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Green River Formation, Wyoming: A Playa-Lake Complex

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ABSTRACT

Recent observations in the Green River Formation suggest that ancient "Lake Gosiute" was a playa-lake complex (Eugster and Surdam, 1973). In this paper, the new playa-lake model is tested in a basin-wide study of surface and subsurface observations. The rocks deposited in and around "Lake Gosiute" can be divided into three distinct facies: (1) marginal silt and sand, (2) carbonate mud flat, and (3) lacustrine. Each lithologic facies has a characteristic carbonate mineral assemblage. The marginal facies is characterized by calcite concretions and calcareous cements. The mud-flat facies is characterized by calcite and (or) dolomite. The lacustrine facies is characterized either by trona (sodium carbonate) or by oil shale (either calcitic or dolomitic). The regional distribution pattern of lithologic facies and mineral zones in the Green River Formation of Wyoming is identical with that of modern playa-lake complexes. Moreover, in the Tipton Shale Member, once supposed to have been deposited in a large, deep, open, fresh-water lake (Bradley, 1963), there is strong evidence demonstrating large fluctuations in the position of the shoreline and progressive increases in salinity and alkalinity of the lake water. By mapping the regional distribution and types of lateral changes characterizing individual stromatolite units, the fluctuations in shoreline position can be quantified. The vertical distribution of fossils and ooliths in the Green River Formation allows an evaluation of water chemistry. In addition, the assemblage of sedimentary structures in the Tipton Shale Member is compatible only with a sedimentologic model characterized by shallow-water deposition and frequent subaerial exposure.

Thus, the deep-water stratified-lake model is untenable not only for the Wilkins Peak Member but also for the Tipton Shale Member of the Green River Formation. In contrast, the playa-lake model is consistent with the physical, chemical, and paleontologic aspects of the Green River Formation of Wyoming. *Key words: sedimentary petrology, Lake Gosiute, Eocene, playa, lake, Tipton Shale Member.*

INTRODUCTION

The Green River Formation deposited in and adjacent to Eocene Lake Gosiute is perhaps the most famous accumulation of lacustrine sediments in the world. Ever since the classic work of Bradley (1929, 1931), it has been under almost continuous scrutiny by geologists. The most modern work is documented by Bradley (1963, 1964), Bradley and Eugster (1969), Culbertson (1966, 1971), Surdam and Parker (1972), Eugster and Surdam (1973), and others. Current interest in the Green River Formation has centered, for economic reasons, on trona and oil shale. The discovery of Magadi-type cherts in the Tipton Shale and Wilkins Peak Members, together with the observations of abundant sedimentary structures indicating subaerial exposure, has led to the suggestion that most of "Lake Gosiute" was in fact a playa-lake complex (Eugster and Surdam, 1971, 1973), rather than a stratified lake, as was postulated by Bradley and Eugster in 1969.

The main thrust of this paper is the evaluation of the sedimentary environment of the lower Wilkins Peak and Tipton Shale Members of the Green River Formation (Fig. 1). The Wilkins Peak Member represents a complicated minimum stand of Lake Gosiute during which time there was a large accumulation of organic-rich alkaline-earth carbonate and trona, as the lake presumably varied from "fresh" water to hypersaline conditions (Deardorff and Mannion, 1971). Alternatively, the Tipton Shale Member represents a complex maximum stand of Lake Gosiute, during which an enormous amount of oil shale was deposited (Bradley, 1963). It would appear that Lake Gosiute is an ancient lake without modern analog.

ESSENTIAL PROBLEMS

The standard model for Lake Gosiute has been a large inland lake, which was never less than 4,000 mi² in area (Bradley and Eugster, 1969). According to this model, the lake was stratified with inferred shoals or bars, facilitating the concentration of brine in local segments of the lake, which eventually precipitated trona. The huge quantity of alkaline-earth carbonates so characteristic of the Wilkins Peak Member was thought to result from stratification of the lake by acquisition of either a perennial or seasonal mixolimnion that was rich in Ca^{+2} and HCO_3^- (Bradley and Eugster, 1969). It was believed that only such a chemically stratified lake with a mixolimnion could have provided water of a salinity low enough to support the algal growth necessary for the formation of

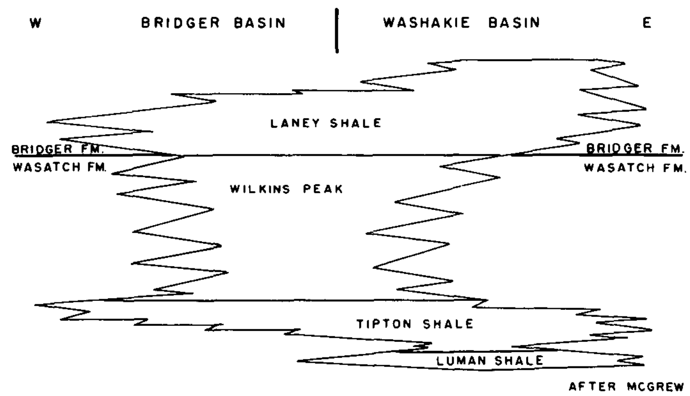

Figure 1. Generalized stratigraphy of Green River Formation in Bridger and Washakie basins of Wyoming.

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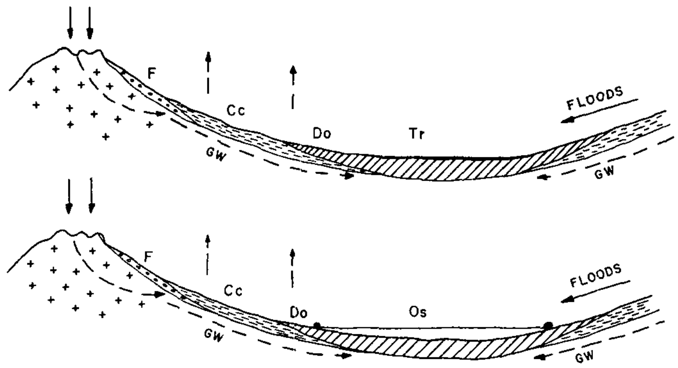

Figure 2. Hydrology, mineralogy, and brine evolution according to playa-lake model at Lake Gosiute for low stage (top) and high stage (bottom). Solid arrows = rain water; crosses = ingenous and metamorphic rocks; F = alluvial fans; dashed pattern = sandstones; diagonal pattern = carbonate muds; GW = ground-water circulation path; Cc = calcite precipitation; Do = dolomite precipitation; Tr = trona; Os = oil shale; black circles = algal beds; vertical dashed arrows = evaporation. Schematic diagram modified after Eugster and Surdam (1973).

oil shales (Bradley and Eugster, 1969). Algae contributed the organic matter present in oil shale, and trona formed when the mixolimnion evaporated.

As pointed out by Eugster and Surdam (1973), this model has some very serious and perhaps fatal flaws. Briefly, it does not explain the following aspects of the lake regime: Ca-Mg supply and Mg dominance; the persistent occurrence of dolomitic oil shales at the base of trona beds; the absence of modern analogs of Lake Gosiute; and an assemblage of sedimentary structures that indicate subaerial exposure and traction-transport of many marlstones and dolostones.

Eugster and Surdam (1973) have presented an alternate model for the deposition of the Green River Formation. They proposed that oil shales and trona beds accumulated in shallow closed-basin lakes that were fringed by large playa flats. In these playa flats, alkaline brines evolved through evaporation and precipitation of calcium carbonate and protodolomite in the capillary zone above the ground-water table. Dolostones, marlstones, and calcareous and siliciclastic sandstones were the products of occasional floods on the playa. The essential features of the playa-lake model are summarized schematically in Figure 2. For a detailed description of the playa-lake model, the reader should refer to Eugster and Surdam (1973).

The playa-lake model appears to be more compatible with the characteristics and chemical constraints inherent to the Green River Formation. This paper reports on a test of this new model in a basin-wide integrated study of both surface and subsurface observations (Fig. 3). Particular attention will be given to the Tipton Shale Member of the Green River Formation because the Tipton in particular was thought to have been deposited in a large, open, fresh-water lake (Bradley, 1964).

LAKE GOSIUTE

The Green River basin extends for approximately 150 mi from the Wyoming Range–Overthrust belt area eastward to the Rawlins uplift. It is bounded on the north by the Gros Ventre Range, the Wind River Range, and the Granite Mountains, and on the south by the Uinta Mountains and their eastward extension (Fig. 3).

Bradley (1963) estimated that the hydrographic basin of Lake Gosiute had an area of about 48,500 mi^2. The rocks exposed in the drainage basin consisted of Precambrian granite and Paleozoic and Mesozoic sedimentary rocks; no older chemogenic deposits were present. The lake was situated approximately 1,000 ft above sea level (MacGinitie, 1969). It developed on a broad, nearly featureless alluvial plain that had an original dip of less than "1 or 2 feet per mile except in a rather narrow belt adjacent to the mountains" (Bradley, 1964, p. A16).

MacGinitie (1969) made a comprehensive study of the flora of the Green River Formation and concluded that the climate was warm temperate to subtropical, with annual temperatures in the range of 60° to 70° F, with a lack of frost, and with high equability. The average annual precipitation at lake level was probably 24 to 30 in., being somewhat higher (45 in. or more) in the watershed. Maximum rainfall probably occurred in late spring and early summer and was followed by diminishing rainfall and near-drought conditions in the late summer and fall. The presence of high mountain ranges encircling a broad basin floor with low relief, the large area of the watershed and lack of a well-documented basin outlet, and the pronounced seasonal aridity of the climate all suggest that playa-lake conditions could have existed in the Green River basin during Eocene time.

The initial Green River sediments were deposited in lakes, ponds, and swamps that developed on the subsiding Wasatch plain. They were common in an east-west depression along the Uinta Mountain front and in a large area east of the present Rock Springs uplift (Roehler, 1965). The continued uplift of the surrounding mountainous areas enhanced the structural downwarping of the basin, and these smaller bodies of water coalesced to form a relatively large, shallow lake. This early stage of Lake Gosiute is represented by the Luman tongue (or Member) of the Green River Formation (Figs. 1 and 4). The Luman rocks consist of low-grade oil shale, coquinal limestone, sandstone, shale, and coal beds, which were deposited under shallow lacustrine and paludal conditions. The Luman Stage of Lake Gosiute extended from the southern Bridger basin northeastward across the site of the present Rock Springs uplift into the Great Divide and Washakie basins (Fig. 4; Roehler, 1965). At the close of Luman time, climatic (or perhaps structural) changes in the basin resulted in a decrease in the size of Lake Gosiute and a corresponding increase in fluviatile and red-bed deposition throughout most of the basin.

This restricted stage was short-lived and soon Lake Gosiute expanded greatly. The rocks deposited during this stage constitute the Tipton Shale Member of the Green River Formation (Fig. 1) and consist primarily of oil shale and dolostone.

With the onset of more arid conditions, the lake was reduced to a very small area. By this time, the lake waters had become so concentrated that trona and sometimes halite precipitated. Periodically, during less arid times, the lake level rose and beds of oil shale were deposited. The alternation of oil shale and trona deposition occurred at least 60 times during this interval (Eugster and Surdam, 1973). These rocks, together with the marlstones and mudstones spatially associated with them, constitute the Wilkins Peak Member of the Green River Formation (Fig. 1).

At the close of the Wilkins Peak time, the climate became more humid and a large, shallow lake again developed. During this stage, the siltstones, marlstones, and sandstones of the Laney Shale Member of the Green River Formation were deposited (Fig. 1). At this time Lake Gosiute and the related mud flats surrounding it may have covered an area of approximately 15,500 mi^2 (Bradley, 1963). Lake Gosiute came to an end during middle Eocene time as the basin filled with sediment.

STRATIGRAPHY OF THE TIPTON SHALE MEMBER

The stratigraphy of the Wilkins Peak Member of the Green River Formation is well documented (Culbertson, 1961, 1966, 1971; Stuart, 1963; Bradley, 1964). For the purposes of this paper, however, a few comments concerning the stratigraphy of the Tipton Shale Member are necessary. In general, the Tipton sequence can be subdivided into three distinct zones (see Fig. 5).

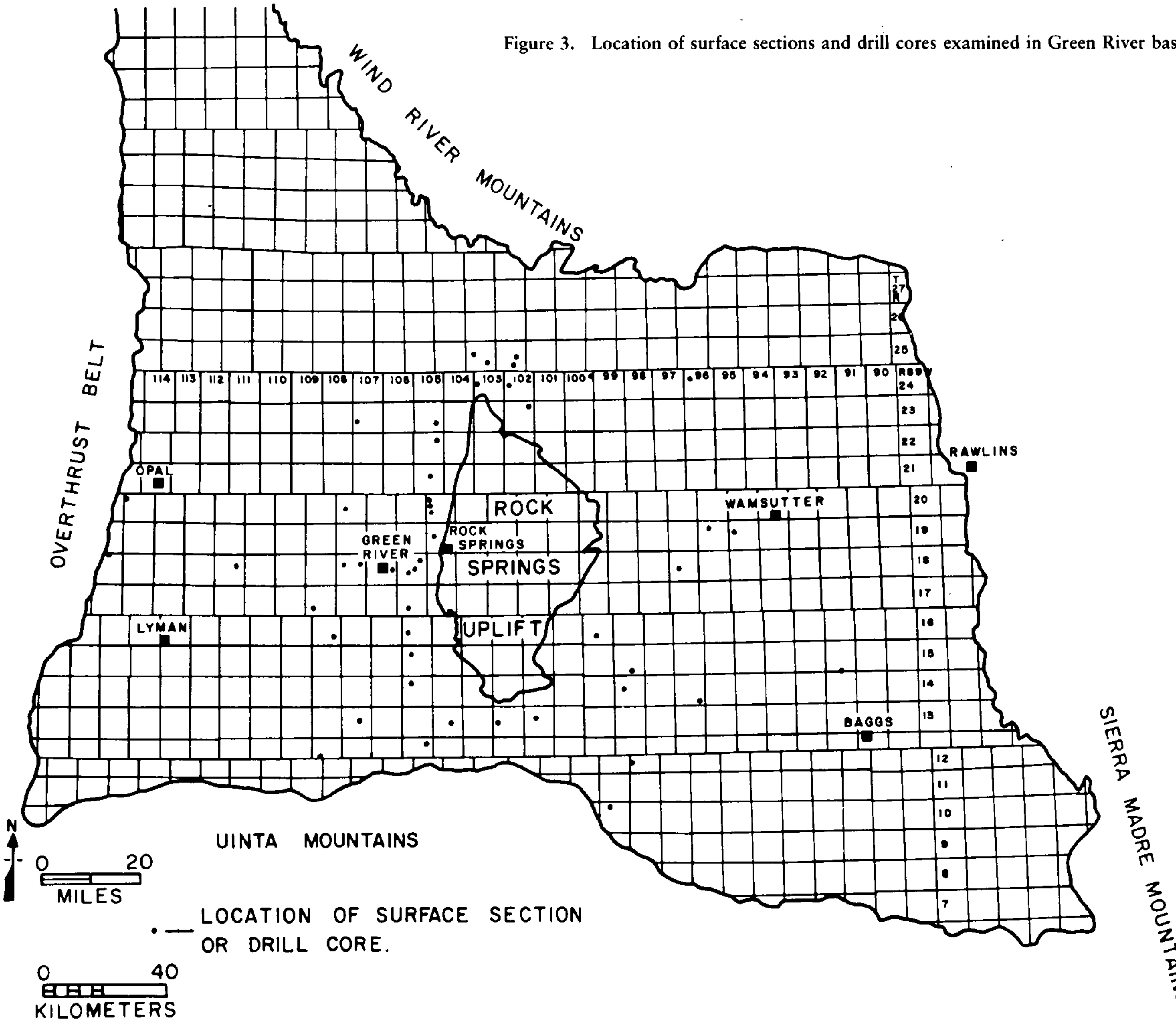

Figure 3. Location of surface sections and drill cores examined in Green River basin.

Zone 1

The lowest 1.5 to 31 m of the Tipton Shale Member consists of limestone, siltstone, and low-grade oil shale. At or very near the base of the Tipton are one or more thin limestone or sandstone units, which contain an abundance of the snail *Goniobasis* sp. These beds are collectively referred to as the *Goniobasis* marker unit. Ostracodes, the snail *Viviparus* sp., and the clam *Lampsilis* sp. can also be found in this unit. Above the *Goniobasis* marker unit of Zone 1 is primarily shale, siltstone, and low-grade oil shale.

It is interesting to note the geographic occurrence of *Lampsilis* sp. in the *Goniobasis* marker unit (Fig. 4), because La Rocque (1960) has stated that this type of clam requires the presence of fish to complete its life cycle. The larvae attach themselves to the gills of fish where they live for a considerable period of time. Therefore, the clams must live in an environment that is also favorable for the fish. This suggests that the geographic area characterized by the presence of *Lampsilis* sp. in the *Goniobasis* marker unit in the lowermost Tipton represents established lacustrine conditions. Outside the *Lampsilis* sp. area, *Physa* sp. sometimes can be found. The presence of this lung breather signifies very shallow water and emergent vegetation.

At or very near the top of Zone 1 is an algal stromatolite that varies in thickness from 2.5 to 62 cm (Fig. 5). This unit can be traced from the Church Buttes uplift eastward to the Green River and from there into the Great Divide and Washakie basins.

Zone 2

Above Zone 1 is a 19- to 50-m-thick interval that consists of oil shale and tongues of fluviatile sandstone (Fig. 5). These rocks contain very few fossils. Rare occurrences of fish remains have been found in the oil shale in the southern Green River basin, whereas the scattered remains of ostracodes and insect larvae have been noted elsewhere; most generally, however, fossiliferous material is absent. The greatest concentration of oil shale in this zone is found in the southern Bridger basin and in the southwestern Washakie basin. In general, these oil shales average 15 to 25 gal/ton.

During deposition of Zone 2 in the northern Bridger basin, oil shale deposition was interrupted by the influx of a large body of sand that entered the basin from the north. This fluviatile tongue is the easternmost extension of the New Fork tongue of the Wasatch Formation (Fig. 9). The sandstone is characterized by an assemblage of sedimentary structures diagnostic of shallow-water and

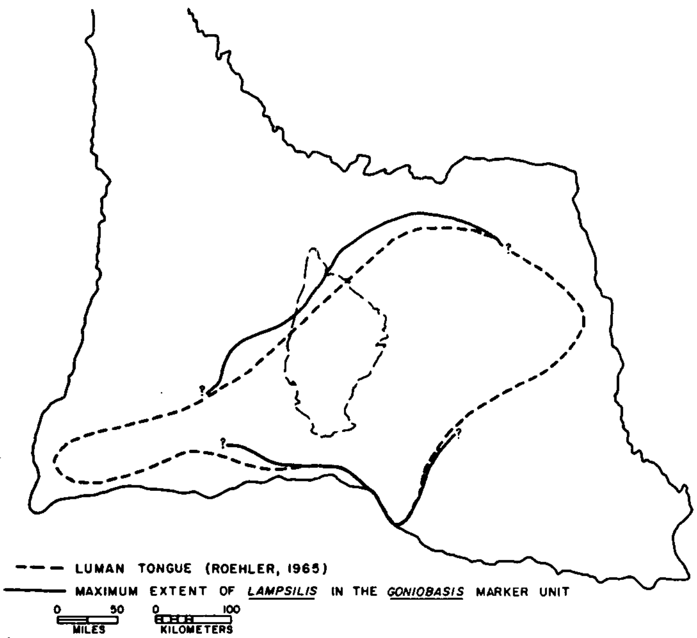

Figure 4. Distribution of Luman tongue (or Member; from Roehler, 1965). Also represented is maximum extent of *Lampsilis* sp. in *Goniobasis* marker bed showing distribution of lacustrine conditions in lower Tipton Member.

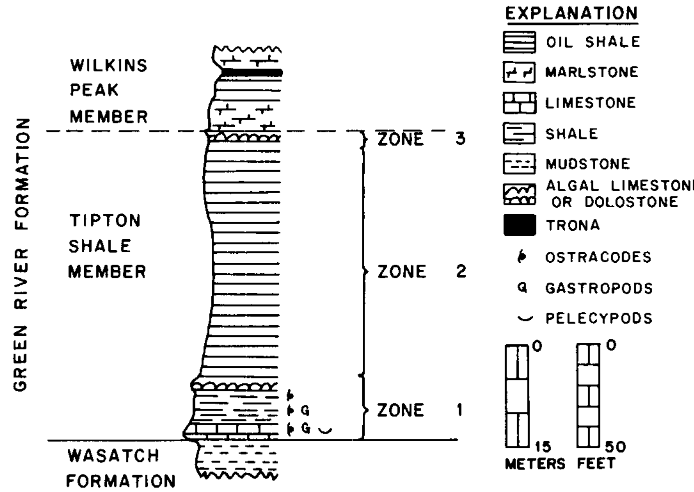

Figure 5. Schematic stratigraphy of Tipton Member of Green River Formation. Zone 1, 5 to 100 ft thick; Zone 2, 60 to 160 ft thick; Zone 3, 3 to 20 ft thick. Lowermost limestone beds represent "*Goniobasis* marker."

subaerial depositional conditions. Near the close of Zone 2, the influx of sand diminished.

Zone 3

Zone 3 marks the transition from Tipton Shale to Wilkins Peak Members of the Green River Formation (Fig. 5). Zone 3 is characterized by several beds of algal stromatolites that can be traced from the northern Bridger basin, around the northern end of the Rock Springs uplift, and into the Great Divide and Washakie basins. These same units may be correlative with the algal beds described by Lawrence (1962) along the extreme western edge of the Bridger basin near the boundary between the Fontenelle tongue (=Tipton Shale Member) and the Middle tongue (=Wilkins Peak Member) of the Green River Formation. The Wilkins Peak Member overlies the algal beds of Zone 3.

PLAYA-LAKE MODEL

It has been suggested that a vastly expanded Deep Springs playa without high sulfate content is probably the best modern analog for Lake Gosiute (Eugster and Surdam, 1973). In order to evaluate the playa-lake model, it is essential to understand thoroughly the physical and chemical processes characterizing a modern playa environment such as Deep Springs Valley. Fortunately, the hydrology and mineralogy of the Deep Springs playa have been well documented by Jones (1965). A summary of the more salient features of the hydrology and hydrochemistry of Deep Springs playa is shown in Figure 6.

Figure 7 is a schematic diagram of the mineral zones resulting from the hydrochemistry at Deep Springs playa (Jones, 1965). The saline crusts (sodium salt zone of Fig. 7), which occupy the central part of the playa, are largely lacustrine rather than efflorescent in origin. Surrounding the saline crusts are calcite- and dolomite-rich muds. Alkaline-earth carbonate may make up as much as 70 percent of the surface muds. Surrounding the carbonate muds are marginal silts and sands. The deposits of the Deep Springs playa can thus be divided into three facies (see Fig. 7): (1) lacustrine (sodium salts), (2) mud flat (carbonates), and (3) marginal (silts and sands).

COMPARISON OF THE ANCIENT LAKE GOSIUTE WITH MODERN DEEP SPRINGS PLAYA

In a similar fashion to Deep Springs playa, the rocks deposited in and around Lake Gosiute can be divided into three distinct facies: (1) marginal silts and sands, (2) mud flat, and (3) lacustrine (including trona and oil shale). The marginal sand and silt facies is particularly well illustrated in the Tipton Shale Member in the Essex Mountain area, approximately 30 mi north of Rock Springs. The sandstones and siltstones there are characterized by mud cracks, ripple marks with flattened crests, ripple marks with mud cracks in the troughs, numerous burrows and root casts, thinly bedded units with current lineations, and fluviatile channel deposits. Algal bioherms, algae-encrusted logs, ostracode coquinas, pulmonate gastropods, caliche, and the bones of terrestrial vertebrates are also characteristically associated with the marginal sandstones and siltstones. The general sequence in this facies is shown in section 1 of Figure 8.

The mud-flat facies consists of clastic lime silts and sands, stromatolitic limestone, oolitic and pisolitic limestone, dolostones, a few siliciclastic sandstones and siltstones, flat-pebble conglomerates, and some oil shale. The assemblage of characteristic sedimentary structures includes cross-bedding and burrows in the siliciclastic sandstones, cross-bedding and graded-bedding in the clastic lime silts and sands, and mud cracks. The dolostones are characterized by mud cracks (including many incipient cracks or incomplete polygons), and salt crystal casts (including trona sprays). Most of the dolostones are very fine grained, but in some of the silt-size dolostones cross-bedding and graded bedding can be discerned. Stromatolitic limestones and algal encrusted logs are also common in this facies.

This mud-flat facies can be divided into two parts, depending on whether the carbonates are nearer to the lacustrine facies or to the marginal facies. Those carbonates nearer to the marginal facies can best be described schematically (see section 2 of Fig. 8), but generally they are calcite rich and consist mainly of lime silts and sands. The carbonates spatially nearer to the lacustrine facies are mainly dolostones and are shown schematically in section 3 of Figure 8.

The lacustrine facies consist mainly of trona, oil shale, Magadi-type chert, and some marlstone (Eugster and Surdam, 1973). Some limestones in this facies, particularly at the base of the Tipton Shale

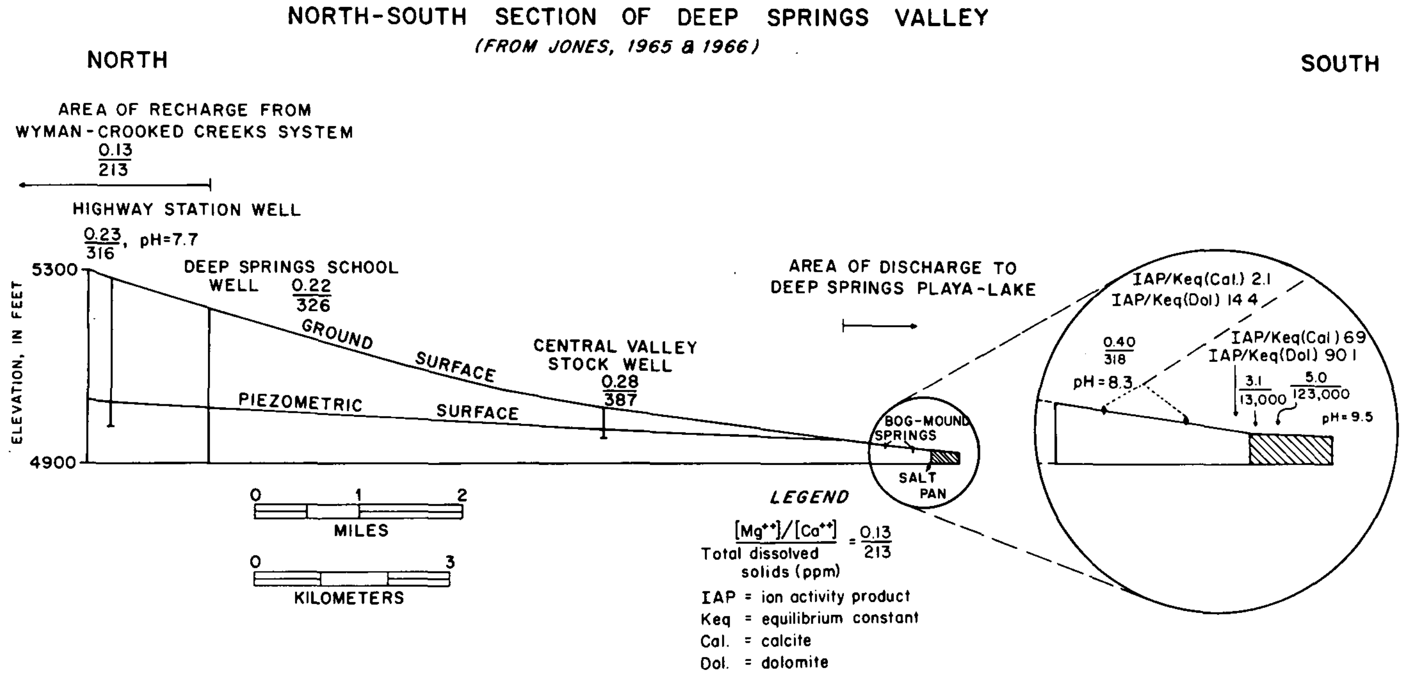

Figure 6. Hydrochemistry of modern playa-lake complex. Note evolution of brines from north to south as they migrate from alluvial deposits to playa. Diagram constructed from Jones (1965, 1966).

Member, are rich in mollusc and ostracode shells (see section 4 of Fig. 8 for a typical lacustrine section).

One of the most striking comparisons between Lake Gosiute and modern playa-lake complexes, such as Deep Springs, is the regional distribution of the lithologic facies. Not only are the same lithologic facies present in both, but the regional patterns are identical. The regional distribution pattern consists of a series of concentric bands; a central lacustrine facies is surrounded by the mud-flat facies, which, in turn, is surrounded by the marginal facies. The facies distribution pattern is independent of water depth in the playa lake. Figure 9 represents a period during middle Tipton time when Lake Gosiute was at a so-called high stand. Note that a central lacustrine facies — in this case, oil shale — is surrounded by the mud-flat facies, which, in turn, is surrounded by the marginal facies. Alternatively, Figure 10 represents a period during middle Wilkins Peak time when Lake Gosiute was at a low stand. Here, a central lacustrine facies — in this case, trona — is surrounded by the mud-flat facies, which, in turn, is surrounded by the marginal facies.

Each of the lithologic facies has a characteristic carbonate mineral assemblage. The marginal sands and silts are characterized by calcite concentrations and some calcareous cements. The mud-flat facies is characterized by calcite and (or) dolomite, depending on the proximity of the other two facies. The inner band of the mud-flat facies is predominantly dolomitic, whereas the outer band is predominantly calcitic. The lacustrine facies is characterized either by trona (sodium carbonate) or by oil shale (calcitic or dolomitic). The distribution of these characteristic carbonate mineral assemblages is what one would expect from a playa-lake complex (Eugster and Surdam, 1973; Jones, 1965).

DYNAMIC PLAYA

A closed-basin playa-lake complex is a dynamic feature. The area of the lake, depth of water, and salinity vary greatly, according to seasonal inflow and evaporation (Langbein, 1961). Figure 11 shows the type of short-term variations in inflow experienced by several modern closed basins in the western United States (Harding, 1965). In addition, there are many examples of longer term

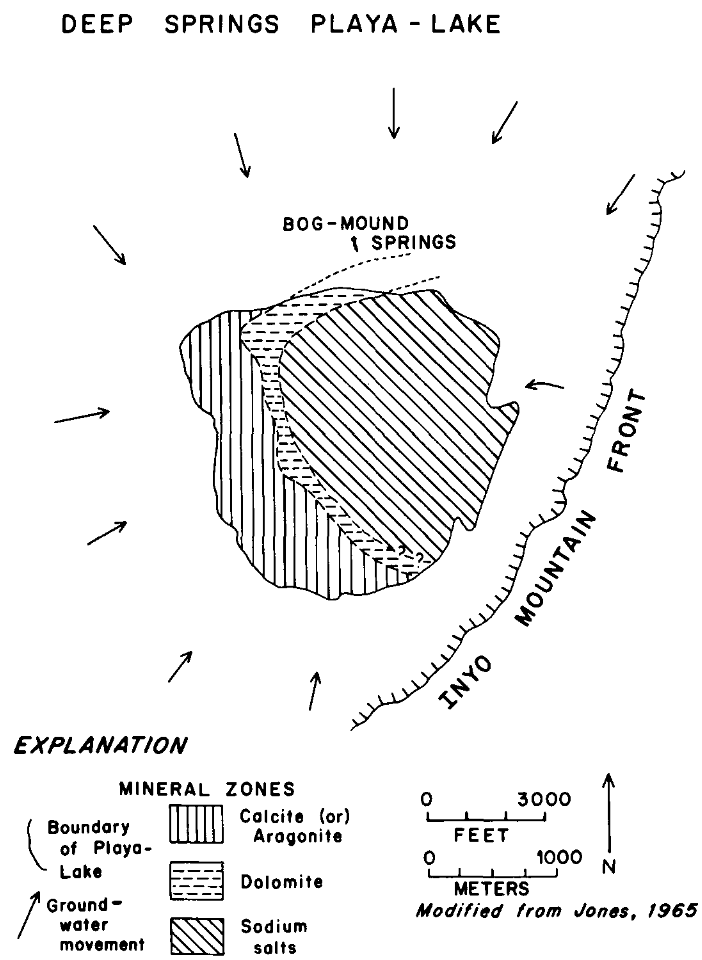

Figure 7. Distribution of mineral zones at modern Deep Springs playa, California (Jones, 1965). Playa-lake deposits surrounded by marginal silts and sands.

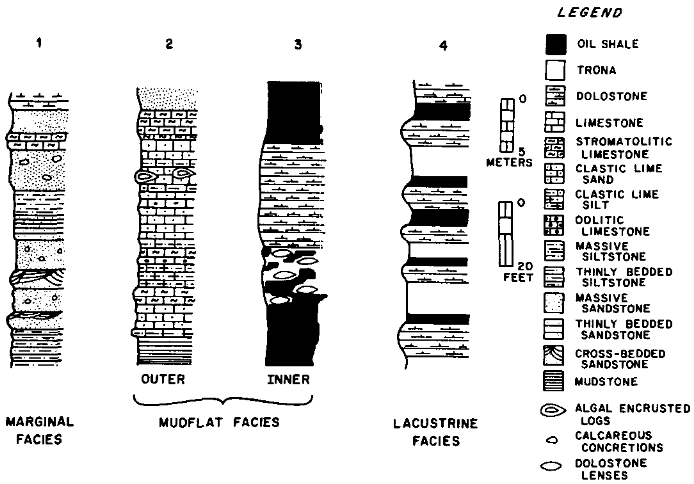

Figure 8. Schematic stratigraphic sections for marginal facies (sec. 1), mud-flat facies (secs. 2 [outer] and 3 [inner]), and lacustrine facies (sec. 4) of Green River Formation.

variations in closed basins. For example, the level of Lake Rudolf, Kenya, has fallen approximately 46 m in the last 3,000 yr as a result of climatic factors (Fuchs, 1939). During the period 1904 to 1930, the average net decrease in lake level was 23 to 30 cm/yr. (Fuchs, 1939). Nonetheless, the balance between inflow and evaporation is a delicate one, and Fuchs (1939) estimated that an increase of 12.7 cm/yr in the rate of precipitation would be sufficient to cause a rise in the level of the lake.

It is reasonable to expect that closed-basin sedimentation should be especially affected by both short- and long-term variations, particularly with respect to inflow fluctuations. In other words, playa-lake deposition should be characterized by cyclic variations, as the sedimentation regime adjusts to long- and short-term fluctuations. An outstanding example of the type of cyclic variations expected from playa-lake sedimentation is the alternating beds of trona, marlstone, and oil shale in the Wilkins Peak Member (Fig. 12). In contrast, the sedimentation regime of a stable, chemically stratified, and presumably deep lake should be characterized by a lack of variation and by gradual transitions.

Bradley (1963), in his study of the Green River basin, inferred that Lake Gosiute had an outlet at the eastern end of the Uinta Mountains during the Luman, Tipton, and Laney Stages. This inference appeared necessary when these stages were envisioned as representing large, deep, stable lakes. However, after a detailed study of the stratigraphy, sedimentology, and mineralogy of the Tipton Shale Member, it is now proposed that the depositional history of the Tipton Stage of Lake Gosiute is more consistent with the closed-basin playa-lake model. If this latter interpretation is correct, there should be evidence in the Tipton rocks of the following: (1) repeated fluctuations in lake level, (2) increasing salinity and alkalinity of the lake water, and (3) sedimentary features indicating desiccation and subaerial exposure.

Shoreline Fluctuations

Algal reefs (stromatolitic limestone units) in the Tipton Shale Member were recognized and first described in detail by Bradley (1929). He noted (Bradley, 1929) that these algal beds are important because they denote very shallow water and the proximity of shore. The stromatolite units are intimately associated with rocks containing mud cracks and other sedimentary structures indicative of subaerial exposure or shallow-water deposition. Both the external morphology and the internal fabrics of the algal stromatolites in the Green River Formation are very similar to Holocene cryptalgal fabrics and structures at Shark Bay described by Logan and others (1974).

The two stromatolitic algal horizons in the Tipton Shale Member, one at the top of Zone 1 and the other in Zone 3 (Fig. 5), are useful in a basin analysis of the Green River Formation. These

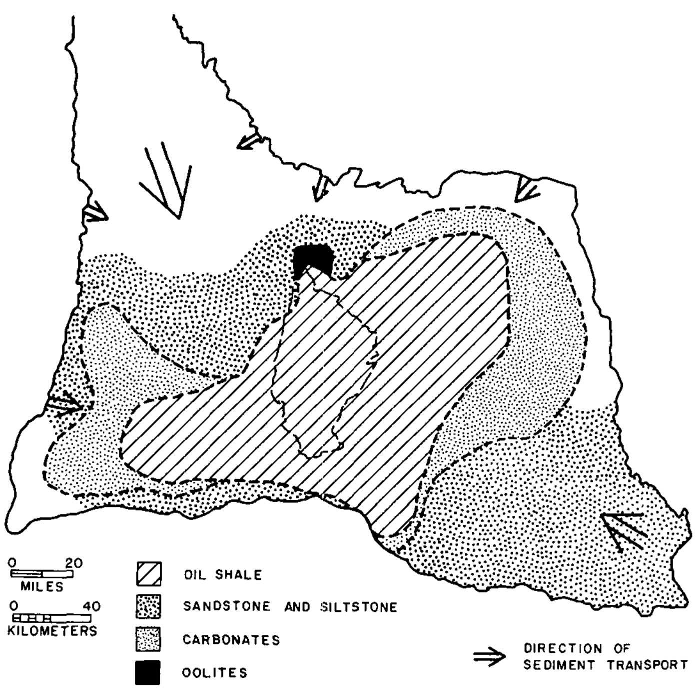

Figure 9. Distribution of lithologic facies during middle Tipton time; this period represents high stand of Lake Gosiute. Lacustrine facies represented by oil shale, mud-flat facies by carbonates, and marginal facies by sandstone and siltstone.

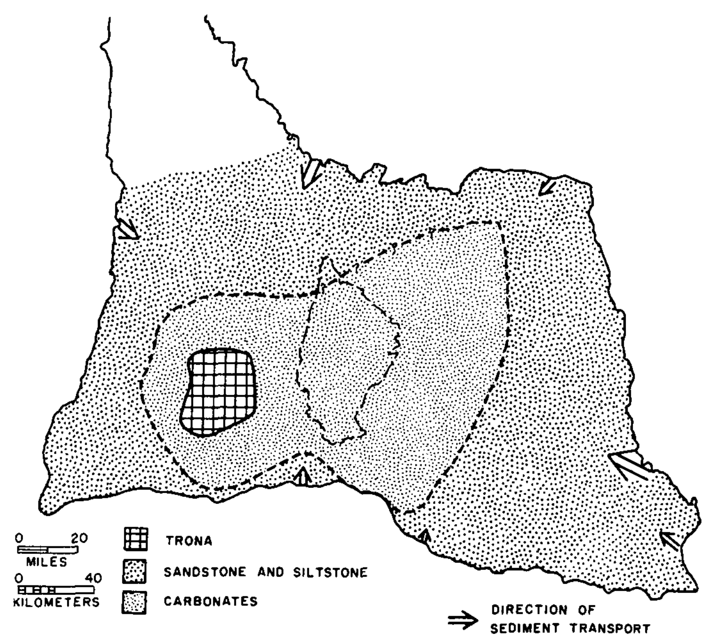

Figure 10. Distribution of lithologic facies during middle Wilkins Peak time; this period represents low stand of Lake Gosiute. Lacustrine facies represented by trona, mud-flat facies by carbonates, and marginal facies by sandstone and siltstone.

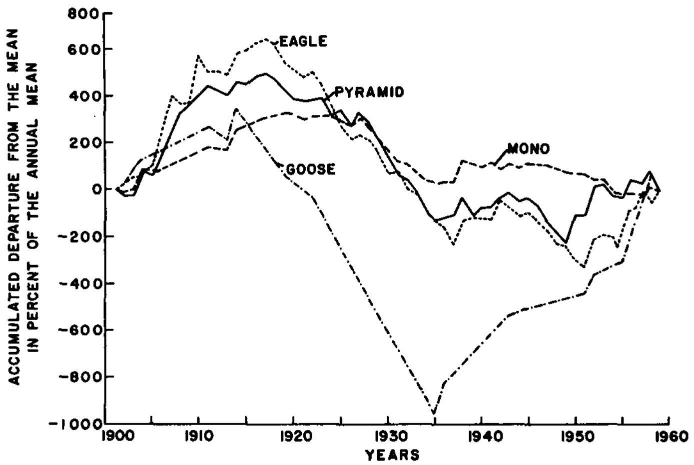

Figure 11. Short-term fluctuations in runoff into closed basins in western United States (after Harding, 1965).

stromatolites are exposed continuously along the White Mountain scarp from Boars Tusk (T. 23 N., R. 104 W.) south to Rock Springs, Wyoming, a distance of approximately 30 mi (Fig. 13). In the vicinity of Boars Tusk (section 5 of Fig. 13), the stromatolites of Zone 3 are as much as 1 m thick, and individual algal heads measure 30 to 40 cm in diameter; however, some heads are so closely packed that the upper surface of the unit has a hummocky appearance (Bradley, 1926, Fig. LXII; 1929). Southward along the White Mountain scarp, the algal heads in Zone 3 become smaller (15 to 30 cm in diameter) and less domal in character. Approximately 12 mi south of Boars Tusk, the stromatolites are more planar, consisting of stromatolitic algal structures 7 to 15 cm across (Fig. 13, sec. 3). Six miles farther south, the unit has lost most of its distinctly stromatolitic algal character. Slightly farther south the unit is composed of very fine grained dolostones characterized by mud cracks, flat-pebble conglomerates, salt crystal casts, and ripple marks (section 1 of Fig. 13).

Preserved tuff beds in this part of the Green River Formation provide a means of interpreting the depositional history within a chronologic framework. The relation between the stromatolite-dolostone lithologic unit and the tuff beds is illustrated in Figure 13. Assuming that the tuff beds are time-stratigraphic units, it is readily apparent that the stromatolite unit is time transgressive. The stromatolite unit is interpreted as representing a transition from a wave-swept shoreline along the northern margin of the basin during a high stand of the lake to a mud-flat environment in the Rock Springs area during a relatively low stand of the lake. During this transition, the shoreline migrated 30 mi to the south, and the stromatolite unit climbed about 2 stratigraphic ft/mi. Bradley (1964), using other geologic evidence, estimated that the topographic gradient in the basin during the deposition of the Green River Formation was 1 to 2 ft/mi.

A good estimate of minimum shoreline fluctuation can be determined by mapping the minimum and maximum extent of algal stromatolites in these diachronous lithologic units (Figs. 14 and 15). The regional distribution of the algal bed at the top of Zone 1 indicates that, during the early part of Tipton time, the lake shrank from 12,500 mi^2 (Bradley, 1963) to 4,500 mi^2 or less (Fig. 14). The regional distribution of algal stromatolites in Zone 3 of the Tipton shows that the lake also shrank thousands of square miles during the transition from Tipton to Wilkins Peak deposition. The areal extent of the lowermost trona bed (No. 1) in the Wilkins Peak Member (Fig. 15), which lies just above the Tipton–Wilkins Peak contact (Culbertson, 1971), marks a minimum stand of the lake and thus delineates the maximum shoreline fluctuation during the Tipton–Wilkins Peak transition. Thousands of square miles of mud flats were exposed during each of these large shoreline fluctuations. Obviously, the Tipton Shale Member does not represent a large, deep, stable lake, but instead was the product of an unstable lake characterized by large fluctuations in size and depth.

Seasonal changes and some small-scale fluctuations in water level can occur in large, open lakes that have a long response time. However, large-scale, long-term fluctuations are possibly only in a closed basin where increased inflow and (or) precipitation can be removed only through evaporation. Several mechanisms such as structural movements, stream piracy in the hydrographic basin, and the blockage of an outlet could account for the water-level fluctuation, but it seems highly unlikely that these mechanisms would have occurred repeatedly to cause the changes observed in the Tipton Shale Member. The regional distribution of the algal units in Zones 1 and 3 necessitates large decreases in the lake level. Such large-scale fluctuations can best be explained as a result of the effect of climatic changes on a closed basin.

Increasing Salinity and Alkalinity

Judging from the fossil remains, the waters of Lake Gosiute were fresh enough to support an abundance of plant and animal life during the lower part of Zone 1. With the exception of algae, however, there was a spectacular decrease in biologic activity by the end of Zone 1. Zones 2 and 3 are nearly devoid of fossil remains. It is not surprising that there is a dramatic correlation in the Tipton between the first appearance of cryptalgal structures and the disappearance of other fossils. Logan and others (1974), studying modern cryptalgal sediments in Hamelin Pool, Western Australia, have shown that hypersaline waters and consequent elimination of browsers are absolutely essential to the preservation of cryptalgal structures. For example, the cryptalgal structures (including algal heads) at Hamelin Pool are found only in hypersaline waters (53,000 to 65,000 ppm); in waters of consistently lower concentra-

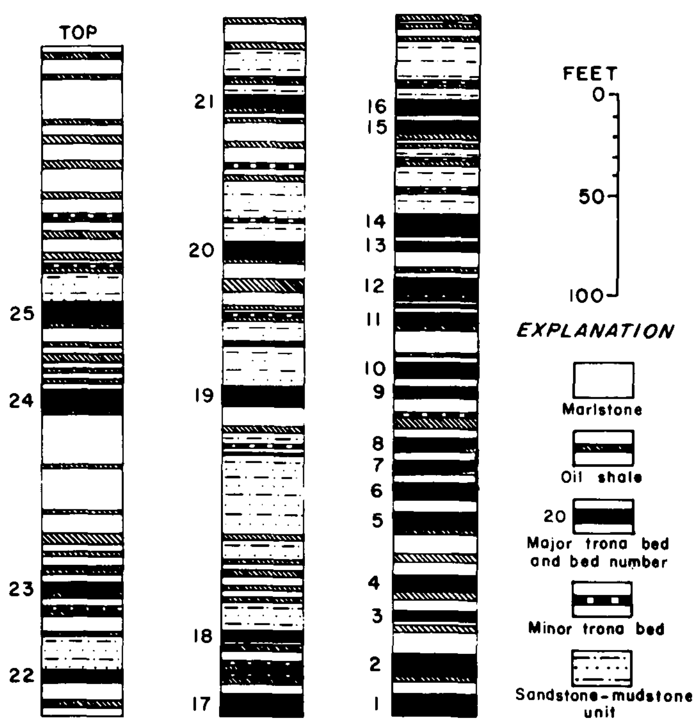

Figure 12. Composite columnar section of Wilkins Peak Member in trona area of Wyoming (after Culbertson, 1971).

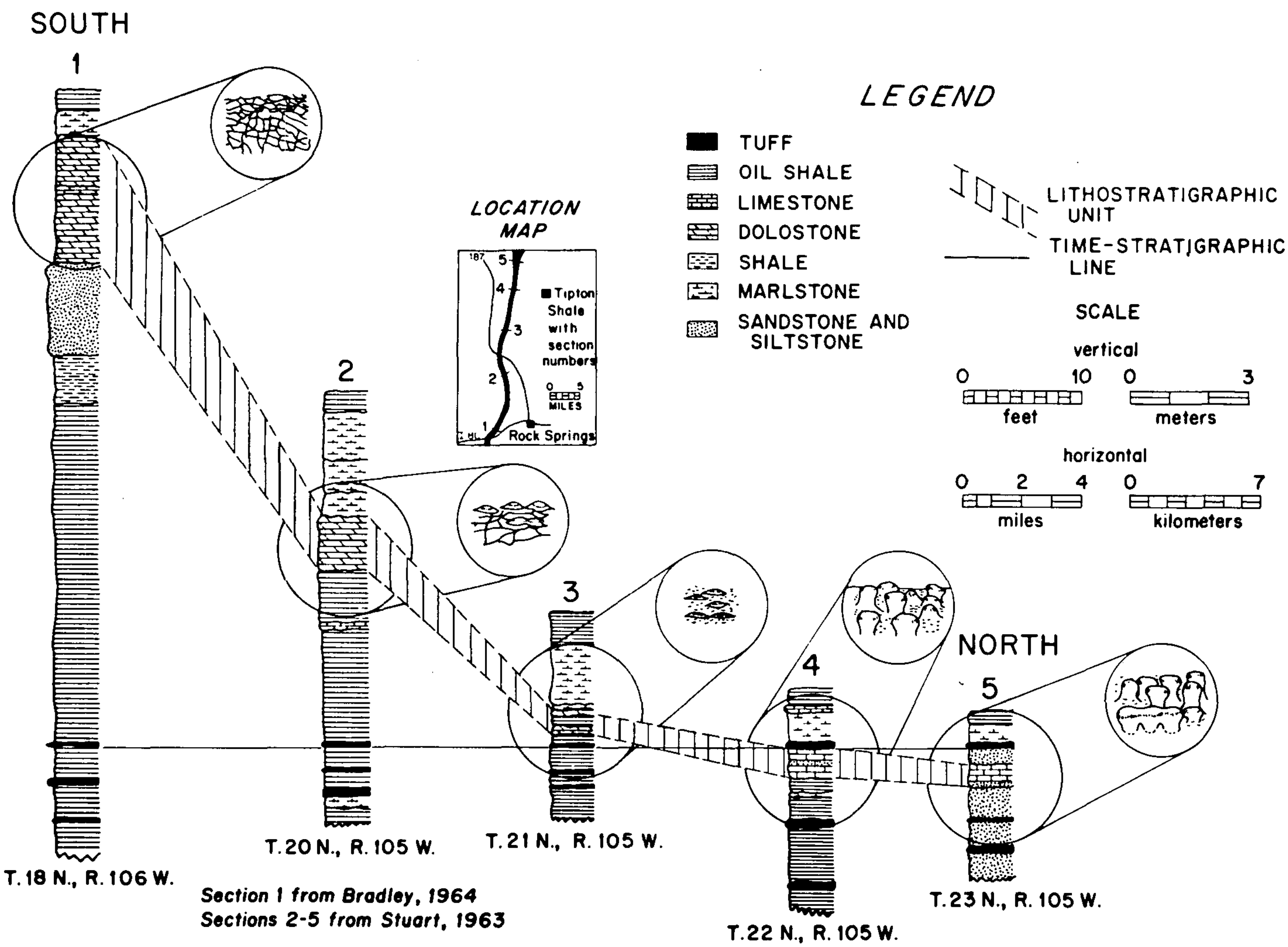

Figure 13. Stratigraphy of uppermost Tipton–lowermost Wilkins Peak section along the White Mountain scarp north of Rock Springs, Wyoming. Section 5 is so-called Boars Tusk section.

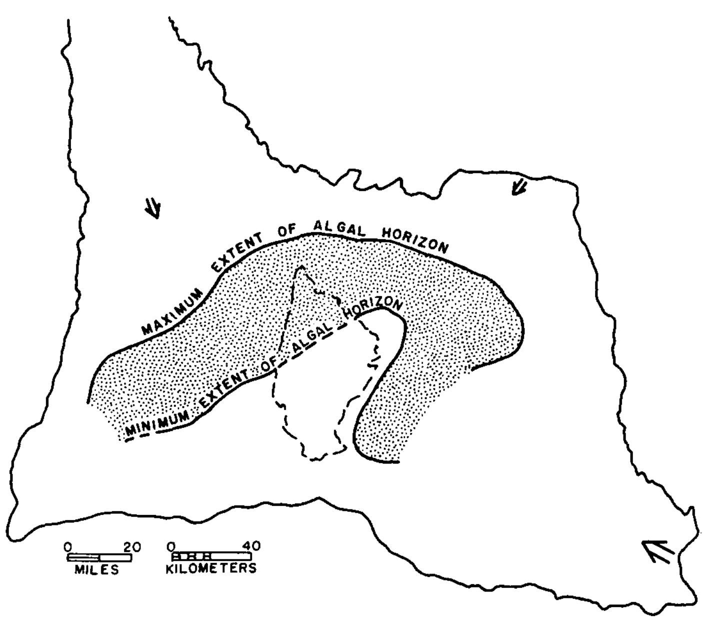

Figure 14. Minimum and maximum extent of stromatolitic algal unit at top of Zone 1 of Tipton Shale Member. Stippled area represents carbonate mud flats exposed as lake receded from high to low stand.

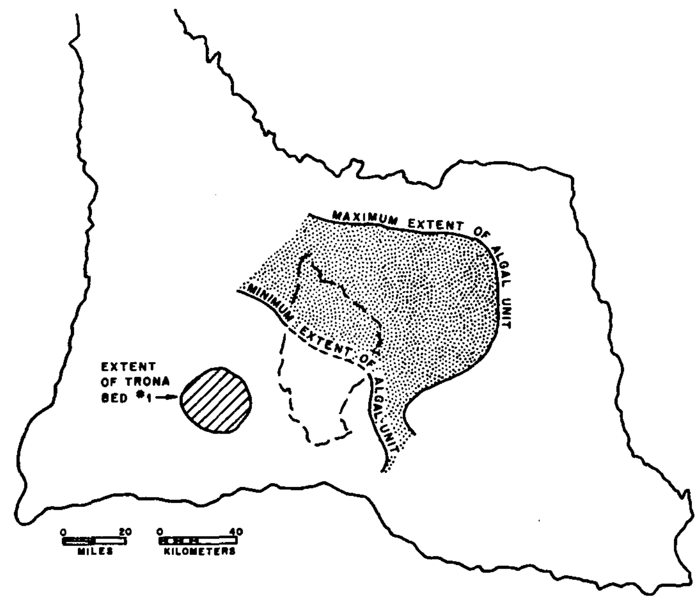

Figure 15. Minimum and maximum extent of stromatolitic algal unit in Zone 3 of Tipton Shale Member. Stippled area represents carbonate mud flats exposed as lake receded from high to low stand. Note areal extent of lowermost trona bed in Wilkins Peak Member of Green River Formation, which represents maximum shoreline recession.

tions (<50,000 ppm), the algal structures are destroyed by the activity of predators. It is suggested that the shrinkage of Lake Gosiute at the end of Zone 1 deposition (Fig. 14) greatly increased the salinity of the lake water, eliminating potential browsers and resulting in a proliferation of cryptalgal sediments. Judging from modern environments characterized by cryptalgal sediments, the waters of Lake Gosiute must have been hypersaline by the end of Zone 1. Subsequent major and minor fluctuations in the lake, particularly at the end of Tipton time, caused further increases in the salinity until, during earliest Wilkins Peak time, the lake brine became saturated with trona.

Evaporation is the major concentrating mechanism affecting waters in closed basins (Jones, 1966; Jones and others, 1967; Hardie and Eugster, 1970; Eugster, 1970). As a consequence, reasonable estimates can be made for the range of water compositions represented by Lake Gosiute. As Bradley and Eugster (1969) have suggested, it is reasonable to assume that the runoff during Eocene time probably had approximately the composition and concentration of the present-day Green River. This assumption is based on the fact that the Green River presently drains a large part of the same hydrographic basin once occupied by Lake Gosiute and in an essentially arid climate (Bradley and Eugster, 1969). Thus it is reasonable to assume that, during its initial stages, Lake Gosiute had a composition similar to the Green River (Table 1) and that this composition represents the very freshest stage of the lake.

The composition of the most saline stages of the lake can be estimated from modern brines saturated with respect to trona. Table 2 shows the range of compositions for modern Lake Magadi brines saturated with trona. It is inferred that the salinity of the Gosiute Lake brine at the trona-salting stage was approximately 280,000 ppm (Bradley and Eugster, 1969). It is significant to note that the brines at Lake Magadi have undergone a concentration factor of 5,000 to 8,000 (river water → trona-saturated brine). In order to go from Green River water to trona saturation, based on chloride or sodium, it would require a concentration factor of 4,000 to 7,000; that is, the complete range of compositions that characterized Lake Gosiute can be represented by the present-day Green River and this same water concentrated 4,000 to 7,000 times.

The most important aspect of the lake chemistry concerns the point at which the lake water became saturated with calcite or aragonite. In other words, assuming equilibrium with atmospheric P_{CO_2}, when did the pH of the lake water exceed 8.4 (Garrels and Christ, 1965)? At that point (pH 8.4), all calcium delivered to the lake via surface recharge would have precipitated immediately as calcium carbonate. Present-day Green River water has a pH of about 8.0 (U.S. Geological Survey, 1963), so that only a slight increase in pH would cause this water to become saturated with respect to $CaCO_3$ and to precipitate aragonite or calcite.

A good example of a modern lake where calcite precipitation can be observed is Pyramid Lake in western Nevada (Table 3). The calcium concentration of this lake has been relatively constant for 90 yr. Thus all the calcium delivered to the lake via the Truckee River during this period was precipitated, probably as calcite. Harding (1965) has shown that the mean annual runoff for the Truckee River Drainage in the period 1901 through 1958 was 600,000 acre ft. The lower Truckee River water characteristically contains 20 to 40 ppm of calcium (U.S. Geological Survey, 1961–1969). Thus, the Truckee River annually contributes to Pyramid Lake approximately 3×10^{10} cm^3 of $CaCO_3$.

This aspect of carbonate chemistry is dramatically demonstrated at the northern end of Pyramid Lake at the Pinnacles. At this locality, water from thermal springs that contains about 220 ppm of Ca with a pH of ~8.0 (Table 4) discharges into the lake (~10 ppm Ca and pH ~9.0). Where the spring water enters the lake, in the zone of mixing, there is an abundance of precipitated calcium carbonate, including ooliths, $CaCO_3$-cemented sand, and $CaCO_3$-coated mud balls. In the zone of mixing, the Ca content of the water is approximately 2 ppm (Table 4). Assuming lake water with a pH of 9.0 and in equilibrium with atmospheric P_{CO_2}, the equilibrium calcium concentration of Pyramid Lake water should be about 2 ppm. Thus in the zone of mixing there is chemical equilibrium with respect to $CaCO_3$ and a precipitation of $CaCO_3$ even though, in general, the lake waters are slightly supersaturated with respect to $CaCO_3$.

The oolitic limestones in the Green River Formation of Wyoming are rarely more than 15 cm thick. They are laterally discontinuous and are confined solely to strandline deposits. Thus it is suggested that most oolitic beds in the Green River Formation formed as a result of the interaction of relatively fresh (pH <8.4) surface waters with lake water (pH >8.4).

The oolith-forming process envisaged for the Green River Formation is directly analogous to the processes presently forming calcite ooliths at Pyramid Lake and vastly different from the processes forming aragonitic ooliths in the Great Salt Lake, Utah. The latter precipitate from wave-agitated water (Ca^{++} = 250 ppm) on shal-

TABLE 1. COMPOSITION OF GREEN RIVER

	Green River at Jensen, Utah (ppm)	Inferred composition from Bradley and Eugster (1969) (ppm)
SiO_2	10.5	15.8
Fe	0.1	0.1
Ca	41.1	61.4
Mg	13.8	20.7
Na	29.8	44.5
K	2.2	3.3
HCO_3	152.2	227.0
SO_4	80.6	120.5
Cl	12.0	18.0
F	0.3	0.4
NO_3	1.0	1.5
BO_3	0.2	0.2

TABLE 2. IDEAL COMPOSITIONS FOR BRINES OF MAGADI BASIN*

	First saturated brine (ppm)	Final saturated brine (ppm)
Cl	50,000	85,000
HCO_3 + CO_3	88,000	110,000
SO_4	1,300	2,300
F	1,300	1,500
Br	180	310
B	60	100
Na	107,000	130,000
K	1,650	2,000
SiO_2	750	1,280

* From Eugster, 1970.

TABLE 3. PYRAMID LAKE WATER

	November 1972 (ppm)	Jones (1966) (ppm)	Mean of 4 samples (Russell, 1882) (ppm)
Cl	2,100	1,980	1,319
HCO_3 + CO_3	1,411	1,153	n.d.
SO_4	263	274	182
Na	1,740	1,630	1,294
K	100	120	70
Ca	10	7	9
Mg	113	117	90
pH	9.2	8.9	n.d.

n.d. = not determined.

TABLE 4. PYRAMID LAKE SPRING WATERS

	Spring water (ppm)	Spring discharge 10 ft from lake (ppm)	Zone of mixing (ppm)	Lake water (ppm)
Cl	1,850	1,890	1,990	2,100
HCO_3 + CO_3	22	27	514	1,411
SO_4	272	282	256	263
Na	1,080	1,080	1,380	1,740
K	26	26	68	100
Ca	216	224	2	10
Mg	0.3	0.6	57	113
SiO_2	127	131	62	13*
pH	8.0	8.0	9.0	9.2

* From Jones, 1966.

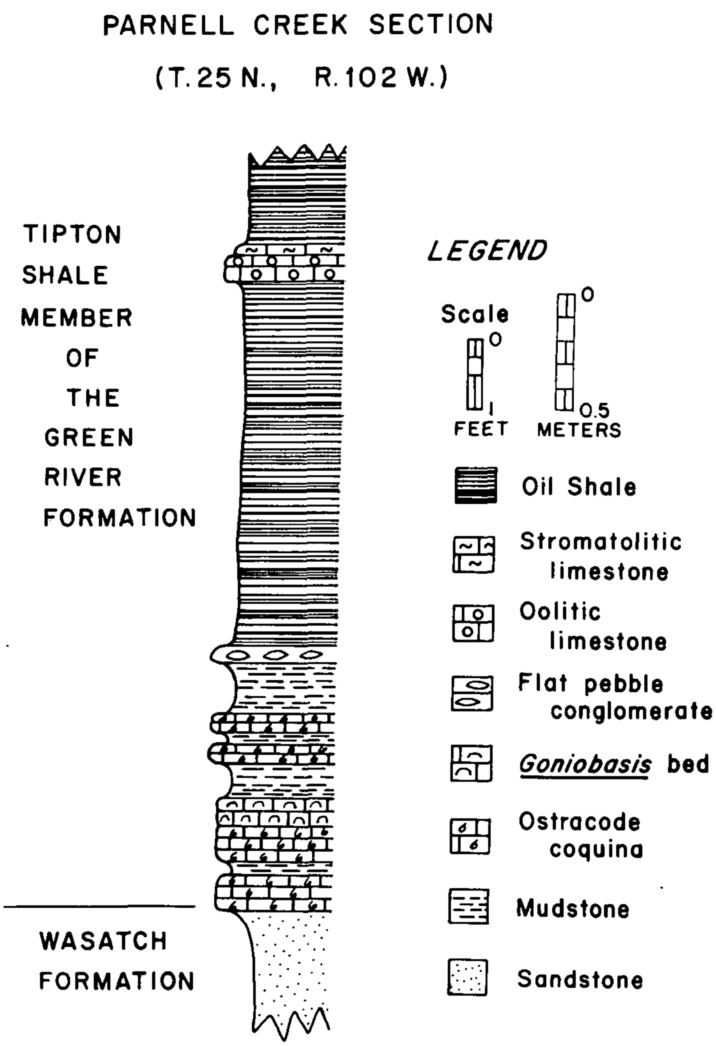

Figure 16. Stratigraphic section of basal Tipton at Parnell Creek. Note vertical scale. Oolitic limestone is approximately 10 ft above base of Tipton.

low shoals (0 to 3.7 m) and form continuous, relatively thick deposits covering hundreds of square miles.

Assuming that the analogy with Pyramid Lake is valid, it is possible to determine when Lake Gosiute became saturated with respect to $CaCO_3$ by evaluating the vertical distribution of oolitic limestones. In several stratigraphic sections, thin oolitic limestones occur in Zone 1 near the base of the Tipton (see Fig. 16). This suggests that, very early in its history, the lake was saturated with respect to $CaCO_3$. Thus, the vertical distribution of oolitic limestones, fossils, and cryptalgal structures strongly suggests that, with the exception of the *lowermost* Tipton, the lake waters were saline (>50,000 ppm) and alkaline (pH >8.4) during most of the history of Lake Gosiute.

Sedimentary Structures

As previously mentioned, sedimentary structures indicative of desiccation and subaerial exposure and shallow water should be common in a playa-lake environment. The rocks of the Tipton Shale Member contain the following assemblage of sedimentary structures:

1. The sandstones and siltstones are characterized by mud cracks, ripple marks with flattened crests, ripple marks with mud cracks in the troughs, many burrows and root casts, and thinly bedded units with current lineations.
2. The carbonate rocks are characterized by mud cracks, saline crystal casts, and plant debris; flat-pebble conglomerates; ooliths and pisoliths; and algal bioherms.
3. The oil shales are characterized by disrupted bedding and a lack of continuous lamination, some mud cracks and numerous incomplete desiccation polygons, and a variety of looped bedding and injection features. Continuous laminations are relatively rare.

Many of these sedimentary structures are not definitive in themselves, but the total assemblage of structures strongly supports the playa-lake model for the Green River Formation.

CONCLUSIONS

The lithologic and mineralogic distribution patterns for modern Deep Springs playa and the Eocene Green River Formation are strikingly similar. Moreover, in the Tipton Shale Member, which previously had been thought to represent a maximum stand of Lake Gosiute, there is strong evidence demonstrating large fluctuations in the position of the shoreline and progressive increases in the salinity and alkalinity of the lake water. The assemblage of sedimentary structures in both the Tipton Shale and Wilkins Peak Members is compatible only with a sedimentologic model containing as fundamental elements (1) shallow-water deposition, and (2) subaerial exposure and desiccation. On this basis, it is suggested that the deep-water stratified-lake model is untenable for both the Wilkins Peak and Tipton Shale Members of the Green River Formation. In contrast, the playa-lake model is consistent with the observations and chemical constraints inherent to the Green River Formation of Wyoming.

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