stippled Pattern is Mixed Marine Sandstone and Shale (After Weimer, 1970)

sediment transport direction of the channels in the embayment was to the southeast. The width of the individual sandstone bodies ranges from a few hundred to a few thousand feet with thicknesses ranging from a few feet to as much as 60 feet. As a result, the streams were generally larger than those during the deposition of the Morrison Formation. The abundance of claystones and siltstones suggests that the streams had a very high sinuosity. Genetic unit two corresponds to the Glencairn Shale. After the deposition of the continental Lytle Sandstone, submergence of the Canon City area as well as other areas of the eastern Colorado occurred. Consequently, the early Cretaceous sea covered the whole region of the present Front Range. The transgression of the sea generated a thin layer of conglomeratic lag deposits at the base of the Glencairn Shale. The conglomeratic lag may indicate the erosion associated with the marine transgression. The following two observations suggest that the studied area was not far away from the shoreline at the time of Glencairn deposition: (1) the interstratification of sandstone and shale beds; and (2) the bioturbation of the top sandstone beds of the member. The interstratification was probably caused by the shift of the streams which provided the area with coarse materials. When the streams shifted from the Canon City area to other places, the studied area could only receive fine materials, which led to the development of shale beds. Intensively burrowed sandstones are observed in the Skyline Drive area. Genetic unit three corresponds to the Dakota Sandstone. It was deposited on a widespread coastal plain during a time of the early Cretaceous sea regression. The transportation direction was to the south-southeast. As previously discussed, the Dakota Sandstone is cross-stratified and lacks shale beds. Therefore, the streams which laid down this member were braided or had a low sinuosity. Alternatively, the sandstones might have been deposited along the shore line, as in the case of a high destructive wave-dominated delta where waves and shore currents winnowed different kinds of clays and silt particles away, leading to fine-grained clean sandstones.

Weimer's (1969) study suggests that there is another unit on top of the Dakota Sandstone in the Southern Front Range area which is composed of interbedded sandstones, siltstones, and shales with abundant trace fossils. The sandstones are fine- to medium-grained and lenticular. Unusually, this unit contains thin kaolinite seams, thin coal or carbonaceous layers, lag and wood chips, bioturbation, ripple marks, and dinosaur footprints. Weimer (1969) interpreted this unit as the deposits of tidal flats, marshes, tidal channels, and fluvial channels on a deltaic plain. This unit may mark the initial submergence of the coastal plain that was associated with the transgression of the Late Cretaceous sea which laid down the overlying Graneros Shale (lower Benton Group).

Benton Group

Outcrop and Stratigraphic Relationships. The Benton Group is divided into three members in an ascending order: Graneros Shales, Greenhorn Limestone, and Carlile Shale. The Graneros Shale is continuous along the margin of the embayment, however, the other two members are locally cut out by faulting.

The Graneros Shale is in transitional contact with the underlying Dakota Sandstone and in a conformable contact with the overlying Greenhorn Limestone, which is conformably overlain by Carlile Shale. The Carlile Shale is disconformably overlain by the Niobrara Group.

Lithology. The Graneros Shale is brittle, hard, and generally bluish-gray; the bottom and top beds are lighter in color than the middle ones. Marine shells are abundant. Occasionally, calcium-carbonate concretions are observed. The total thickness of this member varies from 200 to 325 feet.

The Greenhorn Limestone, which is separated from the Graneros Shale by layers of dark gray, limy shale, is pale to bluish-gray, slabby, and fine-grained. Inoceramus labiatus, a highly characteristic fossil, is abundant in this member. Its total thickness ranges from 40 to 60 feet.

The Carlile Shale member is composed predominantly of dark-gray shale. At its top an interval of yellow sandstone, named the Codell Sandstone, is present. The total thickness of this member ranges from 100 to 150 feet.

Krutak (1970) conducted a comprehensive study of the Codell Sandstone. He divided the unit into two facies: (1) a lower "basal sequence"; and (2) an upper "upper sequence". The basal sequence is composed of pale orange, calcareous sandstones interbedded with very thin olive-gray shales. Its thickness varies from 0 to 10 feet. This sequence is characterized by bioturbation, scattered mollusk remains, irregular oscillation and current ripple marks, and penecontemporaneous deformation structures. The upper sequence is relatively thinner and contains abundant fragments of mollusks. A surface containing extensive borings which penetrate underlying beds separate this sequence from the basal sequence. Krutak (1970) divided the upper sequence into the following five subunits in a descending order: (1) limy sandstone; (2) sandstone containing limonitized borings; (3) shaly to massive sandstone; (4) sandy limestone; and (5) sandy shale with restricted areal extent.

Age. According to Weimer (1960), the Benton Group is of early Coloradan age (from 110 m.y. to 81 m.y. Late Cretaceous). It can be correlated with the lower Mancos shale in the San Juan basin, New Mexico, the Frontier Formation of Northern Colorado, and the combination of the upper Tununk Shale and Ferron Sandstone in Utah.

Depositional Environment. According to Molenaar (1983), the transgression of the Graneros sea was so extensive that the shoreline was about 600 miles southwest of the

Canon City area, reaching the southwestern part of Arizona. Based on his paleontological study, Kauffman (1969) concluded that water depths ranged between 200 to a maximum of 400 feet. Therefore, the studied area was within a sublittoral environment. The predominance of clay-size sediments, laminations, and the continuousness of bentonites in the Graneros suggest a low energy condition.

The predominance of shale in the lower portion of the Greenhorn Limestone throughout southeastern Colorado indicates that the lower Greenhorn was deposited in an outer shelf environment. The shale is interrupted by limestones several times. The limestones are calcarenite and composed of bioclastic debris (predominantly Inoceramus shells and fish bones) in a lime mud matrix (Kauffman, 1969). The occurrence of limestone layers might be due to bottom currents, periodic shoaling, or the cessation of clay-size sediment supplies. Thus, the depositional environment of the lower Greenhorn is about the same as that of the Graneros Shale. According to Kauffman (1969), the upper portion of the Greenhorn Limestone was deposited during the peak transgression of the Greenhorn sea. He defined two biological assemblages in this portion. The first is composed of thin-shelled subequivalves, nonbyssate, and inoceramid bivalves which are adapted to a quiet water, middle and outer shelf environment. The second consists of various ammonite types, deep water types of Pectinoids, and nonbyssate inoceramids.

These two groups suggest a quiet, deep water condition. However, associated with the two assemblages are rudistid bivalves, cerithiid gastropods, cap-shaped gastropods, and gryphioid oysters which occur in zones of dense burrow mottling. The occurrence of the above organisms and burrowed zones indicate a shallow water environment. Kauffman (1969) accommodated the above conflicting observations by suggesting a mid-basin, moderately shallow water environment for the upper portion of the Greenhorn Limestone and the overlying Carlile Shale. This environment was evidently within the photic zone (300 feet or less) because borrowing organisms were able to grow. However, since calcareous siltstone was preserved in the absence of the influence of strong current flow, the environment should have been below the effective wavebase (100 feet or greater).

The Carlile SHale was deposited during the regression of the Greenhorn sea. A water depth of less than 100 feet and even less than 50 feet is evidenced by the widespread occurrence of thin lenses of calcarenite with abundant worn fish bones, ostreid biostromes, ripple bedding, flow casts, and current-sweep surfaces in the basal part of the Carlile (Kauffman, 1969). There was most probably a small transgression of the Greenhorn sea during the deposition of the middle Carlile Shale or the local subsidence rate became greater because the uniform chalky shale with thin persistent bentonites as well as the presence of thinly-shelled nonbyssate bivalves indicates water depths at that time ranged from 200 to 400 feet. Then, the sea began shallowing again and the upper portion of the Carlile Shale was deposited fully within the mid-shelf zone of terrigious clay, as evidenced by the noncalcareous shale. The presence of inoceramid bivalves in the upper Carlile Shale also indicates a quiet to gently agitated water of mid-shelf depth (Kauffman, 1969). Upward, the Carlile Shale becomes more silty and sandy, bedding becomes more irregular, and bioturbation becomes stronger. At last, the Carlile Shale is replaced by the Codell Sandstone of a near-shore origin. Figure 53 shows the depositional environment of the upper Carlile Shale-lower Fort Hays interval.

Niobrara Group

Outcrop and Stratigraphic Relationships. The Niobrara Group is divided into two members, the Fort Hays Limestone member and the Smoky Hill Shale member. The Fort Hays crops out continuously along the edge of the embayment, while the Smokey Hill occupies the whole eastern part of the basin north of the Arkansas River and crops out along the basin margin as well. The group is conformably overlain by the Pierre Shale.

Lithology. The Fort Hays Limestone is composed of gray, hard limestone layers, 1 to 26 inches thick. The layers are separated by thin shale partings, with a total thickness of 40 feet (Scott, 1977). Fossils are abundant.

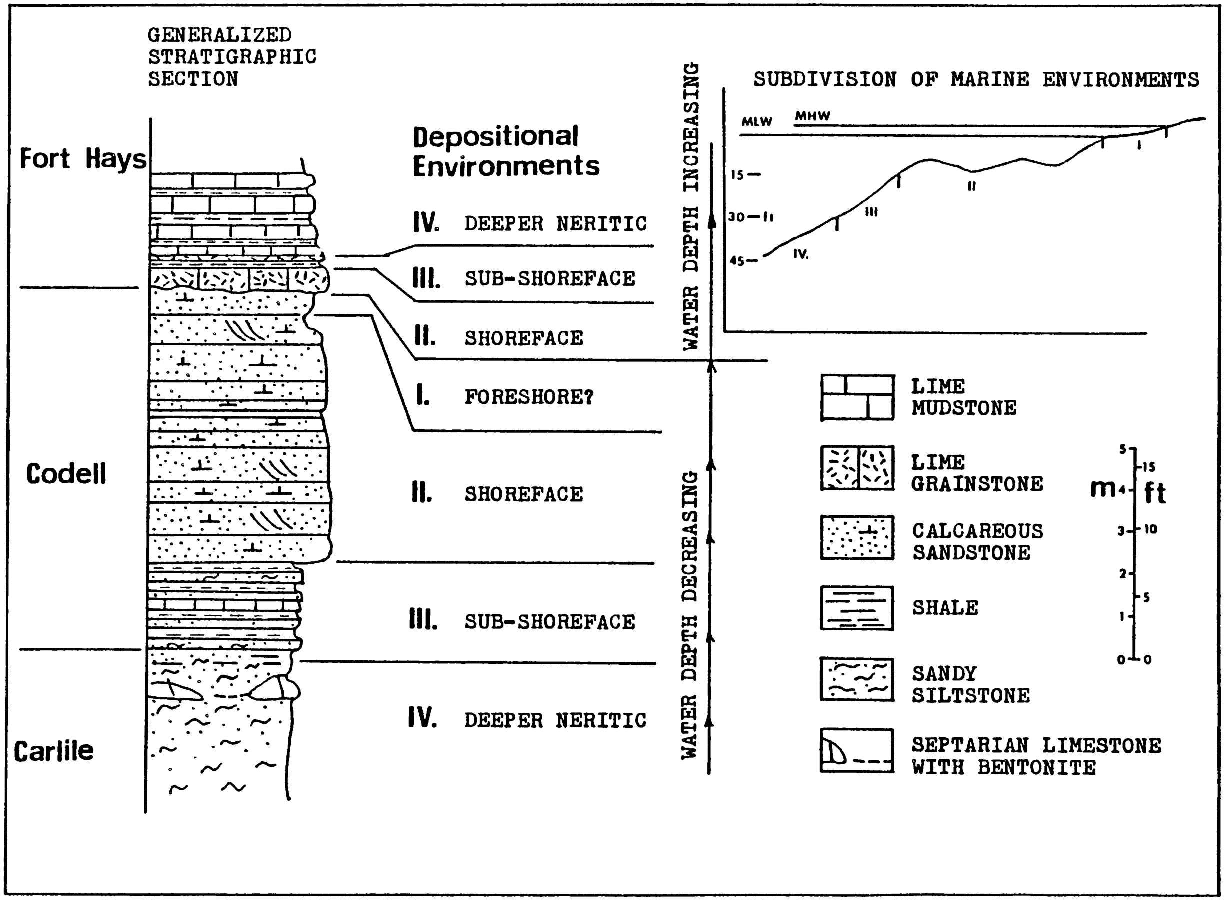


Figure 53. Depositional Environments of the Interval from the Top of the Carlile Shale to the Bottom of the Fort Hays Limestone (After Pinel, 1977)

The Smoky Hill Shale consists of yellow-brown, soft, thinly-bedded, calcareous shale and thinly-layered interbedded limestone. It is divided into seven units from bottom to top: (1) gray shale and limestone; (2) dark-gray fissile to platy lower shale; (3) gray hard platy lower limestone; (4) dark-gray middle shale; (5) yellow-gray middle chalk or limestone; (6) yellowish-orange upper chalky shale; and (7) yellowish-orange upper chalk.

Age. The Niobrara Group is of late Coloradan age (from 110 m.y. to 81 m.y. Late Cretaceous). Southwestward, it is gradually replaced by sandstone; the sandstone is named Gallup sandstone in the San Juan basin, New Mexico. The group is also correlated with the Blue Gate Shale in Utah (Weimer, 1960).

Depositional Environment. Molenaar (1983) determined two transgression and regression cycles during the deposition of the Niobrara Group. The first transgression is represented by the Fort Hays Limestone and the lower portion of the Smoky Hill Shale. At that time, the Canon City area was about 300 miles from the southwestern shoreline of this transgression. However, Kauffman (1969) believed that the Fort Hays Limestone was formed during the time of maximum transgression and the lower part of the Smoky Hill was deposited at the beginning of the regressive facies of his Niobrara marine cycle. Kauffman's (1969) paleontological and sedimentational investigations show that a shoaling midbasin carbonate bank developed during the maximum transgression. This carbonate bank was within the photic zone but below the influence of strong currents (about 200 feet deep), and includes the studied area. During Molenaar's (1983) first regression, the deposition of the middle shale of the Smoky Hill took place; however, the shoreline was still far from the studied area and the deposition was similar to that of the lower shale of this member.

The second transgression is represented by the middle chalk and possible the lower part of the upper shale of the Smoky Hill. The second regression laid down the rest of the Niobrara Group and the lower portion of the Pierre Shale. The shoreline of the regression was about 150 miles southwest of the studied area.

The alternation of shale and limestone in the Smoky Shale Formation indicates that the input of fine- to clay-size clastics fluctuated. When the input of siliclastic materials was greater than the accumulation of carbonate materials from various kinds of organisms, especially Foraminifera, shale was formed; otherwise, chalk or limestone developed.

According to Molenaar (1983), the second cycle did not transgress as far as the first cycle.

Pierre Shale Formation

Outcrop and Stratigraphic Relationships. The Pierre Shale covers the northern part of the Canon City embayment.

In the southern part, the formation is capped by the younger rocks and Quaternary alluvial deposits. The contact between the Pierre and the overlying Trinidad is transitional.

Lithology. The lower and middle Pierre Shale is a remarkably uniform body of dark gray to greenish black shale. The shale is soft at the surface of the ground; however, it is firm and fissile when fresh and unweathered. Many large, hard calcareous concretions are present about 1800 below the top of the Pierre. Some geologists regard these concretions as "teepee buttes" (Washburne, 1909; DeFord, 1929; Griffitts, 1949). The upper 200-300 feet of the formation is composed of gray, thinly-bedded (4 to 8 feet), fine-grained sandstone interbedded with thin beds (2 to 6 feet) of gray to dark gray silty and sandy shale. The sandstone and shale intertongue with and grade into the basal beds of the overlying Trinidad Sandstone.

Age. Griffitts (1949) performed a comprehensive paleontological study of the Pierre Shale and concluded that the fossil content fixes the age of this formation to be the early Montana (from 81. m.y. to 65 m.y. Late Cretaceous). He tentatively correlated the Pierre with the Mancos and Mesaverde formations of Western Colorado.

Depositional Environment. The Pierre Shale was deposited during the withdrawal of the Late Cretaceous sea from the Rocky Mountains region. The lower portion of the shale is barren and contains thin beds of bentonite and pyrite (Griffitts, 1949), indicating that in the early part of Pierre time, the sea was very deep, quiet, and reducing. However, modern studies show that a reducing and quiet environment can also be present in a shallow marine environment (Figure 54). If the lower portion of the shale were deposited in the same environment as that shown in Figure 54, this portion is probably a source rock of the Florence oil field. Therefore, microbiologic studies are needed to decide that depth of water during the deposition of the lower portion. With the fall of the sea level and a consequent better circulation of oxygen, sediments gradually become more sandy and light in color. In the top part of the Pierre Shale, the intercalcation between shales and sandstones indicates that minor oscillations of the sea water occurred in a near-shore environment.

Trinidad Sandstone Formation

Outcrop and Stratigraphic Relationships. The entire outcrop of the Trinidad Sandstone lies in the western part of the Canon City basin. This formation is conformably overlain by the Vermejo Formation.

Lithology. The Trinidad Sandstone Formation is a light-gray to yellowish-gray, friable, fine- to medium-grained, cross-bedded, and micaceous sandstone. The sandstone is massive in the upper portion and thinly-bedded in the lower portion where it grades downward into the Pierre Shale. This formation contains interlayered olive-gray

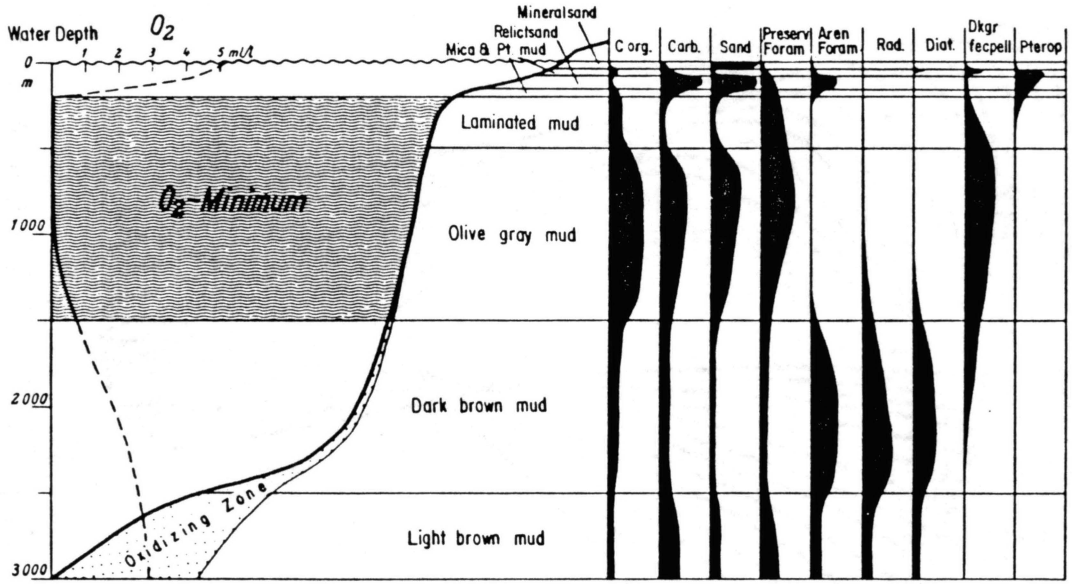


Figure 54. Section Through Oxygen Minimum Layer on the Western Shelf of India (After Steckelberg, 1972)

carbonaceous shale and dark reddish-brown calcareous sandstone concretions. Its thickness ranges from 40 to 90 feet.

Age. The Trinidad Sandstone is of late Montanan age (from 81 m.y. to 65 m.y. Late Cretaceous) and is correlated with the lower Fruitland-Kirtland sequence in the San Juan basin, New Mexico, and the Mesaverde Formation in Utah and Northern Colorado.

Depositional Environment. The bottom two-thirds of the Trinidad Sandstone are burrowed, clay-filled, and deformed penecontemporaneously. The interpretation--that these beds were deposited in a delta front--is supported by these deformation structures because they generally occur in places of high deposition rates on an oversteepened slope, such as the slope of delta front. The top one-third was formed by distributary channels over the delta plain when the delta prograded seaward. The distributary-channel-sandstone is a high angle cross-bedded, fine- to medium-grained and nonburrowed sandstone. Billingsley (1978) believed that the absence of bioturbation indicates that waters were too fresh to allow colorization by marine and brackish-water bioturbating organisms in delta front sands. The channels were low sinuosity because the top one-third does not bear clay or silt.

Vermejo Formation

Outcrop and Stratigraphic Relationship. The Vermejo Formation crops out in the southwestern part of the Canon

City embayment. Its exposure is enclosed by that of the Trinidad Sandstone. The boundary between the Vermejo and the overlying Raton Formation is unconformable.

Lithology. The Vermejo Formation is composed mainly of yellowish-orange, fine- to medium-grained, finely laminate, platy or massive beds of marine and nonmarine sandstones that vary from a few feet to 100 feet thick. Most sandstone beds are friable, shaly and lenticular. Dark to light gray sandy shale interfingers with the sandstones. The shale is blocky to flakey and a few feet to 35 feet thick. A large amount of plant debris and leaves are preserved in the shale beds (Lee, 1923). Coal and lignite beds are interbedded with the sandstone and shale. The total thickness of the Vermejo Formation ranges from 150 to 750 feet. As many as 16 coal layers are present, seven of which have been commercially important in the Canon City area.

Age. The Vermejo Formation is of late Montanan age (from 81 m.y. to 65 m.y. Late Cretaceous), and is correlated with the Upper Fruitland-Kirtland sequence in San Juan Basin, New Mexico, and the combination of Iles Formation and Price River Formation in Utah and Northern Colorado (Molenaar, 1983).

Depositional Environment. The high percentage of shale, presence of coal, lenticular nature of sandstone bodies, and cross-bedding suggest that the Vermejo Formation was deposited by high sinuous rivers on the same delta plain as the top one-third of the Trinidad. A large area of the delta plain was occupied by swamps, marshes, lakes, and crevasse splays.

It is known that carbonaceous materials can be preserved only in a chemically reducing environment, which occurs only beneath the upper limit of the water table. As the delta plain continued to subside and the subsidence rate was greater that that of the accumulation rate of clay-and silt-size detritus, an oxidizing environment above the upper limit of the water table became a reducing one. This transition from oxidation to reduction is reflected in the formation by the gradual increase in carbonaceous materials upward and by the following vertical sequence of lithological variation: noncalcareous claystone x! block, carbonaceous shale x! coal (Billingsley, 1978).

Figure 55 shows the overall deposition model of the rocks from the Pierre Shale to the Vermejo Formation. According to Stevenson (1882), the Trinidad Sandstone is more massive and strongly cross-bedded on the west side of the Cañon City coal field, indicating that a highland occupied the present position of the Wet Mountains, and that the upper Cretaceous sea covered the eastern part of the Cañon City area during the deposition of the Trinidad Sandstone and the Vermejo Formations.

Outcrop and Stratigraphic Relationships. The Raton Formation crops out as a cap of the underlying Vermejo Formation in the southwestern part of the Canon City basin, and is unconformably overlain by the Poison Canyon formation.

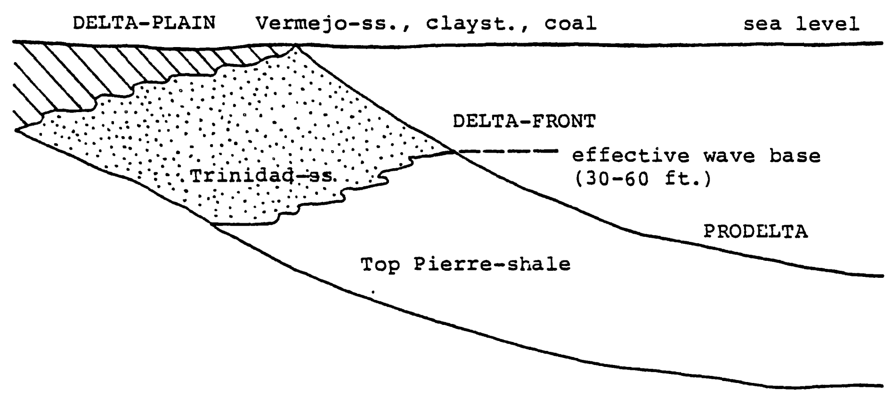


Figure 55. Depositional Environments of the Top Pierre Shale, Trinidad Sandstone, and Vermejo Formation (After Billingsley, 1978)

Lithology. The Raton Formation is yellowish-gray to yellowish-brown, medium- to coarse-grained, massive, cross-stratified, nonmarine sandstone in the lower portion. The upper portion consists of both sandstone and thin carbonaceous shale layers with sandstone dominating. The sandstone beds are thinner and softer than the lower portion beds. Some pollen of the Paleocene age was found about 150 feet below the base of the Poison Canyon Formation (Scott, 1977).

Age. The Raton Formation was deposited during the latest Cretaceous and Paleocene (Scott, 1977); it can be correlated with the McDermott Formation in San Juan Basin, New Mexico, the North Horn Formation, Utah and the top part of the Williams Fork Formation in Northern Colorado (Molenaar, 1983).

Depositional Environment. The Laramide uplift and erosion were accompanied by the accumulation of the alluvial fan deposits in the Canon City embayment. This deposition may have been caused partly by a subsidence of this area. Since marine beds are not present, the subsidence evidently was not sufficient to allow sea water to return. The fining upward sequence of the Raton Formation indicates that this alluvial fan was a retrograding one. According to Leeder (1983), a confined flow changes into a braiding system of shallow channels with a lower flow rate below the intersection point on an alluvial fan. THus, the massive cross-stratified coarse sandstones in the lower portion were most probably laid down by the above shallow channels. Later, the fan became inactive, leading to the deposition of silt-stones, shales, and carbonaceous layers. The retrogradation of the alluvial fan might have been caused by a drying climate.

Another explanation for the fining upward sequence is the possible shifting of the active deposition site to another location over an alluvial fan surface. If so, the shift left the studied inactive, and fine- to clay-size sediments were laid down in the inactive environment with subsequent growing plants and weathering generating some carbonate minerals.

CHAPTER VII

CENOZOIC ERA

Geological Setting

The Laramide Revolution, which created the present Rocky Mountains and the Canon City embayment, began in the latest Cretaceous time and lasted until the end of Eocene or possibly the end of Oligocene time. Van Tuyl and Lovering (1935) believed that the revolution had five to eight movements, plus some minor disturbances. Each movement was separated from another by a period of erosion during which mature topography developed; the youngest peneplain is of Miocene age. They stated that the present altitude of the Front Range was reached by gentle rising through the 15 to 20 million years since Miocene time. However, an investigation conducted by Wahlstrom (1947) showed the following physiographic development of the Front Range area: 1) a long period of erosion occurred after the Laramide Revolution; 2) a peneplain, which had an altitude of 5000-6000 feet above sea level, was developed by late Pliocene or possibly early Pleistocene time; and 3) a series of uplifts raised this peneplain by faulting and upwarping, and the present canyons and valleys were excavated during the uplift.

The major structure features which were formed in this embayment during the Laramie orogeny include folds, normal and reverse faults. The folds, which are mostly monoclinal, exhibit an axial trend of a dominantly north-south direction. Generally, the inclination of the steep limbs of the folds is to the east. From the crest of such a fold, the strata are usually very gently inclined to the west. Therefore, the folds are very asymmetrical.

Petrology and Depositional Environment

Poison Canyon Formation

Outcrop and Stratigraphic Relationships. The Poison Canyon Formation is only present and crops out in the southwestern corner of the Canon City embayment, covering an area of approximately 1.5 x 0.8 square miles. The formation is unconformably overlain by the Late Tertiary and Quarternary alluvium.

Lithology. The Poison Canyon Formation consists of yellowish-gray, yellowish-brown, or yellowish-orange, well-bedded claystone, siltstone, and sandstone in the upper portion, and yellowish-brown or light brown, thickly bedded, medium-to-coarse-grained sandstone and poorly sorted fluvial conglomerate in the lower portion. The conglomerate is cross-bedded and composed of clastics which are subangular to subrounded, gray to white chert, quartz, quartzite fragments, and feldspar.

Age. The Poison Canyon is very different from the underlying formations in the characteristics and coarseness of its materials. The occurrence of the rock fragments from these underlying formations in its basal conglomerate indicates that a long period of uplift (Laramide Revolution) and erosion preceded the deposition of the formation. Due to its involvement with the last great orographic movement of the Rocky Mountains, Washburne (1910) assigned the Poison Canyon Formation an Early Tertiary age and separated this formation from the Late Tertiary gravels. Therefore, it can be correlated with the Arapahoe Formation in the Denver basin.

Depositional Environment. The large scale fining upward sequence which is associated with the formation possibly indicates that the Poison Canyon Formation was deposited in a retrograding alluvial fan environment. The cross-stratified and poorly sorted conglomerate in the lower portion of the formation was laid down in the proximal fan subenvironment; the medium-to-coarse-grained sandstones, which are not present with clay-sized sediments, were deposited by braided streams over the mid-fan subunit; then, the overlying siltstone and claystone were formed over the distal-fan. Alternatively, the siltstone and claystone could also have been build-up by sheet-flood in the abandoned area from which the active deposition site was shifted into somewhere else. According to Scott (1977), channels had a general eastward trend, indicating that the Wet Mountains were the source area. Washburne (1910) found that the basal conglomerate is composed of rock fragments of Late Cretaceous to Pre-Cambrian times.

Volcanism

Epis and Chapin's (1968) studies showed that a single large composite volcanic field occurred in south central Colorado during Oligocene time and that most of the eruptive rocks were accumulated by the end of the Oligocene and covered the whole region. The initial eruptions were largely of andesitic to rhyodactic lavas and breccias. Then the characteristics of activity changed to alternating ash-flow and lava-flow eruptions. This volcanic field was fragmented into the Thirty-nine Mile, West Elk etc. subfields by block faulting and subsequent erosion in Miocene and later time. During Late Cenozoic time, a limited amount of basaltic activity took place. In the studied area, no volcanic rocks were preserved. So, it is impossible to determine the thickness of volcanic layers which once covered the sedimentary Poison Canyon Formation. However, volcanic flows and tuff are present about three miles north of Garden Park (Boos and Boos, 1957).

CHAPTER VIII

SUBSIDENCE AND THERMAL HISTORY

Basin Subsidence Analysis

Figure 56 shows the average rate of subsidence (m/m.y.) of the Canon City area. Based on this diagram, the writer divided the whole evolution into the following four stages:

Stage 1 includes the time period from the deposition of the Sawatch Sandstone to the rise of the Ancestral Rocky Mountains. The subsidence is slow during approximately the first 50 million years. Greater subsidence rates mark the following 25 million years, followed by nonsubsidence to very slow rates which lasted for about 90 million years.

Stage 2 began with the deposition of the Fountain Formation and ended after the Middle Jurassic. The subsidence rate curve is different from that of the first stage such that after peak subsidence, the rate decays slowly; however, it decreases abruptly in the first stage.

Stage 3 corresponded with the Late Jurassic to Cretaceous interval. This stage is characterized by striking subsidence rates.

Stage 4 began from the Laramie Revolution; the subsidence is a continuation of Stage 3, but the subsidence rate is greater than that of the Stage 3.

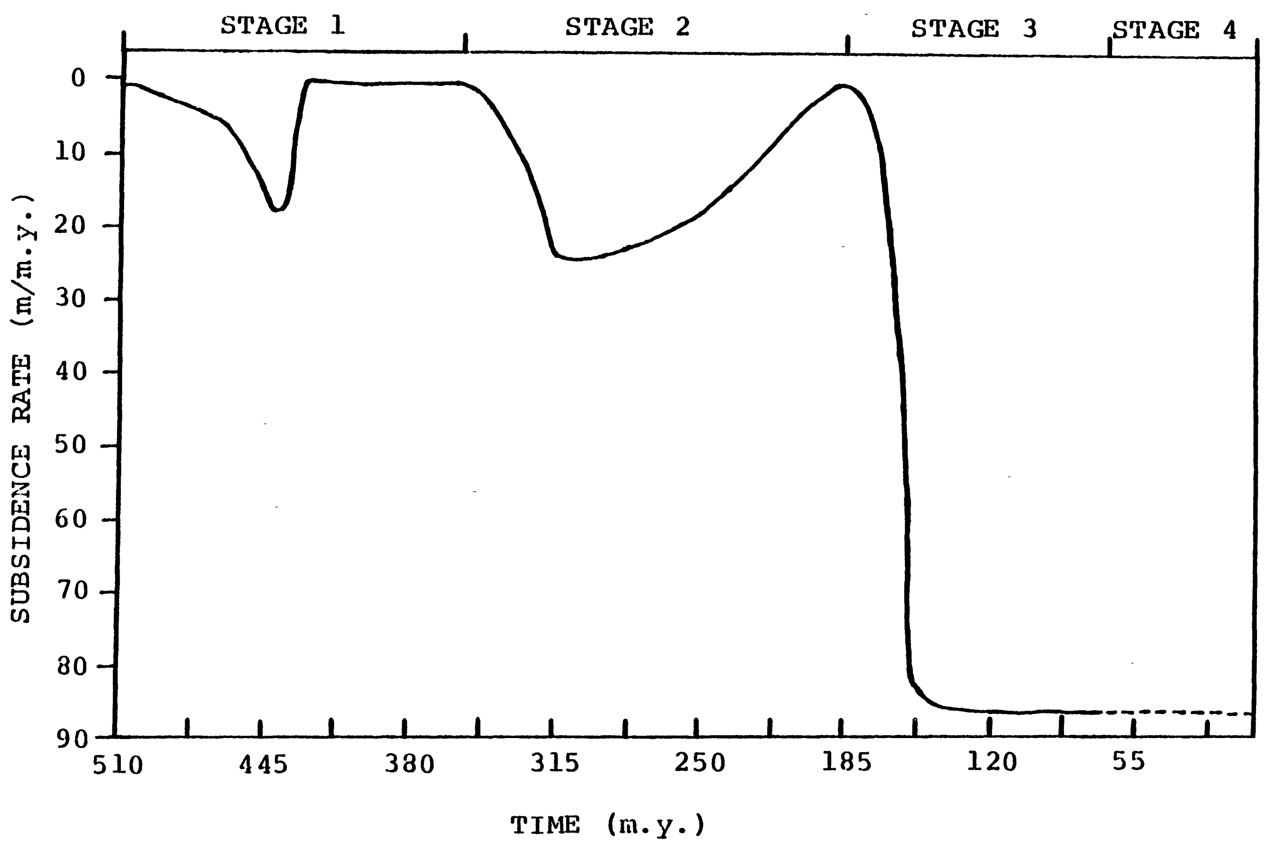


Figure 56. Average Rate of the Subsidence of the Canon City Embayment

Figure 57 compares the subsidence-rate curves of three other basins, the Michigan basin, the Southern Oklahoma aulacogen, and the downwarped platform outside of the aulacogen. The subsidence rates of the three basins decay slowly after peak values while the subsidence rate of the studied area decreases to zero abruptly after a quick subsidence, making it apparent that the curve of the subsidence rate during the first stage is not like the curves of the other three basins. However, the curve of the second stage is much the same as those shown in Figure 57, especially the curve of the Southern Oklahoma aulacogen. The subsidence of Stage 4 is still continuous and will possibly end in 125 million years, like the subsidence in Stage 2, or in a very short time (65 million years), like that of Stage 1.

Geothermal History During Subsidence and Sedimentation

High temperatures can be assumed during the early Pre-Cambrian time when the tectonic framework of the Colorado Plateau was set, and during the uplift of the Ancestral Rocky Mountains. These two thermal anomalies were gradually reduced by heat conduction to the surface from then. At the close of Cretaceous time, the Laramide Revolution lifted the Rocky Mountains and initiated igneous intrusion and volcanic eruption. Therefore, during the Tertiary time the geothermal gradient was probably very high in the Canon City embayment and adjacent areas. From then on, it decreased

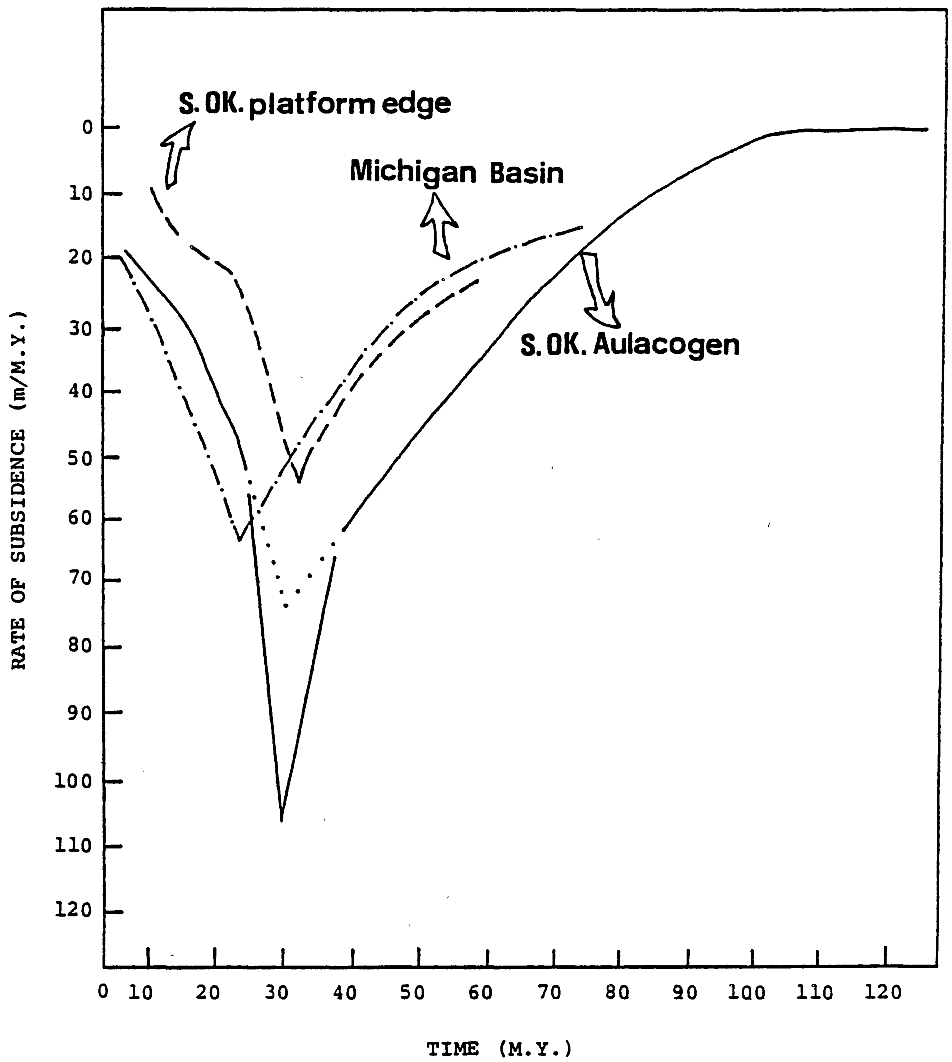


Figure 57. Subsidence Rate of the Michigan Basin, Southern Oklahoma Aulacogen, and the Down-Warped Platform Outside of the Aulacogen (After Feinstein, 1981)

gradually. Parsons and Sclater (1977) stated that during the subsidence of a basin, the geothermal gradient is not constant but decays exponentially with time until it becomes an asymptote. Turcotte and Ahern (1977) gave the following equation for the calculation of the variation of heat flow as a function of time.

$$q = km\ (Tm - To)\ \left[\frac{1}{\overline{PKmt}}\right]$$

t = the time from the beginning of subsidence

Tm = temperature of emplacement of new magma

To = surface temperature

km = thermal conductivity

Km = thermal diffusivity

By dividing the heat flow (q) by the thermal conductivity (K) of the sedimentary rocks, the writer got the estimates of the geothermal gradients at different times in the past. A thermal conductivity of $K=5x10^{-3}$ cal $cm^{-1}s^{-1}c^{-1}$, which is generally used as an average for sedimentary rocks, has been used for this calculation. Figure 58 shows the variation of the geothermal gradient with time. According to Zacharakis and Pearl (1984), geothermal gradients in 11 temperature gradient wells in the Canon City basin range from 2.17°F/100 feet (21.8°C/Km) to a high of 4.92°F/100 feet (89.7°C/Km). The highest value was measured in the Brush Hollow Anticline (east of the studied area) where deep

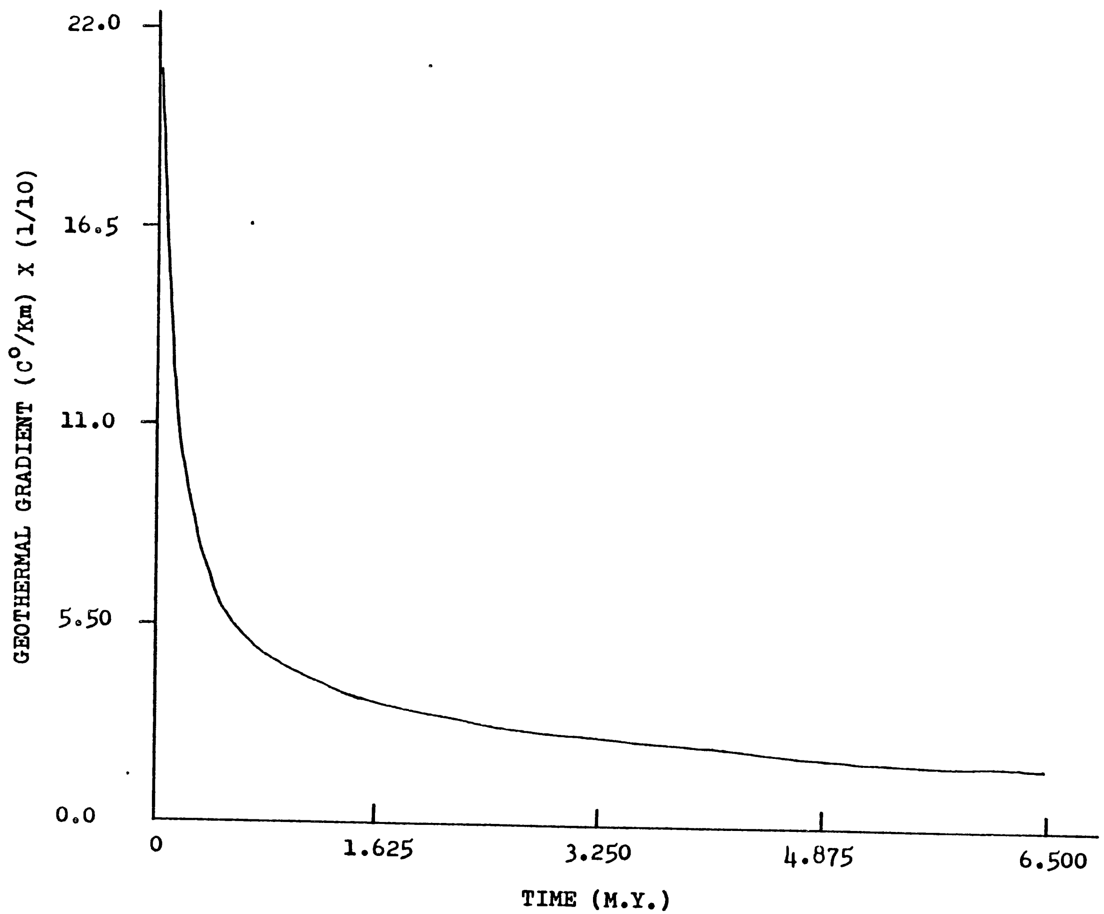


Figure 58. Variation of the Geothermal Gradient with Time in the Canon City Embayment

faults are active and conduct heat from the mantle to the surface; therefore, the calculated value of the present time (22.5°C/Km) is much the same as the lower measured ones. From the geothermal gradients and different depths of sediment burial, four isotherms which are a function of depth and time (Figure 59) have been constructed. This diagram gives valuable information about the thermal history of the sedimentary sequence in the embayment. Nearly one-half of the Pierre Shale and all of the Niobrara and the Benton groups were subjected to the "liquid window" zone between the 65°C and 150°C isotherms since the beginning of the Oligocene time to about 20 million years ago. Possibly, during this time interval (15 million years), oil was generated from the Pierre Shale and the underlying organic shales and limestones.

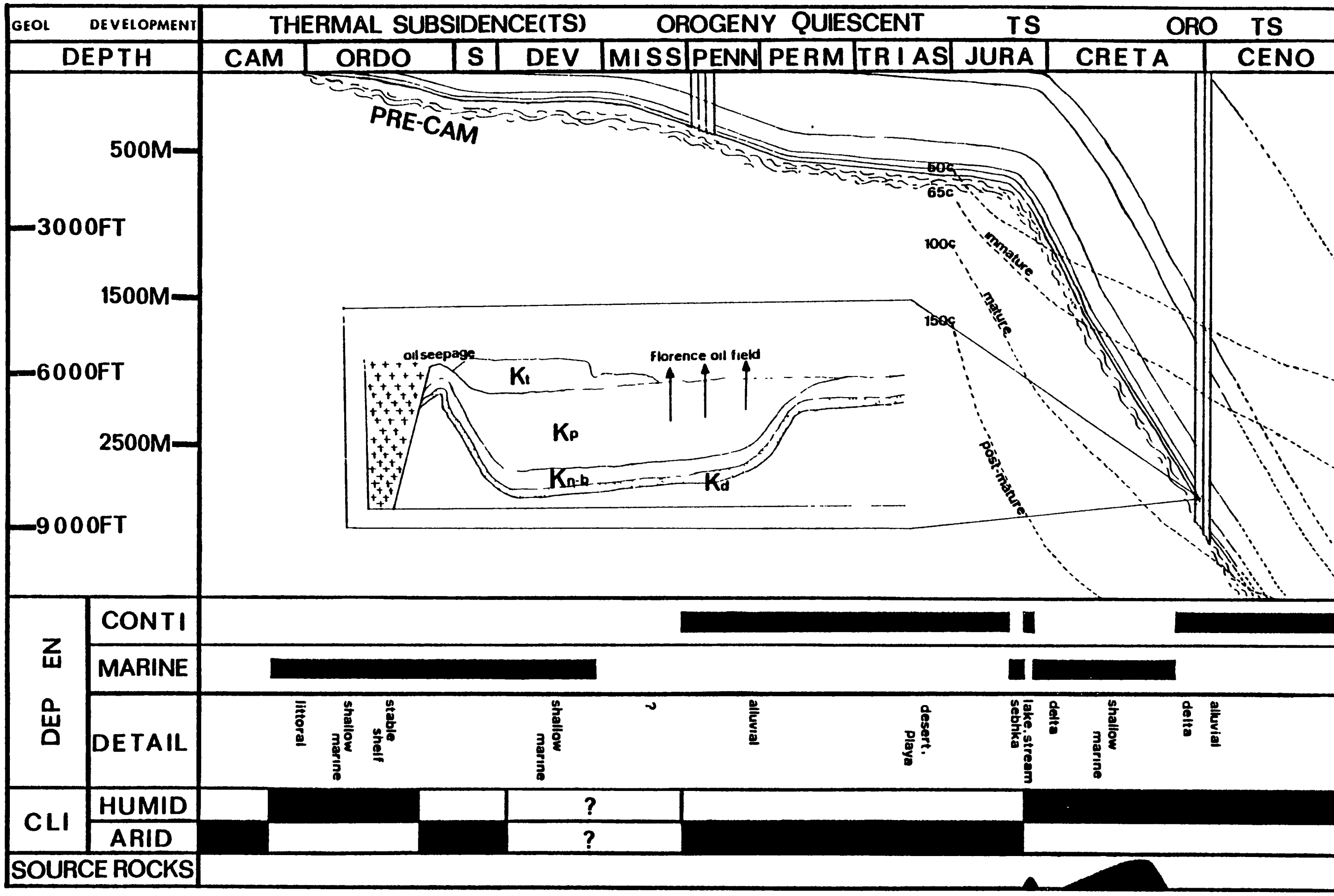


Figure 59. Summary Diagram of the Whole Development of the Canon City Embayment

CHAPTER IX

PETROLEUM GEOLOGY

The second oldest oil field west of the Mississippi River, the Florence Oil Field, is on the east flank of the Canon City embayment. In the embayment syncline, a second order syncline (Figure 59) is developed. This box-shaped syncline bounds the Florence Oil Field and brings the Niobrara Limestone to the surface. No oil was discovered outside the monocline. Washburne (1909) thought that the steep dips and faults along the steep limbs of the monocline let the oil escape along the steep bedding planes and through the faults or permitted water circulation that has dissipated the oil.

In this field, oil is preserved in fissures and cracks. Regarding the generation of the cracks, Cramer (1962) put forward the following theory: During the formation of the first and second order synclines, stresses accumulated in the less competent Pierre Shale, which is held by more competent beds--the Trinidad Sandstone and the Niobrara Limestones; when the overburden of the Trinidad Sandstone and the Vermejo Sandstone were removed by erosion, the stresses were released in the form of little anticlinal flexures, joints, fissures, or even fractures. The oil accumulated in the fractured pore spaces. However, the Teepee Buttes structure can also account for the origin of the fractures. Ronald K. DeFord (1929) found that about 2,400 feet above the Codell Sandstone, the formation contains remarkable Teepee Buttes. The Teepee Butte Reefs are composed chiefly of Lucina occidentalis embedded in a calcareous matrix. These masses might have created some of the cracked holes and fractures in the Pierre Shale.

In this embayment, the possible source rocks for the Florence Oil Field are the Pierre Shale, Niobrara Limestone, and Carlile Shale. Two factors play very important roles in the thermal maturation of these source rocks. The first is the volcanic activity around the embayment. There was a series of eruptions during the Oligocene-Eocene times. The adjacent volcanic fields are the Thirty-nine Mile Field and West Elk Field; therefore, there must have been a higher geothermal gradient than that of the present time. If the initial temperature is assumed to be 1200°C, the calculated results show that nearly all the Upper Cretaceous shales were once subjected to an "oil window" (Figure 59). The second factor is the Laramide Revolution. This revolution uplifted the present Rocky Mountains, generating and rejuvenating some deep faults. These faults conduct heat in the mantle to the lithosphere and surface. Clayton and Swetland (1980) studied the vitrinite reflectance in the Denver basin. They found that the vitrinite reflectance reaches very high values along the foothill belt of the Front Range adjacent to the Denver basin and decreases away from the belt toward the east. Geographically and geologically, the Canon City embayment lies well within the foothill belt; thus, a high vitrinite reflectance is expected in the studied area. Even though almost all the Upper Cretaceous shales and limestones gave out oil when they were mature, the non-porous Greenhorn Limestones might prevent the upward migration of the oil derived from the Niobrara and Benton into the fractures of the Pierre Shale; therefore, the oil in the Florence Oil Field is mainly from the Pierre Shale.

The Denver basin, which is separated from the Canon city basin by the Red Creek Arch formed during the Laramide Revolution, produces a huge amount of oil and gas. Most of the petroleum was generated from the Carlile Shale, the Greenhorn Limestone, and the Graneros Shale. Since the Canon City Basin has received nearly the same Cretaceous sediments, and the organic materials in the sediments were once under a high temperature condition (65°C-150°C), there should be a significant amount of oil and gas below the Pierre Shale. The oil and gas were derived from the Benton and Niobrara. The following observations indicate the migration of oil through the Morrison and Dakota Sandstones: (1) oil seeps issue from both the Morrison Sandstones and the Dakota Sandstones in several locations around the embayment, especially in the Oil Creek area about six miles north of Canon City; (2) Organic matter (bitumen) was found in some thin sections of the Morrison and Dakota; (3) the three carbonate cements, calcite, dolomite, and siderite, were dissolved in an acidic environment. Most likely, the acids were created during the maturation of organic materials and the generation of hydrocarbons and then migrated through the sandstones before the migration of oil. Unfortunately, no deep wells penetrating the Dakota and Morrison Formations have been drilled in the embayment. Thus, future exploration should be conducted, and deep wells may be prospected if the distribution of sandstone bodies is well understood.

CHAPTER X

CONCLUSIONS

Information obtained from this study (Figure 59) permits the following conclusions to be drawn:

(1) The Pre-Cambrian tectonics, the movement of the Ancestral Rocky Mountains, and the Laramide Revolution played very important roles in the development of the Canon City embayment. Deep faults formed during Pre-Cambrian time were active until Laramide time; the orogenic uplift of the Ancestral Rockies caused the accumulation of red beds (Fountain and Lykins formations); the Laramide Revolution created the present Rocky Mountains and the Canon City embayment.

(2) The subsidence of the studied area during the Pennsylvanian-Permian interval is much like that of the Southern Oklahoma aulacogen, the downwarped platform outside of this aulacogen, and the Michigan basin; however, the subsidence during other times is different.

(3) The Ordovician rocks represent a unique Early Paleozoic record in the embayment and thus reflect a period of tectonic stability; the absence of Silurian, Lower-Middle Devonian, and Mississippian rocks indicate the intensive pre-Pennsylvanian erosion while the absence of Triassic and

Lower-Middle Jurassic rocks suggests the occurrence of a positive land during this interval. During the Late Jurassic, the Ralston Creek and Morrison Formations were deposited in sebkha and alluvial streams, respectively. The following Cretaceous time is characterized by a huge accumulation of both clastic and carbonate sediments. The Cenozoic era is represented by alluvial fan depositions (the Upper Raton and Poison Canyon) and volcanism which began in Oligocene time and ended in Miocene time.

(4) The dissolution of matrix such as that in the Fountain Formation and the dissolution of gypsum such as that in the Ralston Creek Formation area can create a significant amount of secondary porosity.

(5) The Lytle and Dakota Sandstone is an excellent reservoir rock, with secondary porosity varying from 10% to 20% and a good connection of pore spaces.

(6) Organic materials in the Morrison Formation, Carlile Shale, and upper Cretaceous rocks were once subjected to an "oil window" (65°C to 150°C). The source rock of the Florence Oil Field is the Pierre Shale.

(7) The comparison of the Canon City embayment with the Denver basin indicates that oil is present in the Morrison and Dakota sandstones, both of which have not yet been drilled in this area.

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