PALEOCENE–EOCENE LAKE FLAGSTAFF OF
CENTRAL UTAH

A Thesis
Presented in Partial Fulfillment of the Requirements
for the Degree Master of Science

by

Neil Andrew Wells, A.B.

The Ohio State University
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Approved by

[Signature]
Adviser
Department of Geology
and Mineralogy
ABSTRACT

Mid Paleocene - early Eocene Lake Flagstaff was the central part of a major lake system that extended SW-NE across Utah; the deposits of the lake compose most of the Flagstaff Formation. The Formation is divided into three Members. The lithologies and features of each reflect different phases of the lake which appear to be related to changes in climate, from humid to arid to subhumid, in relative terms. The lake during the early and late phases was fresh, calcareous, large, and highly productive. In contrast, the middle phase lake was restricted in area and fauna and was saline, on two occasions becoming a gypsum-precipitating brine.

The Lower Member of the Flagstaff Formation is mostly limestone. Dolomite and chert are absent from the lower beds but increase upward. The chert occurs as small crystallaria in voids, growing by nucleation from groundwater. Most of the dolomite is detrital, transported from the surrounding carbonate mudflats by sheetwash, although some is formed in situ. The limestones of the Middle Member are completely dolomitised, except for those previously replaced by nodular chert. The Upper Member is calcareous
like the Lower, but contains silicified beds and more dolomite. This silicification and the dolomitisation of the Middle and Upper Members were probably caused by the evaporative drawing of waters through the exposed mudflats. The model developed herein for the formation of lacustrine limestone suggests that the limestone in the Flagstaff probably was precipitated by pH fluctuation but also by CO₂ degassing by temperature change and photosynthesis.

Early Lake Flagstaff was well-vegetated, productive, and muddy, as shown by the abundant root and rootlet moulds, and burrows, and by the fauna. The fauna and abundant flora also show that the lake was shallow; indeed, it was subject to water level fluctuations (probably caused by wind tides and volume changes) that exposed or inundated vast areas of land. This is indicated by all the internal contraction features caused by dewatering (sheet and prism cracks, craze and skew planes, pull-aparts, etc.), the mechanically re-worked micrites (including sheetwash deposits), and evidence of pedogenesis and calichification, much of which (e.g. marmorisation, nodulisation, pseudoclast formation) is caused only by an oscillating water table. The lake floor and the floodplains were so flat because the underlying North Horn Formation floodplain in this area was very flat and because the combination of lacustrine regression, sheetwash, and transgression maintained the flatness by eroding topographically high areas and filling low ones.
Epiclastic input into the Wasatch Plateau depocenter was extremely limited. Most of the sediments shed from the Sevier orogenic belt, the western shoreline, were trapped in the depocenter in the area of the Juab Valley and the Gunnison Plateau, and those from the uplifts in the east were stranded on the flat floodplain. The floodplain facies comprises very red fine-grained epiclastics.

The lake was chemically simple. It contained much Ca$^{2+}$ and HCO$_3^-$, some Mg$^{2+}$ and SO$_4^{2-}$, and minor iron and silica only. From chemical considerations the early lake was rather alkaline (pH = 8 to 9), making it a hardwater alkaline lake. Its salinity was probably not over 500 ppm. The middle lake phase was saline, containing at least 2000 to 3000 ppm dissolved solids (mostly SO$_4^{2-}$), and its pH fluctuated between 8 and 10.
ACKNOWLEDGEMENTS

This thesis and I owe our continued existence to Leslie Cox, who helped tremendously at all the rough spots. Special thanks are due to Dr. Kenneth O. Stanley, my adviser, for suggesting this topic, for his help throughout, and for his patience, and to Lon McCullough, without whom the field work and later analyses would have been highly difficult. I am indebted to Dr. James Collinson for his valuable ongoing help and suggestions. I would like to acknowledge Drs. Garry McKenzie and C. W. Summerson for reviewing this thesis and respectively serving on my committee and making laboratory equipment available, and also Robert S. Dawson for his toil in the field. I would also like to thank Dr. Gunter Faure, Robert Wilkinson, Jeffrey Franklin, Donald Cooke, and Charles Vavra, for laboratory and technical assistance, Dr. David Elliot for forbearance, and the many others for their gracious assistance.

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INTRODUCTION

Ancient Lake Flagstaff existed in central Utah from mid Paleocene into the Eocene. The major lake system of which it was a part developed in a NNE-SSW oriented, Cretaceous to Eocene sedimentary basin (Figure 1) in which the Flagstaff, North Horn, Colton, and Green River Formations were deposited (Spieker, 1949). The Flagstaff Formation includes lacustrine carbonates and some supralacustrine epiclastic facies and is now exposed in the plateaus and mountain ranges around the San Pete and Sevier Valleys (Figures 2 and 3). The regional geology has been much studied, particularly by Ohio State University geologists led by Dr. E. M. Spieker. The paleontology of the Flagstaff has been studied by Gill (1950), La Rocque (1951, 1956, 1960), and Weiss (1969).

Because of the limited scope of the previous work on the paleoenvironment of Lake Flagstaff and the extensive recent literature on paleolimnology, renewed study of the Flagstaff Formation became imperative. With that in mind this study and its complement by Lon McCullough (1977) were begun during the summer of 1976. We inspected as much
Figure 1  
Index map of Utah, showing location of figures 1 & 2, some structural features, and the possible extent of lacustrine deposition (after Hintze, 1973; Schneider, 1969)
of the Flagstaff as possible in order to understand its variations, and, with Bob Dawson, we measured sections on the Wasatch Plateau (Figures 4 and 5). Our efforts included extensive sampling, detailed section measuring, laboratory analysis, and some theoretical modelling. This approach proved effective, enabling us to recognise the playa nature of the lake and the increased salinity of the middle phase, improve on the stratigraphy (increasing the resolution, improving correlations, and explaining the facies), and begin to answer questions about the paleolimnology of the lake.

The purpose of this thesis is to identify the sedimentary features present and to discuss their implications and the constraints that they impose on the interpretation of the lake. It emphasises the paleolimnology of early Lake Flagstaff, the petrology of the limestone, the origins of the dolomite, gypsum, calcite, and chert, and the chemistry of the lake water. McCullough's (1977) thesis emphasises the origin of the dolomite, the chemistry of the iron and the clay mineralogy of the Flagstaff, the nature of the Middle Member (the playa-lake phase), clastic input, and the paleoclimatology.
Fig 5. Location of Sections (See App. 2)
STRATIGRAPHY

Stratigraphic Setting

Under the Flagstaff Formation lies the North Horn Formation, and beneath that, where present, the Price River Formation. Above are the Colton and (or) the Green River Formations. This sequence represents more or less uninterrupted sedimentation from the late Cretaceous through the Eocene, which was influenced mostly by Sevier and Laramide orogenesis.

The conglomeratic Price River Formation is a flood of debris shed from the previously folded Sevier orogenic belt (see Figure 1) (Spieker, 1949; Harris, 1959; Armstrong, 1968). Sediment was dispersed to the east across flat coastal plains to the sea that was then over Colorado (Hunt, 1956). The unconformity below the Price River, which is marked in the area of this report, dies out eastward (Spieker, 1949).

The early Tertiary in Utah was marked by numerous uplifts in the eastern half of the state, including the San Rafael Swell, the San Juan and Uinta Mountains, and the Circle Cliffs, Monument, and Uncompahgre upwarps (Figure 4)
(Hunt, 1956). These upwarps were sufficient to reverse the drainage and cause ponding. As this occurred fanglomerate and sandstone of the Price River gave way upward and westward to a mixture of lithologies representing the floodplains, rivers, small lakes and ponds, and alluvial fans of the North Horn Formation (Spieker, 1946; Weiss, 1969). Therefore, most of the Price River intertongues and is gradational with the North Horn, but local folding along the western edge of the sedimentary basin led to minor unconformities (Gilliland, 1948). Despite continuing crustal activity (Weiss, 1969; Birsa, 1974), the increased sedimentation gradually buried the paleotopography.

With the onset of internal drainage, Lake Flagstaff was formed. The Flagstaff Formation, in its broadest sense, includes the sediment body of the main lake centred over the Wasatch Plateau, those of minor, mostly synchronous lakes, and most of the supralittoral sediments that are intimately associated with the main lake. Its base is the first lacustrine limestone above the redbeds and sandstone of the North Horn (Spieker, 1946, 1949). Some of the North Horn is synchronous with the lower part of the Flagstaff and represents the supralittoral facies (McGookey, 1960; Birsa, 1974). The Colton has a similar relationship with the Upper Flagstaff and also with part of the Green River Formation (Marcantel and Weiss, 1968). The Colton Formation represents the floodplain and pond environment that existed
adjacent to the Flagstaff lakes and later replaced them. The Flagstaff/Colton contact, as defined by Marcantel and Weiss, is marked in part by the intertonguing of oxidised, epiclastic sediments with the Flagstaff calcilutites (cf. Bonar, 1948). In mid Eocene the Green River lake in the Uinta basin expanded, incorporated the northern remnants of Lake Flagstaff, and inundated the old Flagstaff basin.

Stratigraphic Nomenclature in Central Utah

Hayden introduced the term 'Wasatch Group' in 1869 to describe the red and variegated clastics common in the Early Tertiary of the west, as typified at Fort Bridger, Wyoming. The group was gradually extended south into northwestern Colorado and then west along the Book Cliffs, whence Spieker and Reeside (1925) applied the name Wasatch Formation to the Early Tertiary of the Wasatch Plateau. They subdivided it into the lower, Flagstaff, and upper Members, defining the Flagstaff Member as the main lacustrine facies. When part of the lower Member was found to be Cretaceous in age, and the Flagstaff to be mostly Paleocene, Spieker (1946) changed the members into separate formations (North Horn Formation, Flagstaff Limestone, and Colton Formation respectively), to avoid confusion, because the term Wasatch by then had a distinctly Eocene connotation. The definitions remained the same, but as more people worked in the area the Flagstaff was gradually expanded to encompass both readily
recognisable parts of the lake's shoreline and other geographically separate yet synchronous lake deposits.

In 1948, Gilliland proposed the "Flagstaff Formation" to include the marginal facies in the Valley Mountains. He published the name in 1951. Since then both names have been used. In this paper the "Flagstaff Formation" is used to obtain the broadest definition of the Flagstaff. In 1967 Schneider correlated the Flagstaff and Cedar Breaks Formations, a correlation that Spieker had suspected in 1946, and assigned to the latter some of the southern clastic-rich Flagstaff.

Using recent drill core data from the Uinta basin and referring strictly to the beds of the main lake only, Pouch (1976) proposes renaming the Flagstaff in the Uinta basin as the Flagstaff Member of the Green River Formation. He also recommends reclassifying the North Horn and Colton Formations as the Wasatch Formation wherever they are not separated by the Flagstaff. The latter is a very useful simplifying concept. Comparable demotion of the Flagstaff in the Wasatch Plateau depocenter causes a loss of stratigraphic resolution. As Bachman (1959) writes in discussing this same idea, "the thickness, extent, and distinctive lithology of the Flagstaff merit its formational rank."
Subdivision of the Flagstaff Formation

The Flagstaff is readily divided into the Lower, Middle, and Upper Members, and the Middle Member is in turn divided into the lower, intermediate, and upper units. The Lower Member is fossiliferous. It contains beds of grey mudrock, limestone, and limestone and dolomite, and is larger chert-free, although near the top of the member the rocks are dolomitic and some chert occurs as spheres and fillings in voids and sparse nodules. Except for its intermediate unit the Middle Member is characterised by red mudrock, the absence of gastropods and, in sharp contrast to the Lower Member, white carbonate units that are all 100% dolomite/total carbonate and contain large nodules of chert. The upper and lower units of the member contain some gypsum and (or) red mudrock and are separated by an intermediate unit made up of dolomite and sparse thin layers of 90-99% dolomitised gastropodal limestone (see Figures 6 and 7). The base of the Upper Member is defined by the return of gastropod fossils and of less-than-completely-dolomitised carbonates, although it also contains much pure dolomite. Chert is abundant as silicified limestone beds.

McCullough (1977) and I have defined members for the Flagstaff Formation at Cove Mountain (see Figures 5 and 6, and McCullough, 1977, pp. 79-101), which are essentially those units set up by Gill (1950). (See Table I.)
Fig. 6 Cove Mountain section

% of beds with mollusks

Upper Member

Middle

Lower Member

dolomite
dolomite+calcite

upper unit

middle unit

lower unit
**TABLE I**

**Correlation of Subdivisions of the Flagstaff**

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Compare to Gill (1950) and La Rocque (1960). No vertical scale implied.
While snail populations may shift due to facies changes, their widespread and abrupt disappearance above the Lower Member is probably related to changes in the nature of the lake, especially changes of a chemical nature (La Rocque, 1960). Such changes are probably linked to climate (McCullough, 1977) and were therefore probably synchronous throughout the lake. The gypsum precipitating phases probably also represent time zones. Therefore the Middle Member and its subdivisions are probably close to being time-stratigraphic units. The onset and demise of lacustrine conditions are probably roughly correlable through this region, especially if they occurred in response to one or two major tectonic and (or) climatic events. This would be particularly true in small areas, such as the centre of the Wasatch Plateau, however the Flagstaff Formation overall is by no means a chronostratigraphic unit.
Fig. 8 "Becver Pond" Section
Ferron Mt.
Fig. 9  Correlation of units along Ferron Mountain
PALEOGEOGRAPHY

Paleocene-Eocene lacustrine deposition in Utah occurred in several areas therefore it is necessary and convenient to discuss here the tectonic setting and the regional correlation of these lake deposits. During this time a chain of basins, or perhaps one large basin, extended diagonally across Utah, from the southwest corner of the state north-east into Wyoming (Hunt, 1956; Hintze, 1973). This chain can be divided into three major parts; the Uinta basin in the north, an unnamed basin in the south (see Schneider, 1967), and inbetween, through central Utah, are the several minor basins or depocenters of the Flagstaff Formation (see Figs. 1, 2, and 3). The Flagstaff Formation was bounded on the west side by the Sevier orogenic belt and in the east in part by the San Rafael Swell (Hunt, 1956). It includes the deposits of the lake in the Wasatch Plateau depocenter (Lake Flagstaff) and those of several lesser lakes (cf. Weiss, 1969).

The southwest corner of the Uinta basin is tangent to the northern edge of the Wasatch Plateau depocenter. The Uinta basin was bounded by the early Uinta Mountains along
its northern edge and the Douglas Creek arch to the east; the middle of the southeastern side of the basin lay between the San Rafael Swell and the Uncompahgre uplift, whereas the southern Wasatch Mountains formed the western shore (Osmond, 1964). It contains some lake deposits that are synchronous and almost identical to those of Lake Flagstaff (Ryder et al., 1976). South-southeast of the Wasatch Plateau depocenter are the epiclastics and lacustrine carbonates of the Cedar Breaks Formation (Schneider, 1967). This Schneider correlated with the Flagstaff Formation as defined by Lautenschlager (1952) in the Pavants, where it includes the extralacustrine red clastics that might elsewhere be called the Colton Formation. The Cedar Breaks basin or depocenter extends south to the Kaibab uplift (on the Arizona border), east to the Circle Cliffs and Monument uplifts, and southwest to the corner of the state (Hintze, 1973, Fig. 52; Schneider, 1967, Fig. 6). The western limit of both basins were the mountains of the Sevier orogenic belt (the Sevier arch of Harris, 1959) (Armstrong, 1968; see Figure 1.)

The Flagstaff Formation lies between these two major basins (Figures 1, 2, and 3) and is bounded on the west by the main part of the Sevier orogenic belt, and to the east by the San Rafael Swell (Hunt, 1956). It includes the deposits of the lake in the Wasatch Plateau depocenter (Lake Flagstaff) and those of several lesser lakes (cf. Weiss, 1969).
Use of our field data (Figure 5, Appendix 1) and information from previous research (Figure 10) allows the construction of an isopach map for the lacustrine facies of the Lower Member of the Flagstaff (Figure 11). This map indicates the amount of lacustrine sediments through the area and therefore reflects primarily the local persistence of lacustrine conditions due to tectonism and locally variable rates of sedimentation, rather than the actual bathymetry of the lake floor or the shape of the entire basin. This map is speculative. Beyond the Wasatch Plateau, it rests heavily on the measurements of others, and on my interpretation of their sections, although we have visited and collected most of the areas (Appendix 1). It is however presented as an alternative to the maps of Gunderson and Gilliland (1967) and of Weiss (1969), which it only grossly approximates. The differences with the former map are because Gunderson and Gilliland have mapped the total thickness of the Formation left after erosion and included the supralittoral facies, whereas I considered only the lacustrine facies of the Lower Member and have included no incomplete sections (except on the west side of the Gunnison, where the section is all but complete - Hardy, 1948; Zeller, 1953; Hardy and Zeller, 1953). Differences with Weiss concerning the extent and timing of the lakes are due to new correlations of sections according to gastropod criteria defined by La Rocque (1960) and the definition of the Middle
Figure 10. Previous work in the area

A) Davis, 1967                      W) Katherman, 1948
B) Metter, 1955                     X) Birsa, 1974
C) Khin, 1956                       Y) Taylor, 1948
D) Mase, 1957                       Z) Hunt, 1950
E) Schoff, 1937                    A') Muessig, 1951
F) Pashley, 1956
G) Kucera, 1954
H) Gill, 1950
I) Bonar, 1948
J) Fagadau, 1949
K) Johnson, 1949
L) Bachman, 1959
M) Baughman, 1959
N) McGookey, 1958, 1960
O) Lautenschlager, 1952
P) Maxey, 1946
Q) Tucker, 1954
R) Gilliland, 1948, 1951
S) Babisak, 1949
T) Hardy, 1948
U) Vogel, 1957
V) Zeller, 1949
Member presented in the last section (contains no snails, many redbeds, and some gypsum) (see Table I). The assumptions and evidence for these recorrelations and interpretations are discussed in detail in Appendix 2, but the recorrelations are summarised below.

In brief, Gilliland's (1948) unit B is recorrelated with the lower Flagstaff, and C with the middle Flagstaff on fossil evidence mentioned by Lautenschlager (1952, p. 57) and Gilliland (1968) in accordance with criteria set forward by La Rocque (1960). This has ramifications in other work in the Juab Valley and surrounding areas, and restricts the size of the Middle Flagstaff lake. Locally and with varying degrees of certainty, the "upper North Horn" is shown to be equivalent in age to the Lower Flagstaff. This situation occurred primarily where workers wished to place a lacustrine section at the top of the North Horn Formation in accordance with Spieker and Reeside's type section (1925), and defined the overlying rocks as Middle Flagstaff (cf. Katherman, 1949; McGookey, 1958, 1960). While not incorrect, it has led others to postulate large "North Horn" lakes that are too large and "Lower Flagstaff" lakes that are too small (cf. Weiss, 1969). My distinction between the extralacustrine and lacustrine facies of the Flagstaff also implies that the lakes were smaller than the outcrop area of the Flagstaff Formation. The recognition of nonlacustrine rocks and several other lines of evidence including the
recognition of strongly evaporitic conditions during a more arid climate (McCullough, 1977) show that the middle Lake Flagstaff was much reduced in area and not enlarged relative to the early Lake Flagstaff as proposed by La Rocque (1960) and others. This in turn suggests the recorrelation of some otherwise undated or undatable lacustrine subunits which had previously been assumed middle Flagstaff in age merely because they were thought to demonstrate the expansion of the middle Flagstaff lake. The diagram also shows that the early Flagstaff rocks at the Northern Wasatch Plateau are more certainly a part of Lake Uinta than of Lake Flagstaff, as has previously been assumed (see La Rocque, 1960; Davis, 1967), thus supporting Ryder et al. (1976). Please refer to Appendix 2 for further details.

The isopach map shows that two main depocenters existed in central Utah: (1) the Wasatch Plateau depocenter (Lake Flagstaff) and (2) another west and south of the Gunnison Plateau including Long Ridge, the Juab Valley, the Pavants, the Valley Mountains, and the western Gunnison Plateau (Fig. 11). A complicated arrangement of highs and lows existed in the Gunnison Plateau, forming at least one separate basin along the eastern edge of the Plateau (Birsan, 1974). Possibly another separate basin existed briefly in the Cedar Hills area (Gunderson and Gilliland, 1967; Schoff, 1937).
This map is based on previous suggestions concerning two ridges, one in the central Gunnison (the shoreline of the Juab Valley Lake) and one running the length of the San Pete–Sevier Valley. Evidence for the central Gunnison high consists mainly of the pinching out of the North Horn and Flagstaff on the Indianola at locations such as Reddicks Canyon at the head of Wales Canyon (Hunt, 1950; Birsa, 1974). Evidence for the ridge in what is now the San Pete–Sevier Valley includes the pinching out of the North Horn and Flagstaff over locally tilted Arapien, Morrison, Indianola, and North Horn strata east and south of Salina (Gilliland, 1948) and north of Gunnison (Hunt, 1948; Babisak, 1949); also shoreline deposits on both sides of the valley, sediments shed eastward into the Wasatch Plateau depocenter (Bonar, 1948; Gilliland, 1948; Fagadau, 1949; Johnson, 1949; Gill, 1950; Pashley, 1956) and huge amounts shed into the Gunnison Plateau depocenter (Birsa, 1974).

Gilliland named this ridge the San Pete–Sevier Valley anticline, postulated 4500 to 6000 m of uplift on the structure, and suggested that the earliest movement was late Cretaceous. The extent of crustal shortening is uncertain. He considers it the result of horizontal compression; however it could also have been produced by thrusting or high angle reverse faulting. Neither are expressed at the surface, rather stress on the fault planes might have caused rupture and flowage in the Arapien Shale and diapiric
movement of the salt which in turn could have produced the antiform at the surface—see Bell (1953) and Stokes (1952). Regardless, sporadic local uplift continued from late Cretaceous through late Paleocene, thus local unconformities occur between all formations and occasionally within them (Birsa, 1974). Birsa believed that most of these unconformities were caused by localised salt tectonics and should not be interpreted as regional orogenies as has been suggested by Spieker (1949) and Gilliland (1948, 1951), among others.

The ridge shed much sediment during deposition of the North Horn and Lower Flagstaff but thereafter neither it nor the central Gunnison ridge were active (Birsa, 1974). Undoubtedly they had a greater effect on the Flagstaff lakes as barriers to eastward sediment transport, because while the Flagstaff in the Wasatch Plateau depocenter is entirely dominated by carbonate rocks, that west of the San Pete-Sevier Valley in the Juab Valley and eastern Gunnison depocenters is mostly terrigenous (Tucker, 1954; Vogel, 1954; Muessig, 1957). Most of the epiclastic sediment in the lake system was eroded from the old Sevier orogenic belt and was trapped in the Juab Valley depocenter and never reached Lake Flagstaff (Gilliland, 1948). Indeed, much of it was stored in the thick fans and plains of the western half of the Juab Valley depocenter (not shown on the map). Gradually the San Pete-Sevier Valley became buried and
breached in many places and by late Flagstaff it had been reduced to a chain of islands (Weiss, 1969) and the Cedar Hills (Davis, 1967) and eastern Gunnison (Birsa, 1974) depocenters were joined with the main body of the Flagstaff.

Whether the lakes of Uinta, Flagstaff, and Cedar Breaks basins were ever joined is not known, but one can speculate on it. Despite the lack of intervening uplifts, lakes of the Flagstaff and Cedar Breaks basins were probably not joined, except possibly by occasional streams; the northernmost Cedar Breaks rocks are 80 km from the nearest similar rocks (in the Pavants) and both are dominantly nonlacustrine (Schneider, 1967). Lacustrine rocks are present south of Salina, only 55 km away, but they are dissimilar (McGookey, 1960). In between is the thick Marysvale volcanic field, although poor and isolated exposures of the Flagstaff occur 12.5, 30, and 40 to 50 km away (Weiss, 1965). These now require reinvestigation.

A major conjunction between lake Flagstaff and Lake Uinta prior to the merging at Thistle at the end of the late Flagstaff (La Rocque, 1960; Davis, 1967) is far more likely, but not during the middle Flagstaff because of the progradation of the Colton floodplain (Peterson, 1976). However, the Colton barely expanded to Soldier Summit during its maximum progradation and earlier lay well to the south east, (Henderson, 1964) and because Lower Flagstaff lacustrine beds are known in the northeast quarter of the Wasatch
Plateau (Kucera, 1954), and throughout the Soldier Summit region (Ryder et al., 1976), a connection during the early Flagstaff is probable anywhere along the northeast edge of the Wasatch Plateau. The extent and geography of the junction would have been greatly affected by the amount, extent, and rate of uplifting. The greater the net uplift (and the rate of sediment supply to the floodplain) and the more extensive the area affected, the farther the shorelines would be from the San Rafael Swell, and the more the connection would have been constricted and "bent" around its northwest corner. Either progradation or extensive uplift could separate the two basins. Therefore, whatever connection existed prior to the conjunction at Thistle probably lay between Thistle and Price, but because the critical rocks have been eroded we may never know.
LITHOLOGY OF THE FLAGSTAFF

Introduction

The rocks of each member of the Flagstaff Formation are distinctive and can be generalised into broad categories, which are here typed by colour. Rocks of the Lower Member consist primarily of thin-bedded limestones and dolomitic limestones interbedded with mudrock (sensu Pettijohn, 1957). Limestone beds are commonly 5 to 25 cm thick but can range up to 1 m. They are in the intermediate 5Y hues (generally 5Y 5/2, light olive grey, although the lightness and saturation can vary +1 (Rock Color Chart Committee, 1975). Some are distinctly more orange (in the lighter 10 YR values). The majority of the organic-rich limestones have a very low chroma and are dark (N 4 or darker, in some cases 5 YR 2/1). The bioclastic material is commonly a medium grey, sometimes bluish (N 5½, 5B 5/1). Mudrock is present in thin beds 5 to 15 cm thick and as thin shale partings (commonly less than or equal to 1 cm), although rare beds are well over a metre thick. Where mudrock constitutes less than 10 to 25% of the section, the Flagstaff crops out as cliffs, otherwise the outcrop is a slope with "steps" of more resistant
limestone beds. Mudrock varies in colour: some are shades of red, red-brown, and orange; some are variegated hematite-red, limonite-yellow and brown; whereas a few are dull dark grey green. Most mudrock however occurs in shades of grey.

The rocks of the Middle Member are mostly dolostone and mudrock. These are typically much lighter in colour, commonly in the 5Y hues (8/1, 6/1) or the light neutral greys (N 5 to 0). This member is particularly rich in red mudrock. The dolostones are commonly thin (but range between 1 cm to 2 m); mudrock sequences are generally thicker. These rocks also contain bedded gypsum and abundant chert nodules. This member composes the prominent white caprock of the plateaus.

Rocks of the Upper Member are dolostone, limestone, and mudrock and perhaps exhibit the most variability in colour and thickness. Mudrock sequences are as thick as 3 m and are mostly very light-coloured (white, very light grey, yellowish grey, etc.). The carbonates are also commonly very light and generally are thin-beded, in some cases paper thin, commonly 5 to 15 cm, rarely up to 1 m thick. Beds of chert are common and are also 5 to 15 cm thick, but they display a wider range of colours - blacks, whites, yellows, and browns with minor pale reds and olives.

Channelform sandstones are rare, yet present in each member.
The Mudrock

Four types of mudrock occur in the Flagstaff: blocky grey mudrock, shale partings, variegated beds, and redbeds. The mudrock is almost all unfossiliferous and is equally free of the intraclasts that are so common in the limestone; only four intraclastic mudrocks were found during the entire field season. The blocky grey mudrock is the most common type. It occurs in beds from 5 cm to over 1 m thick and is rarely well exposed. It weathers into (or perhaps occurs naturally as) hard, angular, blocky granules. Its dolomite/carbonate ratio is as variable and as unpredictable as that of the surrounding micrites. The shale partings are by definition thinner than 5 cm, usually no more than 1 cm. They are friable and thin but extensive grey to black shales. In contrast to the other mudrocks these are fissile (or thinly laminated), not blocky. The variegated beds are similar to the grey blocky mudrocks; however they have varying amounts of red, purple, and yellow colour mixed in with the grey. All three types of mudrock may grade into the surrounding limestone, becoming more indurated and losing their blocky structure.

Red mudrocks are unique because apart from being red (from hematite), they are also the only type that is always composed of clay minerals; the others are mostly low-clay calcilutites. Insoluble residue analysis shows that only
8 of 35 non-red mudrocks were clay mineral mudrocks (contained more than 50% clay) whereas all five redbeds were claystones. The two variegated beds were calcilutites. In comparison, the 135 "limestones" tested by McCullough were all very pure carbonates. Because of their high clay content, their dearth of fossils, their high hematite content, and their lateral gradation into thick, coarser extralacustrine sediments (Johnson, 1949; Bachman, 1959; Baughman, 1959), the lack of epiclastics in the other lake sediments, and the likelihood of storage of clastics on the floodplains because of the gentle slope, the red mudrock is interpreted as the clastic, floodplain facies and their presence in an otherwise lacustrine section is assumed to represent a significant regression of the lake or progradation of the shoreline.

Because "mudrocks" are not clay-rich, they pose a problem in classification. Technically the various types of shale are made up of clay-sized particles (A.G.I., 1960), but commonly they are defined as being composed of clay minerals (see discussion in Pettijohn, 1957, p. 341). Marlstone (sensu Pettijohn) applies to these rocks; he describes them in part as "...25-75% clay....The marlstones are less fissile than the shales, generally have a grey to bluish-grey colour, and are marked by a blocky subconchoidal fracture." However this name is inappropriate because marl has many other connotations (e.g. a pure, friable lacustrine
limestone - Davis, 1901) and, as pointed out by Picard
(1953), "marlstone has been defined, redefined, and subse-
quently misused enough to render it useless." Picard sug-
gests that mudrock with over 25% clay minerals and 50%
clay-size particles be mudstones and shales and the re-
mainder be limestones and dolomites (the latter half is
followed herein). The mudrock is micrite (Folk, 1974) or
mudstone (Dunham, 1962) but these names do not sufficiently
distinguish the lithology. Thus the most appropriate term
is Grabau's (1904) calcilutite, which also has the advantage
of not making a calcite/dolomite distinction (A.G.I., 1957).
The clay size of the carbonates is supported by petrographic
and SEM studies.

The characteristic blockiness of the calcilutite may be
the result of flocculation. White (1961) noted that calcium
will readily flocculate clays, thus in calcareous shales the
clay minerals are unlikely to have settled with parallel
orientation and the shale is unlikely to be fissile. The
more flocculation, the blockier is the final shale. Bour-
cart (1946) shows that calcite can be flocculated by organic
acids, but whether calcite with 5 to 30% clay minerals can
be flocculated sufficiently is uncertain. He does note that
blocky calcilutite is common in lake beds.
Micrites of the Lower Member

The discussion of other carbonates is best largely restricted to those of the Lower Member, because the rocks are the least altered and the best exposed. They also form the bulk of my study and are unlike the rocks of the Middle Member (see McCullough, 1977).

The micrites of the Lower Member may be typed according to their sedimentary features. Most of the sedimentary features may be broadly categorised into cracks, clasts, or a group of diverse features that were usually found in association. Thus most micrites fall into one or more of three main non-exclusive groups: a series containing cracks and brecciation features, a wide but gradational variety of intraclastic micrites, and a group showing those diverse but associated features. As the beds are undisturbed the features of the first group result from internal contraction as opposed to external causes like tectonism or slumping. Those of the second are due to mechanical reworking, and the last group's features will be shown to belong to the paludine facies. Minor types of micrite include gastropod coquinas, oil shales, fossiliferous sapropelic limestones, and rocks with fabrics caused by diagenetic recrystallisation. For the most part, fossils, dolomite, and chert occur without regard for lithology (see the sections on dolomite and chert). Fossils include abundant gastropods, common
ostracodes and pelecypods, locally common Chara and macrophyte fragments, and some fish and reptile bones.

Internally Contracted Micrites

Micrites affected by internal contraction, a major lithologic series, include mudcracked and brecciated micrites. The "standard" polygonal V-shaped mudcracks in the Flagstaff are not very common and are not easily recognised. Some V-shaped mudcracks occur in the Middle Member (pl. 1B). More common are non-polygonal, incomplete, nonuniform mudcracks (pl. 2E&G). Most of these are preserved in dolomite or chert, and are filled with fine-grained dolomite. Polygonal mudcracks, filled with intraclastic and bioclastic debris, are also recognised in Upper Member bedded cherts (pl. 2B). Mudcracks in the Lower Member and lower Middle Member generally occur as 1 to 6 cm V-shaped wedges of micrite in beds of laminated calcilutite. These wedges commonly project from the base of the superjacent bed (pl. 1 E,F,&G), but they may "float" in the calcilutite, cropping out as isolated, unattached, upside-down triangles. They are oddly distinctive because the calcilutite has been greatly compacted while they have not. The laminae are "bent" around many of the nodules, suggesting that lamination resulted from compaction. That they are polygonal can rarely be seen (pl. 1F).
But, as noted by Smoot (personal communication), mudcracks are rarely simple, commonly involve complex collapse and cracking of the muds surrounding the cracks, and are further distorted and complicated by later compaction.

Brecciation, as seen in the Flagstaff, has two different styles, both ubiquitous. Style of brecciation is size dependent. Fine-scale brecciation produces fine cracks which are generally invisible in hand sample, are filled with spar only, and give the rocks a mosaic-or jigsaw-like appearance. These voids are planar, their sides are parallel and smooth and they occur as one or more series of subparallel sets, cut the rock irregularly, or are a network of many short, and flat or curved cracks; respectively joint, skew, and craze planes (Brewer, 1964) (pl. 3A&G; 4H; 6B,D). The coarser-scale voids are vertical or horizontal cracks or irregular closed vugs; respectively prism cracks, sheet cracks, or shrinkage pores (Fischer, 1964) (pl. 2C,F; 5C; 6D). The rock is less jigsaw-like; the pieces are less easily reassembled by the imagination. The cracks have irregular edges and are filled with a mixture of fossil fragments, extra- or intraformational clasts, calcilutite, and spar.

The larger cracks are either complex mudcracks (pl. 2E) or, as shown by intermediate types, result from secondary collapse of the walls of the mudcracks (pl. 5C), regularly obliterating the original mudcrack. The smaller ones, while
in places demonstrating creation by secondary collapse of the mudcrack walls (pl. 3A), are generally completely unrelated to desiccation cracks. Freytet (1973) distinguishes the two styles in that the small scale cracks are not part of a mudcrack-like polygonal network like the larger scale sheet and prism cracks.

All these features are produced by internal contraction, commonly by dewatering. Contraction will occur irrespective of sediment composition (Reeves, 1968, p. 90), as long as the sediment behaves like an homogenous mud. It occurs above the water table because of drying by the sun (soft flocculant calcareous muds will crack readily on exposure - Kindle, 1923) or because of soil forming processes (Brewer, 1964). It occurs above or below water by syneresis (Jüngst, 1934; White, 1961), however whether syneresis can occur in carbonate muds is questionable. Jüngst (1934) considered only the interaction of gels of expandable clays (smectite) and saline solutions; White (1961) noted that syneresis occurred when any platy clay minerals were flocculated in a saline solution, because the clays are sedimented not as individual crystals resulting in parallel orientation and fissility but as large and puffy hydrous flocs that later dewater. The ability of calcite to flocculate is unknown although CaCO$_3$ colloids can complex with organic colloids or clays (Wetzel, 1975). However, the question of syneresis may be irrelevant here because desiccation can produce all
the features claimed for syneresis (Smoot, personal communication) and carbonate muds crack very readily in conditions of low salinity (Baria, 1977). Freytet (1973) considers the smaller cracks to be formed above the low water table. In the Lake Flagstaff beds, small cracks are most numerous in areas with paludine features, but, like the large cracks, they are present throughout.

Dewatering can also produce the rare and amazing, postburial, in situ, horizontal pulling-apart of a thin layer of micrite in an otherwise undisturbed bed (pl. 2A; see also Van Houten, 1964, fig. 8).

Mechanically Reworked Micrites

Intraclastic micrite is perhaps the dominant rock type in the Flagstaff Formation. The subtypes are many and varied, in part because the clasts themselves differ greatly in size, shape, colour, angularity, and deformation. Although intramicrites as a group are abundant in many lake deposits (Leamer, 1900; see also Vatan, 1939), the subtypes are not obviously similar and have usually been treated as separate and distinct rock types, due to the absence of intermediates (resulting in a plethora of names in the literature - see Cayeux, 1935a; Boucart, 1946; Schäfer, 1974; and Williamson & Picard, 1976).

In the Flagstaff as in other lake beds, these clasts are all intraformational as is shown by their lithologies,
colours, and the distinctive fossils, textures, and sedimentary features that they contain. Frequently the source bed may be identified. The different subtypes represent the combined effects of erosion or carbonate material in varying degrees of consolidation and lithification, varying extents of transportation and mixing of lithologies, and subsequent diagenesis.

The rocks are polylithologic (pl. 4A,F,G) or monolithologic (pl. 4B). Most commonly clasts from two or three sources are mixed, but in some beds the clasts were derived from the bed below. The size, angularity, and deformation of the clasts depends on the degree of lithification before erosion and the amount of reworking or transport of the clasts. Freytet (1973) explains that originally non-lithified material disintegrates to mud on erosion and transportation, whereas well-lithified material results in angular clasts. Variation between these extremes is continuous. Very angular clasts derived from very well-lithified material are uncommon in the Flagstaff, but are present. Most clasts are rounded and slightly deformed, indicating erosion of moderately soft, partly lithified carbonate mud. Significant degrading may occur after erosion. Less than 1% of the clasts have been squeezed into pore spaces. These clasts were extremely soft during erosion, transportation, and redeposition. Compound clasts (rare) may form when they are rolled on soft mud. These are either single clasts that
are coated or more commonly aggregates of clasts. Also found were a very few complex compound clasts, that is, clasts containing clasts that contained clasts themselves. Compound clasts indicate a certain consistency of the clasts and the mud on erosion — too hard to readily degrade but too soft to collide elastically (Freytet, 1973). The size range of clasts (from medium pebble to clay) in monolithic, locally derived beds suggests that clasts are formed variable in size.

Mudcracking (internal contraction) is the most likely agent because, as noted in the last section, it always occurred throughout the lake basin and involved considerable complex secondary cracking and wall collapse, and could thus produce many irregular fragments. Some Lower Member beds even show clasts in the process of being lifted up from the top of the bed (pl. 5D). Perhaps because of the scarcity of polygonal V-shaped mudcracks, the distinctive flat-pebble conglomerate of Eugster & Hardie (1975) is rare in the Flagstaff. Clasts could also have formed from 1) the brecciation of the sediments due to the escape of gas (Reeves, 1968, p. 86), 2) syneretic cracking, which is irregular and nondirectional thus producing irregular clasts (Jüngst, 1934; Van Houten, 1964), and 3) mud crusts, which are platelets of mud bound by algae and floated by trapped bubbles of photosynthetically produced oxygen (thus forming intramicrites without erosion) (Fagerstrom, 1967).
Reworking can occur by waves, bottom currents, or slump (Vatan, 1939) or by sheetwash on the mudflats (Eugster & Hardie, 1975). Slump is an unlikely process in the Flagstaff because the basin floor was extremely flat (see Hunt, 1956). Waves, currents, and floods are all likely because the lake was shallow enough for waves to touch bottom; because the lake is large enough for winds or surf beat to generate currents (Platzman, 1963; Hutchinson, 1957; Komar, 1975); and because the mudflats were extensive and were undoubtedly subject to sheet floods similar to those postulated by Eugster and Hardie (1975) and McCullough (1977). However, Cayeux (1935b), Vatan (1939), and Freytet (1973) all consider reworking to be a subaqueous process.

Three main textures are formed by diagenesis of intramicrites. The abundant, small, round, intraformational clasts are commonly deformed; their edges become diffuse and they gradually become indistinguishable from the matrix, creating a faint, vaguely peloidal texture (compare pl. 3C and 5B). Another texture is formed by recrystallisation of the matrix to spar making the clasts very prominent (pl. 6C). A variety of this is caused by dissolution of the matrix creating intergranular voids. The third texture occurs in packed rocks when the clasts become diffuse and then the matrix recrystallises forming a "crumbly limestone with sparite filled stellate voids...the calcaire grumeleux of Cayeux" (Freytet, 1973) (3C,E). "Crumbly limestone" is a
poor term, because the limestone doesn't crumble. Williamson and Picard (1974) use "clotted or grumous"; "crumb-like" is also descriptive. The many apparently lithographic limestones of the Flagstaff (La Rocque, 1960), when polished, are seen to be grumelous and cracked.

In short, most of the many intraclasts appear to have been formed by the erosion, transportation, mixing, and redeposition of the micrites affected by internal contraction. Thus the two series may be genetically associated, distinguished only by the transportation of the clasts.

Paludine Micrites

Biological paludine features include snail associations, small sinuous to straight and vertical rootlet moulds, debris-filled root casts, burrows, other pedotubules, vertical moulds of stems or leaves of some macrophyte (possibly a bulrush), and rare vegetation.

The snail evidence has received much attention. La Rocque (1960) studied the species in the Flagstaff and found that many of them preferred shallow, turbid waters and soft, muddy bottoms with lots of vegetation that permits them to climb to the surface to breathe. Using associations described by Hanley (1976) in his paleocommunity analysis of Eocene Lake Gosiate and La Rocque's published collections (La Rocque, 1960) one can conclude that the Lower Flagstaff contains "pond" and littoral lacustrine snail associations
in a 4:1 ratio (plus 4 terrestrial associations). The littoral lacustrine association apparently represents nearshore, shallow, current-swept sites that are not heavily vegetated. The association "delineates shoreline fluctuations." The pond association, which is typical of most of the Flagstaff, is typical of quiet water, either ponds or protected shallow lake water with thick vegetation. Snails of this association could withstand desiccation and occasionally lived in temporary waters.

Single and dendroid circular channels (Brewer, 1964) are extremely common (pl. 1H, 2D). They are about .2 to 2 mm in diameter, branching sporadically, usually smooth-walled and sinuous, sometimes straight, and sometimes ringed by a neocalcitans (Brewer, 1964) or by a leached or iron-rich zone. In two samples they detour around what were hard clasts. They turn to avoid hard surfaces, and will reach down to clay pockets. They are interpreted as rootlets, in agreement with Brewer (1964). The pedotubules (Brewer, 1964) include large root casts that are usually meta-isotubules with neocalcitans and other subcutanic features (tubes lined internally with calcite, etc., and filled with disorganised debris from the bed above: pl. 2F, see also Freytet, 1973, p. 39) and burrows which generally are ortho-, meta-, or para-striotubules (i.e. tubes, showing "watch-glass" and similar structures, and filled with debris from the same bed, or the overlying bed, or somewhere else).
(See pl. 4D, E; 5H; and Brewer, 1964, for illustrations of pedotubules.) Para-striotubules and para-isotubules are the commonest types. Some show solution features at the edges of the tubule and a few are bounded by stylolites that have obliterated the burrow edges (pl. 5H).

Some tubes are almost all vertical and straight, about 1 to 2 dm long, several mm wide, hollow or filled with spar, circular or possibly triangular in cross-section, and present in groups of individuals spaced 2 to 10 cm apart (pl. 11). They are over ten times the size of the rootlets. They never branch, and are interpreted as stem of leaf moulds of, perhaps, bulrushes, although some, especially those depicted, are also similar to the roots of some prairie plants. Possible modern analogues to this facies are pictured on p. 148 in Veatch and Humphreys (1964) and on Plate 6 of Blatchley and Ashley (1900), both of which show large numbers of rushes, etc., each a short distance from the others, growing out of very flat, temporarily exposed, mudcracked, marl mudflats at the edges of shallow marl lakes. The term marsh is not as befitting here as weed bank. These moulds are relatively rare in the Flagstaff and occur mainly in the Lower Member at Flagstaff Peak, Cove Mountain, and New Canyon.

Physical paludine features include mudcracks, remobilisation and plasmic separation features such as marmorised or nodular limestone (described by Freytet, 1973), preburial
solution features such as knobby-topped beds (Freytet, 1973; Walkden, 1974), and pedogenic features like polyphased nodules and various cutanic and subcutanic features that can be produced by deposition of material in or around root voids as described by Brewer (1964). Mudcracks have already been described. Beds with knobby tops appear to be caused by solution. The surfaces are knobby or hummocky on a scale of several cm. In many cases, the upper parts of the beds are decalcitised; rarely, they have a crust-like surface (see Walkden, 1974). Commonly much of the remainder of the bed has been remobilised and nodulised (sensu Freytet, 1973). The dissolution must occur prior to burial, because the bases of the overlying beds are never affected. Likely agents are rainwater, overlying bodies of undersaturated water, or an oscillating phreatic water table (Freytet, 1973). In the knobby-topped bed depicted in Pl. 2C, the oscillating water table is the probable cause. The bed contains lacustrine snails and nearby superjacent beds contain rootlet moulds and possible rush moulds. The overlying bed here is a slightly down-cut channelform fluvial sandstone. Presumably the lake receded, plants grew, and a stream course developed, all of which were probably accompanied by changes in the level of the water table. Also the bed is heavily nodulised, in a manner that Freytet (1973) considers indicative of formation by water table oscillations (see also pl. 3F). Polyphase nodules, rare in the Flagstaff but
identified at Cove Mountain and New Canyon, are nodules that have undergone many episodes of growth and fissuring (pl. 5G), some of which must have taken place above the water table (Freytet, 1973). Besides nodulisation, an oscillating water table can also cause marmorisation and recrystallisation (Freytet). Marmorisation (pl. 4H,I) is a splotchy iron-oxide-colour distribution due to iron remobilisation in hydromorphic soils. Iron remobilisation is greatly increased when the oxygen content and Eh of the substrate fluctuate around zero (Golterman, 1967). Colours involved are pink, red, green, yellow, orange, and purplish-blue (Freytet). This is present in some of the variegated calcilutites.

Curvilinear fissures and craze planes can develop in an homogenous unit in zones of oscillation of the water table. This can cause the in situ production of a "crumb-like" limestone from an homogenous micrite (pl. 3G) (Freytet, 1973); Bourcart (1946) claims this is the origin for Cayeux's (1935a) "structure pseudobrechique." Unequal recrystallisation, sometimes resulting from or in concert with fissuring, may also produce "clasts." Pseudoclasts may also be produced by calichification, a very similar process (Walkden, 1974; Reeves, 1976). Either calichification or fissuring and (or) recrystallisation may virtually create a crumb-like limestone with stellate voids, differing from the previously described diagenetic texture by originating from an
homogenous sediment.

Any or all of these paludine features, when abundant, strongly suggest shallow water. Evidence for weed beds suggests water no more than 3 m deep, the lower limit for surface-attaining plants (Dussart, 1966), whereas the pedogenic features indicate formation within or above the zone of oscillation of the water table. These features were present throughout the Lower Member, however their abundance and association along the east side of the Wasatch Plateau suggest that the eastern shore of the lake was primarily a very wide and shallow weedy shore. Indeed their abundance throughout the lake basin and the abundance of mudcracks and intramicrite indicate that the lake was everywhere very shallow and that the lake floor was frequently exposed.

Minor Lithologies and Features

Other lithologies and features common in the Lower Flagstaff include oil shale, sapropelic limestone, charophytic limestone, flat bedding surfaces, and solution features. Low grade oil shales and sapropelic limestones (dark, nonlaminate, hydrocarbon-rich, gastropodal limestone; pl. 4G, see Freytet, 1973) are abundant in the lower third of the Lower Member only. They are pyritic and lack rootlets, stem moulds, and high energy indicators. They differ in lamination and snail content, but are interpreted to be an accumulation of organic material which probably built up
more than a few metres below the lake surface. Accumulation of organics may occur in several ways. When large amounts of organic debris settle below the thermocline and deplete the hypolimnetic oxygen, decomposition is stopped. Preservation may occur in a chemically stratified lake. If mixis is prevented, the salinity of the monimolimnion may be sufficient to prevent bacterial and fungal decomposition (Wetzel, 1975). An accumulation of organic matter can also preserve itself by adsorbing the nutrients nitrogen, calcium, and phosphorus on humus and by the release of phenolic compounds, which are abundant in humus. The first process removes nutrients required by saprophytic bacteria and fungi, the second inhibits them (Kukkonen, 1973).

Some micrite contains large fragments of Chara (Freytet, 1973). Chara is very readily comminuted to small crystals, therefore large amounts of Chara stems and oogonia indicate that the micrite was deposited near to or where the Chara grew. Chara grows best in the lower sublittoral zone or below, where it is too deep for plants to reach the surface but above the light compensation zone (Dussart, 1966). Therefore this type is also indicative of deeper water.

Lithologic changes are gradational or abrupt and, when abrupt, frequently occur across sharp, distinct diastems. These bedding surfaces are commonly very flat, although they can be undulatory or knobby. As mentioned before, knobby surfaces are are interpreted as solution/remobilisation
features (Freytet, 1973; Walkden, 1974). These surfaces are laterally extensive, and can be visually traced for several kilometers across cliff faces (see Figure B plate 1 and Figure 9, a correlation chart of the same area); however, the beds between the surfaces change laterally in nature and thickness. Lateral changes such as fossiliferous to unfossiliferous, monolithic to poly lithologic, micrite to blocky calcilutite, olive to gray, and + 40% of the bed's thickness are common. This suggests that each surface is the result of a synchronous event that affects different facies equally. The most likely causes of these surfaces are a shift in the wave zone caused by regression or transgression of the lake, erosion by floods, and an increase in wave action due to storms. Regression is supported by the extent of mudcracking and brecciation and by the presence of preburial lithified surfaces; transgression and sheetwash floods are supported by the filling of mudcracks with debris and the frequency with which intraclasts are derived from the underlying bed, indicating erosion. Sheetwash is also supported by the frequent input of dolomite clasts from the mudflats around the lake (see the section on the dolomite). Storms are also supported by the extent of the reworked material. The volume of lake water does not necessarily have to change to produce temporary transgressions, because winds can very effectively move water off or onto marginal mudflats. Wind set-up in Lake Erie, a lake comparable in size
to Lake Flagstaff, under extreme conditions may amount to 4 m (Platzman, 1963). Winds can easily move large bodies of water across a playa (a wind of 70 km/hr was observed to move water at 2 m/minute, whereas even gentle breezes under 10 km/hr can move water a metre every 12 to 15 minutes [Motts, 1970]).

Solution features are very common in these rocks. Many formed prior to lithification. Shells are commonly dissolved (pl. 5B), frequently while the mud was still soft: partial shell solution and obliteration of the resulting voids are seen in many rocks. Similarly, rootlet voids can be partially obliterated by failure of the void walls. Stylolites sometimes form where the two walls become contiguous, in which case the stylolite may be quite early. Stylolites are especially common at the edges of burrows (pl. 5H) and mudcracks, presumably triggered by the difference in permeability between the micrite and the detrital burrow and mudcrack fillings. These are undatable. Solution stringers and poorly developed stylolites are also common (pl. 3F). In most beds, the edges of peloids are at least slightly diffuse from dissolution. Freytet (1973) demonstrates that this occurs prior to lithification. Neomorphism of intergranular micrite produces many stellate "voids" (Freytet). In the more extreme cases (which are associated with dolomitisation) the micrite matrix between the clasts is mostly dissolved, leaving large vugs. In these
cases dissolution occurs when the sediment is relatively solidified, resulting in hollow fossil moulds. However, partly because only thin units are vuggy and dolomitised, it requires that dolomitisation and presumably solution were in each case relatively early (see the section on dolomite).

Dolomicrites of the Middle and Upper Members

The Middle Member

By 114 m above the base of the Flagstaff Formation at Cove Mountain, snails disappear and variegated calcilutites, increasing upward, pass into a 24 m section of red mudrock. A thin carbonate bed at 115 m contains some small detrital clasts of gypsum. These three characteristics are taken to mark the base of the Middle Member, studied by McCullough (1977).

The Middle Member contains many of the lithologies and features present in the Lower Member, but not all, and also contains some new ones. Features are faint or absent, due to their destruction during dolomitisation (see section on dolomite). Rocks include essentially featureless massive white dolomite, tripoli (an incompletely silicified limestone whose limestone has been leached out), and intradolomiticrite (containing mostly clasts that were hard before erosion). As previously mentioned, mudcracks are very common, either as mudcrack fillings, simple or complex cracks, or fillings distorted into wavy lenses by vertical
compression. Detrital and authigenic intercalary gypsum are rare but present in the dolomicrite (see McCullough, 1977). (Intercalary crystals are ones grown in the matrix – Brewer, 1964.) Gypsum seems to occur on two levels – the major 12 m (plus or minus) bedded gypsum in the upper unit of the Middle Member, which is traceable over the central Wasatch Plateau, and a lesser zone of intercalary, secondary, and detrital gypsum that, from a study of the theses shown in Figure 10, seems to exist over a slightly smaller area. Many of the lower dolomicrite beds are vuggy, probably associated with dolomitisation as in the Lower Member. The pores are both intergranular and mouldic (Choquette & Pray, 1970) and commonly contain chert (above this, chert occurs as mudcrack fillings and, higher up the section, as nodules that replace the micrite). Peloidal or grumelous dolomicrites are common. Solution features are evident, including such ones as pods separated by calichification (Reeves, 1976) or by solution stringers (see incipient nodular bedding in Pettijohn and Potter, 1964). The latter are zones of wavy black lines of insolubles that have been concentrated between the carbonate pods during early burial calcite remobilisation. Poorly developed networks of both stylolites of uncertain timing and early or pre-burial’ craze and skew planes are common. Layers and beds are thin and duricrusts are common (as exemplified by one rock with two 2 cm layers of rootlets each covered by a dense impermeable
mudcracked white carbonate crust presumably because the lake was temporary and variable. Numerous rootlet moulds, ostracodes, burrows, and much organic debris (but no gastropods) are preserved in the chert, indicating that the lake was unfavourable for snails but was still productive. Gastropods are present in one or two horizons in the Middle Flagstaff - see Figures 6 & 7 - which are associated with carbonates that are less than 100% dolomite/total carbonate and are interpreted as brief periods of freshening of the lake water. The upper snail zone in Ephraim Canyon is a thin sandy limestone and a channelform sandstone, and is believed to correlate with a "biological delta" observed by Fagadu (1949) and Gill (1950) near the mouth of Manti Canyon.

Caliche features are moderately common in the Middle Member at Cove Mountain and elsewhere. Most common are caliche breccias or conglomerate (pseudoclast and pseudo-conglomerate), caliche crusts, pseudobedding, plates, blebs, and nodules (Reeves, 1976). Calichification can be considered an arid version of the oscillating phreatic water table, although in mechanism, calichification concerns capillary water and evaporative draw rather than groundwater flow and meteoric influx. The most striking resemblance between Flagstaff features and caliche features is in pseudo-bedding and plates, which are flatlying layers ranging in thickness from 1 mm to 5 cm. They strongly resemble the
tops of many dolomite beds from the Middle Member, particularly those which are flat but crinkled and laterally impersistent (pl. 1J). Pseudoconglomerate may occur in a few beds. Algal laminae are locally common but the laminae in some Middle and Upper Member rocks from Ephraim Canyon strongly resemble caliche crusts (see Reeves, 1976, fig. 3-5B; also Walkden, 1974), although they may be algal too. Caliche laminae form when the unit is exposed and the caliche is near enough to the surface to receive infiltrating water. The impermeability of the caliche forces the meteoric water to move laterally, creating layers. Apparently, extended direct exposure of a laminated zone causes fracturing of the laminae (Reeves, 1976); this is a possible source for some of the small intraclasts (see McCullough, 1977, p. 43). Fabrics formed by recrystallisation (Freytet, 1973; Reeves, 1976) are probably common, though not readily identified. One unusual rock contains both moulds of large, euhedral, intercalary evaporite crystals and diffuse, irregular, yellowish-grey plasmic concentrations (Brewer, 1964) of carbonate in a light bluish-grey matrix. These concentrations seem to be "blebs," a unique caliche feature (Reeves, 1976).

In short, calichification was probably very important in the Middle Flagstaff. This is in agreement with the several indisputable caliche features found, with the concept of wide, flat, and frequently exposed mudflats surrounding
the lake, and also with the interpretation of the Middle Flagstaff as a restricted evaporite-precipitating lake phase during more arid conditions, which are suggested by the gypsum, the abundance of redbeds around the edge of the lake (Figs. 6, 7, & 8; cf. Johnson, 1949; Baughman, 1959; and Bachman, 1959), and the lack of snails (McCullough, 1977).

The Upper Member

The Upper Member begins with the return of gastropods and the recurrence of slightly calcareous dolostones. The only decent, complete exposure of the Upper Member in the central Wasatch Plateau is at Ephraim Canyon, so this part of the Flagstaff received relatively little attention. Dolomicrite is less common than light grey blocky calcilutite. Micrite beds are commonly silicified, preserving sedimentary features that show: mudcracks, brecciation, and ostracode- and snail-rich horizons. Shells are commonly concentrated in pockets or layers. Cherts are also intraclastic, grumelous, and fine-grained. In contrast, dolomicrite is generally atextural or at least exhibits many solution features (e.g. vugs, stylolite networks, and inipient nodular bedding). Dolomite is bleached white (although some has secondary limonite stains) and is platy. Gypsum is absent, but paludine features are less common than caliche features. Compared to other members, these rocks contain more epiclastics, being richer in quartz sand and
silt, and clay galls. Five very low grade oil shales are present.

It is likely that these rocks were deposited at the outer reaches of an ephemeral lake or, as Pagadau (1949) and Gill (1950) suggest, on the stream- and pond-filled floodplain above. Flooding would have moved the sediment around, filled mudcracks, etc., and water plants and ostracodes could have flourished. Snails may have required longer periods of inundation, explaining their sparsity compared to the ostracodes. On the retreat of the lake, mudcracks formed and groundwater would be drawn up through the new sediments by evaporation. This caused calichification, and silicification when the groundwater and (or) the lake water were silica-rich (see section on chert). After silicification, dolomitisation occurred. This produced the featureless, atextural bleached dolomites that are so abundant in the Upper and Middle Members. The facies of the Upper Member at Ephraim Canyon are possibly similar to those facies present east of the Wasatch Plateau that produced the dolomite intraclasts of the Lower Member at Cove and Ferron Mountains.
FORMATION OF MICRITE

The origin of marine and lacustrine micrite has long been disputed; however, the question of the origin of marine micrite is mostly settled now by work done during the last decade (cf. Broecker and Takahashi, 1966). CaCO₃ formation in lakes is far less understood than in the ocean. This is to be expected because marine carbonates are vastly more prevalent than lacustrine ones, and have thus received more attention. Also marine conditions are more constant; temperature, salinity, Eh, pH, solutes, and ion activity product of the ocean are much less variable than in lakes, often even less variable than in any one lake alone. The inherent inconstancy of freshwater conditions makes modelling of lacustrine carbonate formation much more difficult. It also means that findings in any one lake system are not necessarily generally applicable.

The many different origins proposed for lacustrine carbonates involve mechanisms that are biological, biochemical, physicochemical, and (or) chemical in nature, depending on whether organisms precipitate the calcite directly, or whether calcite precipitates spontaneously. Summarised
differently, they include deposition of calcite by organisms, by loss or extraction of CO₂ from water, by concentration of Ca²⁺ and HCO₃⁻, and by mixing of different waters. In the next several pages, the various possible origins will be discussed. The general mechanisms of formation will be shown to be controlled primarily by the degree of saturation of the lake water and also by the productivity of the lake.

Previously Proposed Origins

Mixing of different waters occurs when normal Ca²⁺ and HCO₃⁻ - rich rivers or floods flow into alkaline lakes. Because CaCO₃ has a low solubility at high pH values, equilibration between the waters (particularly slow equilibration) causes aragonite precipitation. This is purely a chemical mechanism and occurs mostly in highly alkaline and saline lakes where the lake water is very different from rain water and runoff. Examples are Van Gölü, (Müller et al., 1972) and Lake Magadi, Kenya (although Eugster, 1970, interprets most calcite precipitation there in terms of evaporative concentration). This is also known to occur in Mono Lake where cold HCO₃⁻ - rich springwater enters the lake and creates tufa mounds (Dunn, 1953).

Concentration until precipitation occurs is also a chemical process and may be caused by freezing or evaporation. Hallet (1976) notes the first causing precipitation
of CaCO₃ below glaciers. The second is more common. Müller et al. (1972) note it as the primary mechanism in Tuz Gölü; Hanley and Steidtmann (1973) invoke evaporative concentration to produce limestone laminae in the Caspar formation.

Physicochemical mechanisms involve the loss of CO₂ (degassing), which is caused by changes in physical conditions such as pressure or temperature, as occurs at Plitzvíčkých jezera (Müller et al., 1972). Degassing of CO₂ decreases its partial pressure, which causes dissociation of H₂CO₃ to CO₂ and H₂O. This in turn causes 2HCO₃⁻ to react to produce H₂CO₃ and CO₂, the latter of which reacts with Ca²⁺ to form calcite, if the water is saturated with it already. Each mole of CO₂ lost precipitates 1 mole of CaCO₃.

Thiel (1933) describes precipitation near to artesian springs because the pressure drop allowed degassing; however temperature change has a greater effect because CO₂ solubility decreases rapidly as water warms up. This is mentioned in such early works as Blatchley and Ashley (1900) who ascribed precipitation around springs to warming of cold CaCO₃ - saturated groundwater, Davis (1903) who noted precipitation of tiny calcite crystals in spring-fed marl lakes on warm days, and Kindle (1927, 1929) who concluded that precipitation of calcite was due to the warming of shallow waters or the epilimnion of stratified lakes. In Oklahoma, travertine formations up to 30 m thick form at falls and rapids in streams rich in Ca²⁺ and HCO₃⁻, because
the agitation promotes warming and also evaporative concentration of spray (Emig, 1917).

The effect of warming has prompted Blatchley and Ashley (1900), Bartosh (1971), and Schäfer (1973), studying postglacial deposits in Indiana, the USSR, and Switzerland respectively, to view marl formation as being linked with climatic warming over decades or centuries. This is too simplistic – as soon as the climate warms enough for rains in winter and warm days in summer (i.e. for input in winter and precipitation in summer) the calcite equilibrium is determined by daily and seasonal fluctuations in supply and temperature. In northern lakes, the major degassing occurs after the break-up of winter ice, allowing the escape of CO₂ which had built up through the winter; this degassing is abetted by CO₂ removal by the spring algal bloom (Wetzel, 1975).

Extraction of CO₂ by plants for photosynthesis causes biochemical precipitation. Extraction can be by macrophytes (as at Obi-i-Istada), microphytes (as at Lake Balaton), or both (Müller et al., 1972), and it may be seasonal (during the spring green algal bloom and again during the summer bloom when blue-green algae may use up all the CO₂ (Wetzel, 1975)) or it may be daily. Dean and Megard (1973), working in Minnesota, proved that the daily rates of CaCO₃ precipitation are directly related to daily rates of planktic carbon dioxide assimilation, i.e. to phytoplanktic
productivity (Wetzel, 1975). Note that if only one could prove this origin and show that all CaCO₃ precipitated was deposited and preserved (see Alexanderrson, 1976, and Degens & Stoffers, 1976, for recognition of submicroscopic solution features) then this work could provide a paleolimnological index of productivity. Regardless, this is a very common mechanism, operating both on its own and with physicochemical precipitation (as demonstrated by Terlecky, 1976, and Moxham & Eckhart, 1956).

Dean and Egglestone (1975) show a biochemical origin for calcite in meromictic Fayetteville-Green Lake, near Rochester, New York, the site of Terlecky's marsh. Calcareous algae produce most of the sediment and even form algal reefs (noted elsewhere by Blatchley and Ashley, 1900). They precipitate most of the calcite, but some is also produced physicochemically in the pores of the reef. Takahashi et al. (1968) also studied Green Lake and showed by isotopic evidence that much of the calcite suspended in the lake was precipitated in the surface waters. This is unlike the ocean where the whitings, once thought due to physicochemical precipitation, are due to resuspension of skeletal fragments (Broëcker and Takahashi, 1966). The combination of the studies by Dean and Takahashi suggests that much of the calcite precipitated at the surface is redissolved before reaching the bottom.
In many hardwater lakes the marl comes from *Chara* and other calcareous algae (e.g. *Nitella* and *Tolypella*). They can precipitate calcite internally or externally by changing the pH of their microenvironment by $\text{HCO}_3^-$ uptake and $\text{OH}^-$ extrusion. Their importance in sediment formation was recognised by Davis (1900, 1901, 1903). Cucci (1976) studied production of sediment by *Chara* in Preble-Green Lake and others in New York. The CaCO$_3$ there comes from glacial drift and limestone bedrock and enters the lake through springs, which also bring in enough gypsum to make the lakes meromictic. The pH is 8-9, although *Chara* will grow in waters between pH = 5.5 and 9.5. *Chara* fragments comminute to $0.25 \text{ mm} - 3 \mu\text{m}$, and their accumulation may amount to $0.32 \text{ mm/year}$ (Cucci, 1976).

The biological mechanisms involve mostly formation of shell material by molluscs and ostracodes, although also included in this category is formation of calcite by bacteria in any one of several ways, e.g.

$$\text{Ca(HCO}_3\text{)}_2 + 2\text{NH}_4\text{OH} \rightleftharpoons (\text{NH}_4\text{)}_2\text{CO}_3 + 2\text{H}_2\text{O} + \text{CaCO}_3$$

(Williams & McCoy, 1934) or by anaerobic sulphate and iron bacteria, as observed by Bavendem (1931). However this is not a major source.

It is possible for limestone to be derived solely from skeletal debris. Weber's (1964) study of carbon isotopes in Lake Flagstaff indicates that the limestone could have originated as molluscan shells, which, if true, indicates
that the lake was very productive. Indeed, carbon isotope ratios can distinguish between materials from two different genera of bacteria or snails, or even between a small number of specifics in a known system. However, comparison of work by Craig (1952), Wickman (1952), Rosenfeld & Silverman (1959), and Pardue et al. (1976) (and others) shows that the ratio ranges are so broad as to be useless in a less than fully known system, even before considering diagenesis. Whelan & Roberts (1973) found that $\delta^{13}C$ values are a function of organic debris in the sediment and not depth of burial or temperature as might be expected. This is a great disappointment to the paleolimnologist.

The great variety of possible mechanisms can perhaps be explained after close examination of the literature. Note that all the workers proposing purely chemical precipitation (Eugster, 1970; Müller et al., 1972; Hanley & Steidtmann, 1973) are studying conditions of extreme aridity, salinity or pH which are not conducive to high productivity and are naturally rich or physically enriched in Ca and $\text{HCO}_3^-$ ions. Most of the studies favouring a wholly or partly physicochemical mechanism were done in areas with groundwater rich in $\text{Ca}^{2+}$ and $\text{HCO}_3^-$. Correlations have been noted between marl lakes and porous calcareous outwash (Thiel, 1930), areas of limestone outcrops (Thunmark, 1937; Matthews & McCammon, 1957), and even between the thicknesses of the marl and of the glacial drift (Blatchley & Ashley, 1900).
Lakes where precipitation is wholly a physicochemical process (CO₂ loss due to change in temperature or pressure) are fed by springs bringing up water that is supersaturated with calcite (e.g. Blatchley & Ashley, 1900; Davis, 1903; Terlecky, 1976). (Terlackey's spring water was fivefold supersaturated.) In contrast, purely biochemical precipitation (CO₂ removal required to precipitate calcite) occurs in lakes whose waters are at or close to saturation (e.g. Davis, 1900, 1901; Dean & Megard, 1973). Here organic activity is needed to cause supersaturation. The water entering these lakes is less calcite rich, either because it is mostly surficial or because the substrate is less calcareous. Along this line of reasoning, lakes that are always undersaturated would have only calcite that was derived from shells.

Note that Terlecky (1976) attributed about 15% of his marl to Chara and molluscs. If the lake was as productive as nearby Fayetteville - Green Lake and Dean and Megard's Minnesota lakes (a reasonable assumption) and the 15% represents full production of calcite by organisms, then the lake is producing about six times as much marl as it would were it not over five times supersaturated. In other words, whenever Ca and HCO₃ ions are present in amounts both above saturation and greater than the amount of CO₂ that the biota can use then a proportional amount of the calcite is precipitated physicochemically (see figure 12 for a
Origins of CaCO₃

Figure 12

Biota important

Primary Productivity
very high
high
medium
low

CO₂ extraction
biochemical
(photosynthesis)
biological
(skeletal)
no CaCO₃
undersaturated

CO₂ loss & extraction
biochemical
periodic physicochemical
(changes in P-T)
supersaturated

Biota non-existent

CO₂ loss
chemical

physicochemical

Complete suppression of biota
(position dependent on pH salinity temp.)

saturation with respect to CaCO₃
diagrammatic summary).

An Hypothesis about the Origin
Lacustrine Limestone

Figure 12 is designed to indicate the probable mode of CaCO$_3$ formation in a lake, which is a function of degree of saturation and productivity. The graph is qualitative and not quantitative, because Mg concentration (Pytkowicz, 1965), salinity, colloid reactions, and other factors greatly affect the numbers and because some of the boundaries are gradational. First the diagramme will be discussed, then it will be applied to the Flagstaff.

A major consideration is whether or not the water is supersaturated. If concentrations are such that most of the time the lake is supersaturated then inorganic or physico-chemical precipitation will occur. If the water is not normally or potentially saturated then CaCO$_3$ can not be physicochemically precipitated, but will only be precipitated because of the actions of organisms. When the water is just undersaturated, photosynthetic removal of CO$_2$ will result in CaCO$_3$ precipitation (Dean and Megard, 1973). At very high productivity levels plants will compete with physicochemical reactions (degassing) for CO$_2$, thus the boundary to the left of which there is no physicochemical precipitation leans slightly to the right. In further considering the undersaturated or left side of the graph, it is
obvious that as CaCO₃ (aq) levels drop one eventually reaches the condition where the biota will not extract enough CO₂ to attain saturation of the lake. This is obviously dependent on the productivity of the lake - the more plants the greater the potential for extraction of CO₂, thus with high productivity, water with relatively low concentrations of CaCO₃ may still become supersaturated. Therefore the left hand boundary of the biochemical field climbs to the left. Below this is a region in which CaCO₃ is never directly precipitated but is still deposited and maintained in solid form by animals, notably ostracodes and molluscs. At extreme undersaturation not even molluscs can precipitate calcite.

On the right or supersaturated side of the graph photosynthetic precipitation is still important, at least at low levels of supersaturation, but in terms of calcite produced it is supplemented or exceeded by physicochemical precipitation. At low CaCO₃ (aq) levels, physicochemical precipitation occurs only on the warmest days of the year, but in lakes with increased concentrations of Ca²⁺ and HCO⁻₃ and (or) with higher temperatures, physicochemical precipitation becomes more regular - seasonal, even daily in the tropics. Obviously, at increasingly high saturations, physicochemical precipitation becomes more and more important. At some point hardness will begin to inhibit vegetation and animals. From this point on, the maximum possible productivity will
begin to decline until the lake is almost barren.

pH is the most important variable in considering chemical (and some physicochemical) reactions. A pH change of one unit changes calcite solubility by two orders of magnitude (Garrels, 1960, pp. 54-56; see Figure 13a). In alkaline hardwater lakes, daily metabolism can vary a lake's pH between 8 and 10 or more (Wetzel, 1975, p. 178). pH changes can also result from dilution by rain (mixing of waters) or by concentration of HCO$_3^-$ or of cations by evaporation or concentration. Less marked but still potentially important in chemical reactions is the positive effect of adding ions and raising the Ion Activity Product (multiplying it by 100 doubles the solubility of calcite) (Berner, 1971).

The other part of the diagramme shows probable cause of precipitation in lakes that are under extreme conditions and are thus unproductive. The factors, pH and salinity, are only symptomatic of the environment; they are not the underlying cause of precipitation as were saturation and productivity. High actual or potential supersaturation is assumed. Under the less severe conditions, CaCO$_3$ is precipitated by changes in pressure, temperature, and pH. Increases in pH and salinity indicate evaporation, which causes concentration and ionic complexing and/or precipitation (Eugster, 1970). At very high salinity and pH, incoming fresh water will flow out over the lake water thus
causing precipitation by the slow mixing of waters of different pH (Müller et al., 1972).

Source of Ca$^{2+}$ and CO$_3^{2-}$ and Origin of Micrite in Lake Flagstaff

The previous section summarises the modes of formation of lacustrine micrite. The next two questions to be dealt with are how the limestone formed in Lake Flagstaff and whence the constituents came. Much of the Ca$^{2+}$ and CO$_3^{2-}$ undoubtedly came from the surrounding rocks, which are very rich in limestone. (The Paleozoic section that forms the mountains west of the thrust belt is particularly rich.) However, because none of the possible sources are free of sand and because epiclastic sediment in Lake Flagstaff is minimal, it is implausible that clastic input of allochthonous detrital limestone was a major source of micrite; therefore, most of the calcite probably entered the lake in solution. Part of the calcium and bicarbonate must have been transported in streams and part was probably carried by groundwater, entering the lake through springs and seeps. Springs are expected where the sediments are fine-grained (Reeves, 1968), but calcite-rich springs normally create massive tufa mounds as in Mono Lake (Dunn, 1953) and in the Kharga Oasis area, Egypt (Caton-Thomas & Gardner, 1932), but none have yet been observed in the Flagstaff Formation. If the groundwater entered the lake by
seepage (the Flagstaff rocks though fine-grained are of moderate porosity due to fissures, voids, and moulds), then no such evidence would have been left (Leakey, 1931).

Possible mechanisms of formation of micrite in Lake Flagstaff are: from comminution of shell or Chara debris, by precipitation due to photosynthetic removal of CO₂ or degassing due to rising temperature, or by increase in pH by evaporation. Mixing of waters could not have occurred in the fossiliferous stages and only possible several times in the middle phase because the required salinity would have killed the snails and ostracodes.

Evidence for the origin from shells comprises 1) the known abundance of snails and ostracodes and 2) the carbon isotope data of Weber (1964). Snails undoubtedly contributed to the limestones: in the richest coquinas, recognizable shell material makes up 25% of the rock (La Rocque, 1960). It is unlikely, though, that all the limestone came from shells. Using a conservative volume of 1 x 10¹¹ m³ for the Lower Member rocks and assuming that the medium-size snail (3cm³) is 90% hollow and that the Lower Member lasted 7 million years (half the Paleocene) approximately 4 x 10⁹ adult snails must have died each year. Also, much of the limestone was formed when the lake contained no snails.

Weber's (1964) isotopic data show that the ratios for the snail shells and the limestones fall in the same general wide ranges, although the ranges of values are extremely
broad and seem related to diagenesis.

Similarly *Chara* certainly contributed a little calcite. A few beds are rich in *Chara*, although none are packed. *Chara* is most common in the lower part of the Lower Member although even there it is not abundant. Overall, recognisable *Chara* fragments were found in less than 3% of the beds. Had *Chara* been a major source of micrite mud, it should be more prevalent.

This leaves photosynthetic removal of CO₂, degassing due to rising temperature, and pH changes as probable causes. The lake was certainly very productive and weedy during the early substage as the abundant snails, root moulds, and organic material attest. The lake was probably episodically productive during the middle substage, as evidenced by preservation of rootlet moulds and organics in chert, some algal-laminated sediments (Weiss, 1969; McCullough, 1977, p. 59), and ostracode blooms. The late substage was probably less productive than the first. Temperature changes were probably very important through the history of the lake because the lake was shallow and would have been subjected to great and continual temperature fluctuations. However, the main cause was most likely change in pH, given the ease with which pH changes can drastically affect calcite solubility. Inorganic precipitation is supported by tufa on twigs stuck upright in the mud and on stones that were
perhaps too large to be moved by currents and most waves (see Figure E plate 5 for a large oncolite, also Eardley, 1932).
FORMATION OF THE DOLOMITE

Dolomite is abundant in the Flagstaff, being present in all the members and constituting almost half the carbonates on the Wasatch Plateau, according to our measured sections. The Middle Member contains the most dolomite – McCullough shows that almost all the Middle Member carbonates are pure dolomite – and the Upper Member is also dominantly dolomitic. Rocks of the Lower Member are less dolomitic, varying between 0 and 100% dolomite/total carbonate. At Cove Mountain the lowest 35 m of the lacustrine beds, except the basal 5 m, are 100% calcite/total carbonate; above this, the dolomite increases irregularly to 100% of total carbonate by the top of the member. Vertical variations are so extreme that two adjacent beds may differ in their percentage of dolomite by more than 90 points.

The Pure and Calcareous Dolomites of the Lower Member

In the Lower Member, the very pure (>95%) dolomite beds are distinct from the others. They are generally a bleached white (approximately N9), and very vuggy due to the
dissolution of any fossils and fine micritic matrix. On close inspection most contain faint ghosts or remnants of the structures and fabrics commonly found in the micrite (fossils, "crumbs," rootlet moulds, stem or leaf(?) moulds, burrows, cracks, etc.), all of which lead to the conclusion that the dolomite has replaced the limestone and has largely destroyed its original texture and fabric.

The impure dolomites (>10% calcite) are, however, indistinguishable from the dolomite-free carbonates around them. Therefore the dolomitisation of the mixed dolomite/calcite beds is somehow different from that of the pure dolostones. If preferential dolomitisation can be ruled out or allowed for (e.g. Schmidt, 1965), then, as Lumsden (1976) points out, because dolomitisation is unlikely to be partial, intermediate dolomite values result when a chemical change arrests the process or when dolomite and calcite are mixed.

Physical mixing of the dolomite and the calcite seems to have occurred in the Flagstaff. Some of the intraclastic micrites show mixing quite well in that frequently the percentage volume of the dolomite clasts is approximately equal to the percentage of dolomite in the rock. This is best seen in rocks that have readily identifiable white dolomitic clasts such as the one illustrated in Figure B, Plate 4. The relationship between dolomitic clasts and the analysed dolomite content of the rock is not perfect because 1) part
of the detrital fraction is microscopic and 2) subsamples may not have been representative. The colour relationship, while dramatic, is less accurate because not all the white clasts are dolomitic nor are all dolomite clasts white. The dolomite clasts are rounded to subangular and were probably moderately to well lithified when eroded. The clasts were probably formed like the micrite intraclasts, by complex syneresis or mudcracking and later erosion and transporta-
tion by waves or floods. The occurrence of dolomite clasts in micrite but never the reverse suggests that the sites of dolomitisation were topographically higher than the sites of deposition of micrite.

These clasts could not have been preferentially dolo-
mitised: not only are they slightly denser (inhibiting dolomitisation) but more significantly most intramicrites contain clasts of both dolomite and calcite. Clays may be agents in dolomitisation by acting as nuclei or by being in-
volved in ion sorbtion or clay-carbonate reactions (Kahle, 1965), but levels of insoluble residues in the Flagstaff are neither significant nor are they correlable with dolomite content. The dolomite beds that were the source for these intraclasts were formed by replacement of limestone: the same ghost features that can be seen in the very pure dolo-
mite are present in the larger clasts.
Dolomite of the Middle Member

Dolomite of the Middle Member is similarly formed by replacement of limestone. These dolomites are not identical to the pure dolomites of the Lower Member; they are not vuggy, they contain no snails, and generally they are finer grained and are sometimes completely featureless. Many comprise silt- to granule-sized intraclastic material (McCullough, 1977). The original limestone though was essentially similar to the Lower Member limestone, as is proven by the preservation in chert nodules of structures and fabrics and structures that were present in the original limestone but have been destroyed in the surrounding dolomite, as well as by features poorly preserved in some of the less atextural dolomite. An interim conclusion is that at least most of the dolomite replaced limestone, but that much has been reworked and deposited as intramicrite.

Wolfbauer and Surdam (1974) list four primary mechanisms for dolomite accumulation in Lake Gosiuie in the Green River basin: inwash of allochthonous detrital (pre-Eocene) dolomite, late diagenesis, primary lacustrine precipitation, and penecontemporaneous dolomitisation of the mudflats. Inwash is unlikely for the Flagstaff because there are no sufficient source rocks in the area (McCullough, 1977). It is also unlikely because the input of epiclastics into the Flagstaff basin was very limited and because at
least the larger dolomite intraclasts in the Lower Member are identifiable as intraformational, from snails, ostracodes, and characteristic sedimentary features. Dolomitisation by a permeating brine is impossible in the Upper and Lower Members due to the abrupt vertical changes in dolomite content in rocks that show no evidence of differential replacement. In the Middle Member there are thick section of pure dolomite that could have been dolomitised as a unit, but 1) the rare calcareous dolomites, 2) the history of dolomitisation and silicification suggested by the chert nodules (see "Formation of Chert"), and 3) the strongly evaporative conditions indicated by the mudcracks and caliche features (McCullough, 1977) argue against a pervasive late dolomitisation. However, the accumulation of several metres or tens of metres prior to a recurrence of dolomitisation is quite possible. McCullough (1977) considers that protodolomite may have precipitated directly (the third alternative), but concludes that it was not significant in the Flagstaff. Again, at least part of the dolomite replaces limestone.

Evaporative Pumping

The most likely mode of dolomite formation is replacement caused by the evaporative pumping of magnesium-rich waters through exposed calcareous muds. This is the model proposed by Hsu and Siegenthaler (1969) and accepted as the
cause of dolomitisation in Lake Gosiute by Wolfbauer and Surdam (1974) and Surdam and Wolfbauer (1975) and also by McCullough (1977) for the Middle Member of the Flagstaff. Presumably dolomitisation of carbonate mudflats occurs where the carbonate muds are relatively flat, open, and unvegetated, such as the margins of the lake that are frequently inundated. Here continual evaporation draws up the brine and concentrates the magnesium which forms dolomite. Given the wide and flat fringes on Lake Flagstaff's eastern shore, evaporative pumping is probable. Reflux of brines or infiltration of low-pH rain water (Badiozamani, 1973) may have occurred and caused precipitation, but as Hsü and Siegenthaler (1969) point out, these processes are very slow and self-inhibiting, and are thus rarely significant.

Evaporative pumping explains many characteristics of the clasts and can be applied to the Lower Member as well. It explains how so much limestone was dolomitised, how so many clasts came to be formed (extensive mudcracking on exposed carbonate flats) and it permits vertical variability in dolomite content and can dolomitise thick sections, if the climate remains arid long enough. In like manner, a brief period of evaporation to concentrate magnesium in the lake water and draw water through the exposed muds could plausibly replace single beds of limestone, explaining the thin beds of pure dolostone in the Lower Member. It also explains why dolomite clasts are common in micritic matrices
but never the reverse - if the carbonate shoals and lake-
margin flats were the site of dolomitisation then the net
transport of clasts would have been lakeward into the
micritic lacustrine sediments. Instances of encroachment
of dolomitising conditions into sites of micrite deposition
have been identified. The onset of dolomitisation is
usually preceded by the input of dolomite clasts.

In brief, the dolomite was generally formed by the
evaporative drawing of groundwater through an exposed accu-
mulation of limestone. At times dolomitisation was followed
and (or) preceded by some mudcracking. The dolomite may
later have been eroded and transported basin-ward as clasts,
to be deposited as dolomitic intraclasts. The lake marginal
mudflats were an important source of clastic dolomite
throughout the history of the lake.
FORMATION OF CHERT

Petrology

Chert is volumetrically more important than sandstone in the Flagstaff of the east-central Wasatch Plateau region, but it is much less important than limestone or dolomite. It increases upward in the section and takes different forms at different levels. It first occurs as small spheres in voids, then as void fillings. In the Middle Member it occurs as nodules, and above that as beds of silicified limestone.

Void-controlled chert of the Lower Member

The basal Flagstaff is free of chert. At Cove Mountain, the first chert occurs 95 m above the base and the remainder of the Lower Member contains only a very small amount. At first, the chert occurs as $<8 \, \text{mm}^3$ spheres and crystal aggregates, forming mainly in voids that were created by the partial dissolution of the intergranular micritic matrix in since dolomitised rocks and in rootlet moulds. These spheres rarely fill the voids. Near the top of the Lower Member, the chert has increased and diversified. It is
present as larger and more complete void fillings as both spheres and crystal tubes (term from Brewer, 1964) and as void linings (cutans). Voids include stellate packing voids (Freytet, 1973) (i.e. Brewer's orthovoids); orthovughs (Brewer) - shrinkage pores (Fischer, 1964); and shell-related voids (voids inside or below shells, or moulds); as well as rootlet moulds and vugs in dolomitised rocks. Concomitant with an increase in chert is additional precipitation in voids that are smaller and farther from the main network of voids, vughs and channels, i.e. associated with less groundwater movement. Much of this chert is a pale blue in colour.

Thus, the chert of the Lower Member is characterised by being precipitated as spheres and cutans (linings) in voids, especially ones that were probably major conduits of groundwater. The spheres suggest that they grew by nucleation as saturated groundwater moved past. Temperature was probably not a factor, but the lower pressure in the voids could have caused precipitation from a weakly supersaturated solution (see Biggs, 1957). Linings may have occurred similarly, or because of pH fluctuation. The presence of chert in voids caused by solution related to dolomitisation implies a mostly post-dolomite age.
Chert horizons of the Lower Middle Member

Chert is abundant in the Middle Member. Most of it is present as 10 to 5000 cm³ glaebules (Brewer, 1964), which were formed by replacement of the limestones and are unaffected by voids, but some of the earlier chert is possibly of a gel origin, as shown below.

The latter, unusual Lower Middle Member cherts, are discrete nodules with a smooth and round surface. They almost completely lack structure, texture, and fabric, and tend to occur along certain horizons. While some of these layers are in dolomite beds some seem to be on surfaces of erosion (or non-deposition). On one horizon, nodules occur in a thin laminated calcilutite bed that separates two much more massive dolomite beds (fig. D, plate 1). Some nodules significantly indent the dolomite above and (or) below. In two less extreme cases the calcilutite lamina tions bend around the nodules. There is also a subtype: in some layers the nodules are wedge-shaped, angular, and vertically elongated, and are rammed into the bed above (see fig. G, plate 1). These occur in beds of calcilutite.

The nodules were not formed by replacement of limestone - there are no relic structures, no included dolomite, only a few ostracodes and some possible root voids. Neither are they void fillings (as suggested by one nodule's quartz/dolomite druse- and gypsum-filled centre) for none are in voids or display concentric structure. Because of both the
vertical orientation of some and the surrounding nonclastic beds, they were probably not transported to the site. They cannot all be postdepositional, formed by water moving along bedding planes or joints because some layers of nodules occur in the middle of unbroken dolomite beds.

That they are predepositional, formed from a colloidal gel, to me seems inescapable. The wedge-shaped clasts are conclusive. On the strength of their similarity to micrite-filled mudcracks in calcilutites at the base of massive micrites (compare figures E, F, & G, plate 1), these are interpreted as mudcrack fillings, filled by the creep, inwash, or precipitation of silica gel. Their odd appearance and their indenting the beds above and below is due to compression of the surrounding calcilutite after hardening of the chert. This origin also accounts for their presence in horizons: they formed from a suddenly precipitated colloidal gel, perhaps when the lake's pH dropped sharply after a build-up of silica. The gel may have been precipitated as nodules or as a soft layer, which was later reworked into balls. Being composed of a gel, the nodules would be soft at first (thus easily rounded by rolling or by currents) and atextural and would have a nongradational boundary (see Biggs, 1957). After dewatering and hardening, the round nodules might sink or be compressed into the underlying carbonates. Subaqueous deposition of more calcilutite could easily have buried the nodule, allowing
later dewatering to compress the calcilutite and "bend" it around the nodule.

Limestone-replacing chert nodules of the Middle Member

Through the remainder of the saline phase the chert is nodular and formed by silicification of the limestone. Almost all the chert formed prior to dolomitisation. This is evident in the features preserved in the chert that are obliterated in the surrounding white dolomite. Preserved in the chert are organic material, rootlets, burrows, ostracodes, oolites, beds, and laminae (see also Pittman, 1959) (see figures A & B, plate 3 and C & D, plate 4).

The chert nodules may be brownish-black (organic rich), greenish-grey (from pyrite and reduced iron), pink to red (from hematite), or grey to light brown. The brown organic-rich cherts are commonly internally cracked (fig. D, plate 6). The cracks are filled with a clear, pale blue chert identical to that filling voids elsewhere. This chert etches preferentially when boiled in NaOH, suggesting (according to Pelto's work on similar cherts in 1956) that it is composed of smaller crystals or is less ordered or possibly contains more water. It might be opal or 2-d-cristobalite but this is not well-supported by X-ray diffraction data (see Jensen et al., 1957, for methods of determination). It is certainly cryptocrystalline and is free of the impurities in the dolomite. Some of the organic-rich
cherts are rich in pyrite. Others have extremely large amounts of ostracodes and testify to occasional ostracodal (and algal?) blooms.

These nodules form because, in an alkaline environment, nodular chert is more stable than dispersed amorphous silica. When the latter dissolves it migrates to sites of precipitation (areas of relatively low pH). The detailed replacement of limestone occurs because the same low pH favours dissolution of calcite (Pittman, 1959).

The nodules at first sight appear to be marked by a chert "front," a sharp, regularly pyrite-rich, line delimiting the dolomite and the chert. Although the boundary is in some cases quite sharp, microscopic investigation commonly shows a gradational zone about .3 to 1.5 cm wide around the chert (Figs. D & E, plate 3). Its gradational nature is shown by X-ray diffraction data, but the zone can be more readily identified by scoring a polished surface with a knife. Very poorly preserved in the zone are silicified ostracodes, shrinkage pores, and rootlets, all obliterated in the dolomite (Fig. D, plate 3).

The order of silicification can be determined in a traverse of this transitional zone. First the ostracodes and some rootlets are silicified. The ostracodes can be filled by crystalline, fibrous, and (or) clear blue chert. The rootlet moulds may first be lined with clear blue chert before being filled with crystals. Next in the process is
initial silification of the carbonate. The result is very patchy or fine and irregular and gives the boundary either a mottled or a feathered appearance (Figs. D & E, plate 3). As silification continues the remaining carbonate becomes isolated into chunks or flecks with diffuse or concave edges. These are degraded and become entirely replaced. Thus the transitional zone expands in front of the growing nodule.

Silification was very early. Not only are all the nodules predolomitisation but the many that are cut by mudcracks show the chert to be prior to mudcrack formation. The dolomite fillings of these cracks argues further against later silification (see Fig. C, plate 1). In some beds the cracks skirt the nodules but do not cut them; then, which controlled the other is uncertain. The proposed method of growth and the occasionally seen partial silification of dolomite at the edges of the nodules or in the mudcracks suggest the following history of chert formation. Amorphous silica was precipitated but being unstable in an alkaline environment, it was redistributed to form nodules, which grew in the lime muds. Next the muds were cracked, perhaps due to a lacustrine regression. Then the mudcracks were filled and dolomitisation occurred, though not necessarily in that order. Silification, which was slow relative to dolomitisation, continued briefly and then stopped. If opaline at first, the silica was soon converted to
α-quartz by the alkaline groundwater.

Silicified beds of the Upper Member

In contrast to the pure chert nodules and void fillings of other members the chert in the Upper Member replaces whole beds of limestone. The silica is dispersed through the bed, although it does not completely replace the carbonate. The beds are thin and extensive. The chert is still formed by replacement of limestone, as it shows preservation of features. It is thus prior to dolomitisation, although the occasional silicification of previously dolomitised clasts shows that silicification continued afterward as well. Ostracodes, snails, and voids are preferentially silicified. Most of the beds are mudcracked and the micritic and bioclastic debris is silicified, as are the shrinkage pores, sheets, and prisms. The intraclastic debris has been at least partly silicified (some show very diffuse and faint edges). The main part of the bed, regularly an ostracodal micrite, is at least 40% finely dispersed silica. The dolomite is also fine and dispersed, thus the rock will react equally to either hot NaOH or HCl, although neither etching is patchy or deep. For this reason, the yellowish-white weathered surfaces do not react to HCl - the carbonate has all been leached from the immediate surface. Sometime in the process of silicification, iron is mobilised in large quantities, as many chert/dolomite boundaries are marked by
pyrite and hematite rinds and the mudcrack fillings can contain up to 40% hematite that is pseudomorphous after pyrite. When in situ, the pyrite is generally clustered right inside the chert/dolomite boundary, while the pseudomorphous hematite is right outside. Both sharply decrease away from the boundary, in size and in numbers. Obviously the pyrite is forming right at the boundary, possibly because of the release and expulsion of iron during the silicification of iron-rich dolomite and calcite (see McCullough, 1977), but is only preserved in the impermeable chert-rich areas.

Preferential silicification of low-pressure areas such as fossil- and intergranular voids and replacement of entire beds suggest precipitation of silica from groundwater. Either evaporative pumping or the flushing down of low-pH meteoric waters seems likely to cause the silicification of extensive but thin units.

Geochemistry of the Chert

Source of the silica

Using criteria from Bourcart et al. (1933), analysis of Flagstaff sandstones, North Horn sandstones collected by Birsa, and remarks in the literature (e.g. Cadigan, 1972a, b; Peterson, 1976) shows that minor solution features on quartz grains are extremely widespread in this basin. Some feldspar grains also show solution features. Too little lacustrine sandstone is in the Flagstaff to be the sole
source of silica, although in other lakes this is a common source (Bourcart et al., 1933). The sandstones in the immediate area are too mature for heavy minerals to be a source for much silica. Possibly clay diagenesis in the Middle Member freed some silica. McCullough (1977) notes that the Lower and Upper Members contain mixed-layer clays while the Middle Member clays are illitic. Montmorillonite can change into illite (Weaver, 1958) acquiring $K_xAl_{x+y}^-$ and releasing $(Mg, Ca)_y$ and $Si_x^-$ and a minor amount of exchangeable cations. (Tome, 1962; Berner, 1971). Generally the silica seems to have come from the dissolution of the detrital quartz, chert, and feldspar in both the clastic-rich Indianola, North Horn, and Colton Formation within the basin, and the Mesozoic beds in the uplifts nearby.

Transportation into the lake

Silica probably entered the lake in streams or groundwater in solution or as colloids, for epiclastics were rarely transported across the floodplain. From experimental evidence, rivers can carry as much as 100 to 140 ppm silica, mostly as colloids (Krauskopf, 1959). Dissolved silica was probably minimal in the Flagstaff streams because the pH was probably below 9 ($pH = 8.4$ when water is in equilibrium with air and calcite [Garrels, 1960, p. 52] or between 7 and 8 if typical of streams in arid regions [Krauskopf, 1967, p. 35]). Below $pH = 9$, silica forms $H_4SiO_4$ which is relatively
insoluble and not pH controlled. In this range $K_{sp}$ of amorphous silica is 120 ppm and $K_{sp}$ of $\alpha$-quartz is 16 ppm (Berner, 1971). The normal range in rivers is 3 to 120 ppm (including colloids?) (Berner, 1971) or 10 to 60 ppm (Krauskopf, 1967). Livingstone (1963) gives an average value of 21 ppm. Regardless, $\alpha$-quartz is always super-saturated, but does not precipitate, due partly to kinetics. With time colloids will form, and will produce a gel if sufficiently concentrated (Krauskopf, 1967).

Groundwater in limestone, when closed to the atmosphere, has a pH of 9.9 (Graf, 1960) and can move much more silica into the lake than streams. At this level, silicic acid ionises to $H_3SiO_4^-$ and $H_2SiO_4^{2-}$ (Krauskopf, 1959), and solubility is greatly increased and, furthermore, pH dependent. Jones et al. (1967) indicate that $\alpha H_3SiO_4^-$ at pH = 9.9 will be 450 ppm.

Precipitation of silica

Once in the lake silica concentration will be increased by either increased solubility due to increased pH or evaporation of the lake, until precipitation is induced. At surface temperatures, dissolved silica may be brought out of solution by pH fluctuations, organisms, adsorption, reaction with cations to form silicates, by a slow equilibration with crystalline silica, and by a decrease in pressure (Krauskopf, 1959; Jones et al., 1967). Colloidal silica
is precipitated (as a gel) by evaporation, cooling, or addition of electrolytes. The main concern is the precipitation of dissolved silica, because the formation of a gel from colloids, thus depositing silica, is merely a question of time, although the process can be speeded up. Anything that speeds precipitation promotes formation of a gel rather than individual colloids (Krauskopf, 1959).

Uptake of silica by organisms is unlikely to have been important in Lake Flagstaff as no fossils of silica-requiring organisms have been found. If any existed, the most likely is Equisetum, the horsetail. Fresh water diatoms are not known until the Miocene (Lohman, 1964) and because they completely control silica cycles in modern lakes (Wetzel, 1975, pp. 278-286; for an example see Hecky and Kilham, 1973), silica cycles in lakes today may not be analogous to Lake Flagstaff.

Complexing and adsorption are of minor importance in the silica cycle (Krauskopf, 1959). silica may be taken up in the diagenesis of clays (Weaver, 1958) and, if cations are abundant, it may form silicates (Eugster and Chou, 1973), particularly sepiolite (Hardie and Eugster, 1970) and magadiite (Eugster, 1969). Neither is present in the Flagstaff.

The role of pressure has already been mentioned: the relatively low pressure in voids prompts precipitation from groundwater.
By far the most important silica control would be pH. As it rises above 9, silica is concentrated in solution but when it falls below, silica is precipitated. pH fluctuations can occur in many ways. pH is usually raised by evaporative concentration of bicarbonate-rich waters (Eugster & Chou, 1973) or by CO₂ removal during photosynthesis (Hutchinson, 1957). The former can raise pH above 12, especially when Na is present (Hutchinson, 1957) and in the latter case, the pH may commonly rise to 10.2 (Peterson & Borch, 1965). This often occurs seasonally or during algal blooms but may also happen on a daily basis when the waters are alkaline to begin with. Heating the lake also raises pH by driving off CO₂. On the other hand, increased CO₂ will lower the pH. The major source of CO₂ is organic decomposition in alkaline lakes (Peterson & Borch, 1965), but it is also produced during "dark" photosynthesis and by animals' respiration, and is taken up from the atmosphere as the lake cools. Chemical and thermal stratification of the lake may be very important. In the summer, CO₂ concentration is inversely clinograde (Hutchinson, 1957) – there is more in the cold hypolimnion than in the warm, productive epilimnion. As the epilimnion warms, the CO₂ becomes trapped below it. If the thermocline is also a chemocline, the boundary will be very abrupt and pH-dependent reactions can occur. Similarly, there is frequently a boundary at or just above the sediment-water interface between the acid to neutral waters that
are rich in decomposition-derived CO₂ and the overlying alkaline lake water where silica can be precipitated (Peterson & Borch, 1965).

pH may also be lowered by freshwater flooding. When approximately neutral freshwater flows into a saline-alkaline lake it may float on top and not mix, causing chemical stratification. Precipitation of silica can occur at the chemocline as dilution and lowering of pH take place (Eugster, 1969). The necessary pH fluctuations can also occur when meteoric water, in equilibrium with calcite and the air, mixes with calcareous groundwater not in contact with the air (Graf, 1960; see also Badiozamani, 1973). The former theoretically has a pH of 8.4, the latter, 9.9. Mixing of the two during infiltration of rainwater or lake water may supersaturate the silica and cause precipitation. High levels of supersaturation may be reached when the water is not in contact with silicate muds (Jones et al., 1967) and low in sodium (precluding magadiite formation); this could have been the case in Lake Flagstaff.
CHEMISTRY OF THE LAKE WATER

Chemical Composition of the Water and Trends over Time

The waters of Lake Flagstaff were chemically simple. This simplicity is due to the limited diversity of the surrounding rocks and is attested to by the incomplex mineralogy of the lake sediments. All the possible source rocks are sedimentary; conglomerate, sandstone, limestone, dolostone, and shale, with minor gypsum and halite. More importantly, the post-Jurassic beds are either relatively pure precipitates or are multicycle sediments, being ultimately derived from the lower Paleozoic, mostly marine limestone, dolostone, shale, and orthoquartzite (Hintze, 1973). Therefore only a limited number of elements entered the lake.

The lake deposits include limestone, dolostone, chert, gypsum, and some terrigenous shale and sandstone, all of whose important minerals are calcite, dolomite, quartz, and gypsum. Lesser constituents are various clays, hydrocarbons, pyrite, hematite, and some feldspar (in the sandstones). No salts were found. Consideration of the
minerals in the sediments and the source rocks therefore suggests that the lake was dominated by Ca, Mg, CO$_3^{2-}$, occasionally appreciable SO$_4^{2-}$, and minor Fe.

Before attempting any chemical approximations of Lake Flagstaff, it is convenient to establish the trends of the changes in composition during the lake's existence. Recall that the lowermost part of the Lower Member at Cove Mountain is almost pure limestone. After its first occurrence, dolomite increases gradually (but irregularly) up section. The same is true of chert, although it does not appear much before the top of the member. All the carbonate beds in the Middle Member are dolomite, and the member also contains many large chert nodules, and some gypsum in its upper and lower units. As demonstrated earlier, dolomitisation is frequently pre- or early post-burial and silicification is predolomitisation for the most part, thus the chemistry of a bed is usually indicative of the chemistry of the lake water. Therefore, from the gross lithology it can be concluded that 1) calcium and carbonate were always abundant in the lake, 2) levels of magnesium, at first low, gradually increased through the end of the early lake phase (despite large fluctuations), remained very high through the middle phase (with two minor decreases), and dropped only slightly during the late phase, 3) silica followed the same pattern, except that it further increased during the late phase, and 4) sulphate was low during the early and late Flagstaff but
reached high concentrations during the middle Flagstaff, especially during precipitation of gypsum in the upper and lower units.

Lake Flagstaff lends itself to chemical modelling because it has so few components. In an open system such as Lake Flagstaff calcium, total carbonate, and pH are interdependent (see Garrels, 1960, pp. 43-60, and Krauskopf, 1967, pp. 50-88). Also, because the ions present form mainly calcite, dolomite, and gypsum, and because any two of these three minerals will be competing for a common ion, formulae can profitably be matched with one another to determine likely ionic concentrations. Furthermore, pH controls silica.

Maximal Ionic Concentrations During the Earliest or Calcareous Lake Phase

Prior to the appearance of dolomite, early Lake Flagstaff must have been a very simple hardwater lake. Carbonate equilibria in a system open to atmospheric CO₂ and in contact with unlimited CaCO₃ create a pH = 8.4 (Garrels, 1960, pp. 54-56). Use of Garrels' method 4 (pp. 54-56) shows that this would have caused concentrations of \([\text{HCO}_3^-]\) = 61 ppm and of \([\text{Ca}^{2+}]\) = 12 ppm. (See figure 11a for variations in \([\text{Ca}^{2+}]\) with pH.) The absence of dolomite suggests that \([\text{Mg}^{2+}]\) was probably not much higher than \([\text{Ca}^{2+}]\). Müller et al. (1972) show that the precipitation
and diagenesis of carbonate minerals in lakes are controlled by the Mg/Ca ratio. In a lakes characterised by precipitation of calcite only and by muds of pure calcite, the Mg/Ca ratio is between 0 and 2. Assuming the latter, \([\text{Mg}^{2+}]\) in the Flagstaff was 24 ppm, giving a total theoretical hardness of 97 ppm. Minor iron and sulphur were probably present as iron oxides and \(\text{H}_2\text{S}\), and in organics.

Approximate chemical analogues to early Lake Flagstaff are lakes such as Preble-Green Lake and Tully-Green Lake, small springfed marl lakes in New York (Cucci, 1974) and (when magnesium begins to increase) Lake Balaton, in Hungary (Müller et al., 1972). The pH in Tully-Green Lake remains around 8.3, and \(\text{Ca}^{2+}\) is approximately 67 ppm. One set of values from Preble-Green Lake shows the pH = 8.25 (low), \([\text{HCO}_3^-]\) = 178.5 ppm, and \([\text{Ca}^{2+}]\) = 57 ppm. The pH there varies between 8.1 and 8.52, \([\text{HCO}_3^-]\) from 177 to 215 ppm, and \([\text{Ca}^{2+}]\) to 100 ppm. In comparison, Lake Balaton, a 600 km\(^2\) lake with a maximum depth of 12 m but an average depth shallow enough that fishing boats frequently hit bottom when tossed by storm waves (Salanki and Ponyi, 1975), has a pH of 8.6 and a hardness of 500 ppm, comprising mostly \(\text{HCO}_3^-\), \(\text{Mg}^{2+}\), and \(\text{Ca}^{2+}\), and also some \(\text{Na}^+\) and \(\text{SO}_4^{2-}\). Its Mg/Ca ratio annually fluctuates between 0.8 and 2.4, depending on evaporation and input. So, the earliest Lake Flagstaff probably contained \(\text{HCO}_3^-\), \(\text{Mg}^{2+}\), and \(\text{Ca}^{2+}\), had a pH of 8.3 to 8.6, and a hardness of up to, perhaps, 500 ppm.
Probable Concentrations During the Late
Early and Middle or Dolomitie Phase

Modelling the dolomite-containing lake is best done by applying the work by Müller et al. (1972) and others on the correlation between Mg/Ca ratios and the diagenetic production of dolomite. Frequent precursors are high-Mg calcite (Schmidt, 1965) and protodolomite (Müller, 1970). Estimates of the necessary conditions for dolomite have ranged from Mg/Ca between 4.7 and 22 (Wolfbauer and Surdam, 1976) to Mg/Ca > 10 and dissolved solids > 10,000 ppm (Jones, 1965). Alderman (1965) found that high-Mg calcite forms when pH is about 9 and Mg/Ca > 6, while dolomite assemblages occur when pH > 9 and Mg/Ca = 7 to 10. In a series of studies covering many lakes Müller and his coworkers concluded that for dolomite to form the interstitial liquids and (or) the lake water must have an Mg/Ca ratio above 7 and high-Mg calcite must be present (Müller, 1973, 1975; Müller and Förstner, 1975; Müller et al., 1972).

Folk (1974) and Folk and Land (1975) have a slightly different but less comprehensive interpretation. They suggest that as salinity (Na⁺ and SO₄²⁻ content) decreases, dolomitisation occurs at lower ratios: at salinities common in lakes and streams, i.e. less than 5% of the ocean, dolomitisation will occur at Mg/Ca = 1. Under these conditions the dolomite precursor is typically low-Mg calcite.
They also predict crystal shape on the basis of salinity and Mg/Ca ratio. SEM examination of the Flagstaff micrites show much of the dolomite and micrite to be present in small, very fresh, clean, equant blocks, or unit rhombs, that are between .5 and 3 μm in length. Microspar like this is distinctive of "relatively low" salinity and low Mg/Ca conditions (generally less than 3500 ppm, or $10^2$-$10^3$ ppm, and Mg/Ca less than 3:1). Dolomite with these characteristics is also found in slightly more saline, Mg-rich conditions.

The $\text{HCO}_3^-$ and $\text{SO}_4^{2-}$ contents are also best expressed as ratios. If $\sum \text{CO}_3^{2-}/(\text{Ca}^{2+}+\text{Mg}^{2+}) \gg 1$ then on evaporative concentration Ca and Mg would be depleted by calcite and dolomite formation and the next precipitate would be another metal carbonate. If less than 1, the next precipitate would be another Ca or Mg salt - in the Flagstaff it was $\text{CaSO}_4 \cdot \text{H}_2\text{O}$. Sulphate is presumably present and increasing by the end of the early phase of the Flagstaff but due to the high solubility of gypsum it may leave no mineralogic evidence of its presence. In fact, because the $\text{CO}_3^{2-}$ and $\text{SO}_4^{2-}$ anions were competing for $\text{Ca}^{2+}$, this comparison

$$\frac{K_{\text{CaSO}_4}}{K_{\text{CaCO}_3}} = \frac{10^{-4.6}}{10^{-8.3}} = 10^{+3.7}$$

show that $[\text{SO}_4^{2-}]$ may be up to 5000 times $[\sum \text{CO}_3^{2-}]$ before gypsum is precipitates in preference to calcite.
Chert is present in the uppermost Lower Member and above: its presence signifies that SiO₂ is increasing in the lake water, which in turn implies a higher but fluctuating pH (as described in the section on the formation of chert). The mudcrack-filling cherts in particular evince periods when the pH exceeded 9, allowing the build up of silica. As the pH rises above 10.3, CO₃⁻ replaces HCO₃⁻ as the dominant species of ΣCO₃⁻ (Krauskopf, 1967, p. 50).

A possible analogue is the Neusiedler See, a large (290 km²) but very shallow lake in the Austrian - Hungarian border region and which was studied by Schroll and Wieden (as quoted in Müller, 1975). Although the salinity is "only" 1,500 ppm, high-Mg calcite and protodolomite are forming because of high Mg/Ca values. The Neusiedler See has a maximum depth of 1.5 m (Müller et al., 1975) and a pH of 9; it also has fairly large amounts of Na⁺ and Cl⁻.

In conclusion, the dolomite-producing phase of the lake was probably characterised by occasionally high pH, salinity on the order of 1,500 ppm, a high Mg/Ca ratio, possibly large amounts of SO₄²⁻, and ΣCO₃⁻ less than Mg+Ca, although possibly not all at once.

Minimal Concentrations During the Gypsum Precipitation

Indirect Control of Sulphate by pH

The last case, the most complex but also the most informative, is the composition of the water that
precipitated the gypsum. As previously mentioned, gypsum is precipitated in preference to calcite by increasing the $\text{SO}_4^{2-}/\text{CO}_3^{2-}$ ratio, by increasing the sulphate or decreasing carbonate. The second is unlikely because the lake was in contact with the atmosphere. When a lake is an open system, $\text{CO}_2$ will be taken from the atmosphere to form $\text{H}_2\text{CO}_3$ until $[\text{H}_2\text{CO}_3]$ reaches $10^{-5.0}$. $\text{H}_2\text{CO}_3$ will dissociate to $\text{HCO}_3^-$ and then to $\text{CO}_3^{2-}$ to an extent that is governed by pH. ($\text{CO}_3^{2-}$ may then react with $\text{Ca}^{2+}$ to form calcite.) As $\text{H}_2\text{CO}_3$ dissociates, more $\text{CO}_2$ will be taken up to maintain $[\text{H}_2\text{CO}_3]$ at $10^{-5}$. Therefore $\text{CO}_3^{2-}$ will always be saturated and $[\text{CO}_3^{2-}]$ will be governed by pH. As pH rises, $[\text{CO}_3^{2-}]$ will increase, precipitating $\text{CaCO}_3$ and lowering $[\text{Ca}^{2+}]$. Therefore $\text{Ca}^{2+}$ will be indirectly governed by pH. Given the right pH conditions, calcite could be continuously precipitated until $\text{Ca}^{2+}$ is depleted. This makes gypsum precipitation impossible through $\text{CO}_3^{2-}$ depletion by excess $\text{Ca}^{2+}$, thus requiring $\text{CaSO}_4$ to be precipitated below $\text{CaCO}_3$ saturation. This is easiest achieved by concentrating $\text{SO}_4^{2-}$. The concentration required will be a function of $[\text{Ca}^{2+}]$, and thus indirectly of pH, which is convenient as it might allow supersaturation and precipitation by reasonable pH fluctuation instead of physical concentration. Also the Mg/Ca ratio must not be high enough to favour $\text{MgSO}_4 \cdot \text{H}_2\text{O}$ (kieserite). Thus pH fluctuation, evaporative concentration, and dolomitisation are likely agents in producing a gypsum-precipitating,
Ca - SO$_4$ brine.

Using formulae from Garrels (1960, pp. 54-56),
log[Ca$^{2+}$] may be plotted as a function of pH (figure 13a).
This produces the relationship

$$\log \text{Ca}^{2+} = -2 \text{pH} + 13.8$$

(Note: the ion activity product, here set equal to 0.10
which is an order of magnitude above the IAP for most fresh
water -Berner, 1971-, is not a major factor. For example,
at pH = 9, Ca=2.5 ppm at IAP=0.10, while at IAP=0.001
Ca=1.2 ppm.) Equilibrium concentrations of Ca$^{2+}$ and SO$_4^{2-}$
may be plotted using the following equation (see figure 13c)

$$K_{\text{CaSO}_4} = [\text{Ca}^{2+}][\text{SO}_4^{2-}] = 10^{-4.6}$$
or,

$$\log \text{SO}_4^{2-} = -\log \text{Ca}^{2+} - 4.6$$

Because maximum [SO$_4^{2-}$] is determined by the reaction
CaSO$_4$ ⇌ Ca$^{2+}$ + SO$_4^{2-}$ and because Ca is controlled by pH, it
follows that pH, through keeping Ca at equilibrium values,
also sets the maximum or saturation level of SO$_4^{2-}$. Sub-
stituting for Ca$^{2+}$ in the gypsum equation yields

$$\log \text{SO}_4^{2-} = 2 \text{pH} - 18.4$$

This which shows the dependence of maximum SO$_4^{2-}$ levels on
pH (see fig. 11b).

Implications for precipitation of Gypsum

Because pH controls calcium which controls sulphate,
one can draw a horizontal line on figure 13c that repre-
sents the concentration of calcium at any given pH. Note
Equilibrium concentrations of Ca from CaCO$_3$ vs. pH

\[ \log \text{Ca} = -2 \text{pH} + 13.8 \]

Equilibrium \( \text{M}_{\text{SO}_4}^- \) from CaSO$_4$ at varying pH

<table>
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<th>( \text{SO}_4^{2-} ) (ppm)</th>
<th>pH</th>
</tr>
</thead>
<tbody>
<tr>
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<td>6.71</td>
</tr>
<tr>
<td>10</td>
<td>7.20</td>
</tr>
<tr>
<td>100</td>
<td>7.71</td>
</tr>
<tr>
<td>500</td>
<td>8.06</td>
</tr>
<tr>
<td>1000</td>
<td>8.20</td>
</tr>
<tr>
<td>10000</td>
<td>8.71</td>
</tr>
</tbody>
</table>

Figure 13

Equilibrium concentrations of Ca$^{2+}$ and SO$_4^{2-}$ and calcite and gypsum fields at a given pH
that the lines intersect above \( \log \text{SO}_4^{2-} = 2 \text{pH} - 18.4 \) and that the resulting quadrants have been labelled. At the given pH, it can be seen that both CaCO\(_3\) and CaSO\(_4\) are supersaturated in quadrant I, although at lower SO\(_4^{2-}\) concentrations (quadrant II) only CaCO\(_3\) will be precipitated. In quadrant III (low Ca\(^{2+}\) and SO\(_4^{2-}\)) nothing will be precipitated, but in quadrant IV (high SO\(_4^{2-}\) concentrations) only CaSO\(_4\) is precipitated.

But this graph may be simplified and made more useful. Consider conditions in which both CaCO\(_3\) and CaSO\(_4\) are undersaturated, i.e. in quadrant IV. CaSO\(_4\) may not always be precipitated by increasing [SO\(_4^{2-}\)] but also by increasing [Ca\(^{2+}\)] if the original \( \log \text{SO}_4^{2-} \) is greater than 2 pH - 18.4. Therefore CaCO\(_3\) will only be precipitated by increasing [Ca\(^{2+}\)] when \( \log \text{SO}_4^{2-} \) is less than 2 pH - 18.4. This separates the calcite and gypsum fields and eliminates the boundaries of quadrant I because those concentrations could not be attained unless the rate of concentration exceeded that of precipitation. (Quadrant I concentrations might also be attained if ion complexing occurred or if pH rose drastically while the waters were precipitating gypsum.)

Expressed in reverse, when Ca\(^{2+}\) is at saturation levels, if pH is less than or equal to \( 1/2(\log \text{SO}_4 + 18.4) \), then gypsum will precipitate. This is shown in figure 13b. Note that low-pH but Ca-rich waters will precipitate gypsum at very low concentrations of SO\(_4^{2-}\). Because the Flagstaff
was generally quite alkaline, gypsum formation indicates a rather high $\text{SO}_4^{2-}$ - salinity.

Implications for the salinity

How high could the salinity have been? (Note that $\text{Na}^+$ and $\text{Cl}^-$ have not been accounted for.) Once the Ca is removed from the system, pH and salinity can reach very high levels (over 10 and $10^3$ ppm respectively) (Livingstone, 1963; Eugster, 1970). Without $\text{Ca}^{2+}$, $\text{CO}_3^{2-}$ and $\text{SO}_4^{2-}$ can increase and form complexes with other bases (usually $\text{Mg}^{2+}$ and $\text{Na}^+$) (Livingstone, 1963). The next step is the precipitation of trona at 290,000 ppm (Bradley and Eugster, 1969) or halite (if the brine salinity is over 150,000 ppm $\text{Cl}^-$ or 10,000 ppm $\text{Mg}^{2+}$ - Kinsman, 1976). However, concentrations this high are unlikely in Lake Flagstaff, because nothing suggests that $\text{Ca}^{2+}$ was ever depleted.

Salinity may still be quite high, even with $\text{Ca}^{2+}$. Livingstone (1963) lists Bourcart's values from Lac Ritom, in France, which may well be roughly equivalent to the saline Flagstaff. The total salinity is 2432 ppm, of which $\text{SO}_4^{2-}=1658$ ppm, $\text{HCO}_3^-=109$, $\text{Ca}^{2+}=525$, $\text{Mg}^{2+}=118$, $\text{SiO}_2=10$, $\text{Fe}=8.3$, and $\text{Na}^+ + \text{K}^+ < 5$. $\text{Mn}^{2+}$, $\text{Al}^{3+}$, $\text{PO}_4^{3-}$, $\text{NO}_x$, and $\text{Cl}^-$ are absent. The ion activity is 0.143 and at this salinity ion association or complexing may be important, particularly affecting $\text{CO}_3^{2-}$, $\text{SO}_4^{2-}$, $\text{HCO}_3^-$, $\text{Ca}^{2+}$, and $\text{Mg}^{2+}$ (in probable descending order) (Garrels and Thompson, 1962).
In summation, if the lake had a pH > 8.4 (the pH of water in contact with the atmosphere and CaCO\textsubscript{3} - Krauskopf, 1967) then from diagramme 11a [Ca\textsuperscript{2+}] must have been greater than 10\textsuperscript{-2.8} and for gypsum precipitation [SO\textsubscript{4}\textsuperscript{2-}] must have been above 10\textsuperscript{-1.8}. Because \(\sum CO\textsubscript{3}\textsuperscript{2-}/(Mg+Ca) < 1\) (shown earlier) and Mg/Ca=3:1 (from Folk, 1975), then middle Lake Flagstaff had a salinity of at least 2,000 ppm: 1500 ppm SO\textsubscript{4}\textsuperscript{2-}, 65 ppm Ca\textsuperscript{2+}, 195 ppm Mg\textsuperscript{2+}, and 250 ppm CO\textsubscript{3}\textsuperscript{2-} (almost entirely HCO\textsubscript{3}\textsuperscript{-}). At higher pH values, SO\textsubscript{4}\textsuperscript{2-} increases but all else decrease (e.g. at pH = 9, SO\textsubscript{4}\textsuperscript{2-} = 38,200 ppm, while Ca\textsuperscript{2+} = 2.5 ppm). Clearly a minor drop in pH after the evaporative concentration of SO\textsubscript{4}\textsuperscript{2-} can cause major gypsum precipitation.
PALEOLIMNOLOGY

Before discussing individual topics, it might be well to briefly describe the early Lake Flagstaff. Lake Flagstaff was created by ponding between the old Sevier orogenic belt in the west and early Tertiary uplifts in the east. Gradually alluvial fans, alluvial plains, streams, and ponds became inundated until a very large NE-SW oriented lake was formed. The lake was in a very flat basin and was shallow and weedy throughout. Lake level fluctuated constantly in response to rainfall and evaporation, causing exposure or inundation of very large areas. The occasionally inundated carbonate mudflats are nonclastic, were probably barren to weedy, and were crossed by a few streams. Exposure of the flats caused extensive mudcracking; prolonged exposure allowed soil forming processes and calichification to create some rather distinctive features such as polyphased nodules and recrystallisation fabrics. Flash floods and wave action during reexpansion of the lake created erosion surfaces, removed partly lithified micrite fragments, redeposited them as intraclasts, and filled mudcracks with detritus. Frequent erosion and shifting of the sediment kept the lake
floor flat.

Wind was probably a very important agent. Because the lake basin was flat and the lake extensive and shallow, the fetch was very long and storm waves could probably disturb the entire bottom. The largest possible waves in the Flagstaff can be determined by height = 0.105 \sqrt{\text{fetch}} (Wetzel, 1975, p. 95), assuming wavelength to be less than water depth. Given a north-south fetch of 120 km (a conservative estimate) the maximum height is 3.6 m. (The maximum east-west height with a 50 km fetch is .7 m.) Because of the increased fetch and the flooding of wave-baffling vegetation, waves must have been significantly larger during periods of active lake transgression and thus transgression should be associated with large-scale reworking of sediment, which can perhaps be seen in the reworking of clasts derived from mud-cracks. During low lake levels, friction on the lake floor and baffling by vegetation probably greatly prevented the build-up of waves.

The wind can also cause a transgression; in fact, barometric pressure alone causes a water level difference of 1 cm/millibar (Muir-Wood, 1969), which, assuming a 10 mb differential across Lake Flagstaff and a 1°/mile slope for the mudflats, could have caused a 5.5 m transgression. The wind is more efficient, as shown by equations from Muir-Wood. After the wind has blown long enough to establish steady conditions \( \tau + \tau_b = g \frac{1}{2} \rho_w h \).
where $\tau = \text{water surface shear stress}$, $\tau_b = \text{shear stress between the returning current, and the lake floor}$, $g = \text{gravity}$, $S = \text{surface gradient or wind slope}$, $\rho_w = \text{density of water}$, and $h = \text{mean water depth}$.

$\tau_b$ is usually small with respect to $\tau$, and as

$$\tau = C \rho_a U^2$$

(where $C = \text{coefficient of friction}$, $\rho_a = \text{density of air}$, and $U = \text{wind speed at 10 m}$) then

$$S = \frac{C \rho_a U^2}{\rho_w g h}$$

$C = 2.5 - 3 \times 10^{-3}$, applied to $U_{10m}$ for high wind speeds. $\rho_a = 1.25 \times 10^{-3}$ g/cm$^3$ (values from Muir-Wood, 1969) and $\rho_w = 1.0$ (density of water at 25°C is 0.9983 but with a 2 ppt salinity is 1.00169 - Wetzel, 1975, p. 10).

Assuming the lake averages 1 m deep, a wind of 50 km/hr creates a windslope of 6 cm/km which could create a 300 cm high pileup of water across the lake. Assuming a 1°/mile slope, this causes a 0.2 km transgression down wind. A wind of 100 km/hr blowing the length of the lake causes a denivellation of 3000 cm at one end and a transgression across 1.7 km of lake plain at the other. If the average water depth is 10 m (which is highly unlikely considering the vast flat eastern shoreline) then the above values would be reduced by an order of magnitude. In very flat
playas, wind has been known to blow bodies of water across the basin floor (Motts, 1970).

Wind would also greatly effect evaporation and water temperature change, making both more rapid as wind speed increases, which leads into other paleoclimatic considerations.

It would be nice to be able to use Langbein (1963) and Schumm (1965, 1968) to more accurately show changes in evaporation, run-off, the volume and area of the lake, sediment influx, temperature, and rainfall, etc., and how they affect the lake and each other, but too much is uncertain and an attempt rapidly becomes an exercise in fantasy. These parameters are much better known for the Green River basin because of studies such as McGinitie (1969). The best information on the Paleocene climate can be summarised briefly. The absence of halite (as discussed previously) may be because the relative humidity never fell below 76% (as calculated for sea water by Kinsman, 1976). Similarly, the presence of gypsum indicates that, at least during gypsum precipitation, it had to be below 93%. The crocodile found by Davis (McCullough, personal communication) indicates that the Upper Flagstaff remained above 10°C in the coldest month (Berg, 1964). Brown (1962) analysed the floral composition and concluded that the Paleocene climate in the Rocky Mountain area was latitudinally zoned (breadfruit and cinnamon only grew south of central
Wyoming whereas ginkgos, birches, etc., did not grow south of Wyoming), and overall was warm temperate with medium precipitation distributed throughout the year. The best paleoclimatic study was done by Tidwell et al. (1976), who analysed leaf physiognomy, which is the most accurate basis for floral analysis. They agree with Brown that the early Paleocene climate was zoned (from subtropical to warm temperate) and add that the Fort Union flora at Baggs, Wyoming indicates a warm temperate to subtropical, seasonably warm to moist climate. Thus the Flagstaff climate was probably warm and could have been dry to humid, depending in part on the rainshadow effect of mountains along the Sevier orogenic belt.

The salinity and reduced volume of the early Eocene saline phase argues for a temporary increase in temperature or decrease in rainfall. This peak can be seen in Figures 3 & 5 in Dorf (1969). If these curves are correct, the late Flagstaff was a little warmer than the early Flagstaff.

The lake was very productive and full of vegetation but was weedy rather than swampy (as shown by the rootlets and the lack of extensive organic deposits and pyrite). Aquatic macrophytes were probably the most important biotic constituent of the lake, as they were the pioneers, recolonising recently inundated floodplain after a regression/reexpansion cycle. They provided food, habitat, and hiding places for the organisms that followed (McLachlan, 1975).
They also allowed trophic diversity over what would otherwise be a uniform, featureless lakefloor and provided temperature buffers for fish (McCarragher, 1970). Mollusks and ostracodes were abundant, fish and reptiles probably were too (La Rocque, 1960).

The lake floor was probably soft and "muddy" lime most marls. The water was probably very turbid. This is supported by snail evidence (La Rocque, 1960) and is expected of shallow, calcite-precipitating lakes. Indeed, shallow lakes frequently have a Secchi disk transparency as low as 6, 10 or 15 cm (McLachlan, 1975; Walker, 1973; Melack & Kilham, 1974, respectively). The turbidity made Lake Flagstaff an "optically shallow" lake, because of the abiological absorption of light by stirred up mud (Salanki & Ponyi, 1975). This has the effect of compressing the floral zones, which are caused by light availability.

Lake Flagstaff almost certainly had very irregular temperatures, which probably affected species composition. Shallow lakes change temperature more rapidly than deep ones because of more rapid mixing and greater relative sediment/water and air/water interface areas. Salinity also causes erratic heating (Walker, 1973). A high salinity lowers evaporation, therefore the heat normally lost by evaporation has to be lost by other means. In saline lakes, the sediments can be a very important heat sink. This was probably not significant in the early lake.
The lake water was very hard, containing much Ca$^{2+}$ and CO$_3^{2-}$ and minor but gradually increasing amounts of Mg$^{2+}$, SO$_4^{2-}$, and SiO$_2$. The early Lake Flagstaff almost certainly contained less than 500 ppm ions. The Mg/Ca ration was very low at first, perhaps less than 1. Neither dolomite nor chert was formed then. pH probably started at about 8.4 and also gradually increased. Upper Lower Member dolomite and chert suggest that the pH and Mg$^{2+}$ and SiO$_2$ concentrations had risen considerably by then, and, with SO$_4^{2-}$, continued to do so into the Middle Member. The calcareous nature of Lake Flagstaff undoubtedly gave it a very green colour (Reeves, 1968).

At first Lake Flagstaff was probably a normal hardwater lake, distinguished mainly by its "playa lake" nature. "Playa lakes" are usually associated with extreme desert conditions and thick deposits of evaporite salts, however they can occur in much more humid condition and are really only characterised by their relative flatness and shallowness and fluctuating water levels (Reeves, 1972). Much of a playa lake may be permanently lacustrine. At first, Lake Flagstaff may even have had an outlet. The ecology of the early lake was probably much like that of the small playa lakes of Nebraska described by McCarraher (1970) and that of the less saline lakes of the Kirghiz Steppe (Mozley, 1937). Later the situation was quite different. The high pH and high concentrations of Mg$^{2+}$ and SO$_4^{2-}$ undoubtedly had a
restrictive effect on the biota (Mozley, 1937) explaining the absence of snails. The evaporites show that the climate was hotter and (or) more arid, and the lake probably experienced greater absolute fluctuations in temperature, pH, salinity, and volume, which probably also affected the biota.

The saline phase probably produced less biomass overall, conditions being severe and the lake smaller, although saline lakes can be highly productive (Melack & Kilham, 1974). From fossil evidence the biota of middle Lake Flagstaff was certainly less diverse. Typically, saline lakes follow a definite cyclical succession but have only one or maybe two species of phytoplankton and of rotifers, etc., a season (Walker, 1973). The saline lake was subject to massive ostracodal blooms, which may have occurred after an expansion of the lake.

As in most lakes the most critical nutrient was probably phosphorus (Vollenweider, 1968). This is a very involved topic, but it and iron were probably in sufficient supply to cause a significant amount of nutrient (phosphate) regeneration, which can greatly increase a lake's productivity (see Einsele, 1941, and Mortimer, 1961). Nutrient regeneration is dependent on Eh reactions across an anoxic sediment—oxic water interface involving iron phosphate and oxides, and colloid and sorption reactions (see Golterman, 1967) but the Flagstaff lake floor might
well have been almost continuously oxic, as is probable in very shallow lakes due to the ease with which they are mixed. An attempt to use the criteria set up by Degens and Stoffers (1976) to determine whether the carbonates were deposited in oxic or anoxic waters suggested that the water was oxic (i.e. at magnifications of 5 to 50 x 10^3 no solution features were seen on the carbonate rhombs; see Figure A, plate 5), however the effects of diagenesis are not allowed for in their work and thus bring the results into question. Solution, stylolite formation, and recrystallisation certainly occurred in many beds. If the bottom was usually oxic, nutrient regeneration would be impeded, lowering productivity. Another factor that could have lowered productivity during the saline phase is the acceleration of the change of nitrate to ammonia at a high pH, thus causing nitrogen to be lost to the atmosphere and increasing the toxicity of the water. Despite this, the lake was probably very productive. Fossil evidence is strong on this point (La Rocque, 1960). Also, Hutchinson (1957) notes that, in large lakes, productivity is inversely proportional to the mean depth because of the increase in area of the littoral zone in shallow lakes. Vollenweider (1968) shows another reason why productivity should be high - the loading of phosphorus and nitrogen required to eutropify a shallow lake is extremely low (2 g/m^2/yr of N and 1 of P for a 5 m deep lake).
Evaporation to dryness, the precipitation of the dolomite and gypsum and their burial probably "freshened" the lake (see Gilbert, 1890) because except for high silica concentrations and possible existence as a group of smaller lakes, the late Lake Flagstaff was probably similar to the lake at the end of the early phase and the early middle phase.
CONCLUSION

Throughout its existence, Lake Flagstaff was a large and very shallow calcareous alkaline lake surrounded by wide and flat mudflats. Early Lake Flagstaff was very large, fresh, and calcareous. It was very productive, as evidenced by the abundance of vegetation, but it mainly precipitated calcite physicochemically. Some of the higher beds were dolomitised in situ, but most of the dolomite is detrital, probably forming by evaporative draw on the mudflats (as in the Middle Member) and being reworked into the lake by sheetwash floods and waves. Chert spheres occur in some voids. Because of a more arid climate, Middle Lake Flagstaff was restricted and more saline. Limestones were deposited as before. In some cases, they were replaced by nodular chert, but always they were dolomitised. Gypsum was at times precipitated; snails were absent. Amelioration of the climate produced the final stage of the lakes.

Owing to its very low volume/area ratio, the lake was very sensitive to climatic fluctuation. Evaporation and precipitation controlled the concentration of dissolved solids and the pH of the lake, its area, and the chemistry
of the mudflats. Its shallowness permitted frequent stirring of the sediments and allowed plants to grow in profusion, both important in determining the productivity and nature of the lake. Wind tides and volume changes exposed and inundated vast areas around the lake, which, with sheetwash, perpetuated the flat lake floor. Many distinctive pedogenic and caliche features formed during periods of exposure.

The occurrence and processes of formation of the major lithologies are as follows.

1) Micrite. In all members, but dolomitised in the Middle Member. Formed by physicochemical (temperature and pH fluctuation) and biochemical (photosynthetic CO₂ degassing) precipitation.

2) Dolomitic intramicrite/intradolomicrite. Primarily upper Lower and Upper Members, also Middle Member. Formed by inwash of fragments of dolomitised limestone from lake marginal carbonate mudflats.

3) Vuggy 100% dolomitic dolostone. Primarily Lower Member and middle unit of Middle Member. Formed by replacement of thin limestone units, probably due to evaporative draw, preceded by solution.

4) 100% dolomitic dolostone. Middle and Upper Members. Evaporative draw, pre- and (or) early burial.
5) Void-filling or void controlled chert. Top of Lower Member, base of Middle Member. Late nucleation from groundwater in voids.

6) Mudcrack-filling chert. In lower unit of Middle Member. Appears to have formed from an opaline gel, which possibly precipitated directly from the lake water by pH fluctuation. Hardening prior to burial and compaction.

7) Chert nodules. Middle and Upper Members. Replacement of limestone prior to pre- or early burial dolomitisation.

8) Silicified limestone beds. Upper Member. Evaporative draw of siliceous high pH water through mud-flats and silicification of limestones prior to dolomitisation, and (or) precipitation due to mixing of same groundwater and infiltrating meteoric water.

9) Bedded gypsum. Upper unit of Middle Member. Precipitated from alkaline brine due to pH fluctuation.

Profitable future studies on the Flagstaff Formation include determining the cause of the blocky calcilutites and the investigation of facies arrangements in very small time zones, to identify shoals, patterns of lake currents, extent of regressions and transgressions, and distribution of snail assemblages, for example. This can be done by following
individual beds up canyons and around mountains. Lake Flagstaff is superbly suited for a study of sediments in a very large lake, perhaps even more so than modern lakes due to its excellent exposures. The Flagstaff and North Horn beds of the Gunnison Plateau need far more study. Continued application of micromorphological studies such as Brewer's (1964) should be rewarding.
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APPENDIX 1

Location of Sections

1) New Canyon - Ephraim Canyon

Ephraim Quadrangle, T. 17 S., R. 3 E.

SE 1/4, Sec. 12: 0.75 km up New Canyon, ascend north wall to top of ridge.

Offset west along top of ridge towards Ephraim Canyon pumping station, beginning in north-central Sec. 13

280 m. Complete section

2) Cove Mountain - Skyline Drive

Ferron Reservoir Quadrangle

Ascends SE side of Cove Mountain (N-C, Sec. 2, T. 19 S., R. 4 E., and SW 1/4, Sec. 31, T. 18 S., R. 4 E.

Offset west towards Skyline Drive. Section starts below ridge, SE to E-C Sec. 33, T. 18 S., R. 4 E.

304 m. Lower, Middle, and 10 m of Upper Member.

3) Ferron Reservoir

Ferron Reservoir Quadrangle, T. 19 S., R. 4 E.

Below Skyline Drive NW of reservoir, NE corner, Sec. 22.

Lower Flagstaff sampled, not measured.
4) Ferron North, Ferron Mountain

Ferron Reservoir Quadrangle, T. 21 S., R. 2 E.
Cliff above White Slides, south of Mill Stream and
Jeep Trail, NW 1/4, Sec. 31.
Lower Member sampled, not measured.

5) Flagstaff Peak

Flagstaff Peak Quadrangle, T. 20 S., R. 5 E.
West end of Peak, Centre, Sec. 13.
80 m. Lowermost Lower Member.

6) Ferron East, Ferron Mountain

Flagstaff Peak Quadrangle, T. 20 S., R. 5 E.
East ridge of Mountain, E-C, Sec. 11
132 m. Lower Member.

7) Jason's Cabin, Ferron Mountain

Flagstaff Peak Quadrangle, T. 20 S., R. 5 E.
NW Sec. 3, NE of Jason's Cabin and Springs.
121 m. Lower Member.

8) Ferron South, Ferron Mountain

Heliotrope Mountain Quadrangle, T. 19 1/2 S.,
R. 5 E.
Ascends gully NE Sec. 32.
81 m. Lower Member.

9) Ferron Top, Ferron Mountain

Ferron Reservoir Quadrangle, T. 19 S., R. 4 E.
South side of the peak in SW Sec. 25.
Includes 37 m above top of Beaver Pond section.

Middle Member.

10) Beaver Pond, Ferron Mountain
    Heliotrope Mountain Quadrangle, T. 19 1/2 S.,
    R. 5 E.
    Ascends gully, NW Sec. 32.
    218 m. Lower and Middle Members.

11) Heliotrope Mountain
    Heliotrope Mountain Quadrangle, T. 20 S., R. 4 E.
    West side of mountain, Central Sec. 2.
    145 m. Lower and Middle Members.

12) Upper Sixmile Canyon
    Black Mountain Quadrangle, T. 19 S., R. 3 E.
    Cliff at top of landslide, SE Sec. 3.
    89 m. Lower Member.

13) Middle Sixmile Canyon
    Black Mountain Quadrangle, T. 18 S., R. 3 E.
    North wall of canyon, NW Sec. 26
    141 m. Middle and Upper Members (?)

14) Palisade Lake, lower Sixmile Canyon
    Sterling Quadrangle, T. 18 S., R. 2 E.
    Above NW corner of golf course, SE Sec. 26.
    49 m. Upper Member Only?

15) Horse Mountain
    Wales Quadrangle, T. 16 S., R. 2 E.
Above end of left fork of Rock Canyon road, east
    front of Gunnison Plateau, SE Sec. 28.

Middle and Upper Members.

16) Wales Canyon

Wales Quadrangle, T. 15 S., R. 2 E.

Up east front of Plateau south of Canyon, W-C Sec.

26 and E-C Sec. 27.

116.5 m. Middle and Upper Flagstaff.

The Flagstaff was also sampled: at Sage Flat east of Flag-
staff Peak; along the north wall of Willow Creek Canyon
(sec. 4, T. 21 S., R. 2 E.); in the hogbacks east of Axtell
and Redmond; in Fairview Canyon, Fairview (roadcur in north
wall); at Birdseye Quarry, north of Indianola and east of
US 89) in Willow and Deer Creek Canyons in the Pavant Range;
at Long Ridge; west of Thistle along US 6; south of Gilluly
in railroad cuts; and north of Castlegate along State 33.
APPENDIX 2

Discussion of Figure 11

For the areas outside the central Wasatch Plateau, Figure 11 relies heavily on the work of others (mapped in Figure 10) and on my analysis and recorrelation of their measured sections and other data. Their work can be found in various Ohio State University theses and dissertations, nearly all unpublished. Considerations of space unfortunately prohibit a complete summary of their work relative to the Flagstaff, but the main pertinent points are presented below.

1) Gilliland (1948) described units A through E in the Valley Mountains and La Rocque correlated A with his lower Flagstaff, B with his middle Flagstaff, and C, D, and E with his upper Flagstaff. B and C are incorrectly correlated. Unit B comprises light grey to black, massive, locally fossiliferous, cliff-forming limestones (Gilliland, 1948). C is unfossiliferous, and comprises red limestone, siltstone, and sandstone. D and E are fossiliferous. There are two fossil faunas described from unit B. One was collected by Gilliland (1948) and contains *Viviparus trochiformis* and *Holospira leidyi* (both unique to the lower Flagstaff, according to La Rocque, 1960), *Physa bridgerensis* (La Rocque lists this as upper Flagstaff, but it is found elsewhere in the lower Flagstaff - see Davis, 1967), and
Planorbis sp., which is not useful. The other collection was collected by Lautenschlager (1952, p. 57), from unit B along Sweet Creek in the centre of the Pavants. It is lower Flagstaff because it also contains V. trochiformis (La Rocque, 1960).

La Rocque (1960) does not mention the first collection, however it appears that he includes the second, as collection 53, dated as lower Flagstaff and attributed to Lautenschlager. The uncertainty is because although the species and even the numbers of specimens of each species in each description are correlative, La Rocque lists it as "Round Valley, Valley Mountains, lower part of the Flagstaff, exact stratigraphic position unknown." Further confusion seems to have resulted from Lautenschlager's comment, "This assemblage is considered middle Flagstaff (A. La Rocque, personal communication)." The apparent ambiguity is a function of La Rocque's revision of his zonation between 1952 and 1960 (La Rocque, 1951, 1956, 1960).

Thus from the fossil evidence units A and B are combined and correlated with the lower Flagstaff. Note that the supralittoral nature of C (no fossils, highly epiclastic, red – see Gilliland, 1948; Lautenschlager, 1952) no longer requires Middle Lake Flagstaff to expand west of the Gunnison Plateau (cf. Weiss, 1969, p. 1118).

2) The northern end of the Juab Valley depocenter is not well known, but Muessig (1951) indicates that the
northern part of Long Ridge is entirely nonlacustrine. Vogel (1957) and Tucker (1954) show that the lake lay primarily to the south and east. Note that this depocenter is tectonic in origin, lying within the eastern edge of the Sevier orogenic belt (the Gunnison Plateau being the easternmost edge) and that the lake facies lies only in the eastern half of the basin, while clastics (fanglemerates, etc.) dominate the western half.

3) In Salina Canyon the North Horn, Flagstaff, and Colton Formations lap onto an old high (Gilliland, 1948; Bachman, 1959; Baughman, 1959; McGookey, 1958, 1960). The Middle Flagstaff here is a wedge of red deltaic and extralacustrine clastics that pinches out north and east into the lake basin (Johnson, 1949; Baughman, 1959). The red, arenaceous "basal" Flagstaff south of the Salina Canyon high by physical tracing (McGookey, 1960) and by ostracodal evidence (Swain, 1956) is of Middle Flagstaff age, whereas the underlying "upper North Horn," although reddish, is lacustrine (McGookey, 1958) and contains Lower Flagstaff ostracodes (Swain, in McGookey, 1960). Thus, needless to say, it is Lower Flagstaff, and the Salina high formed a peninsula that jutted into the lake. This also shows the early Lake Flagstaff to be again more extensive than Middle Lake Flagstaff.

4) The problem of extending the lake south into the Fishlake Plateau and beyond has already been discussed.
The rocks are covered by the Marysvale volcanics, but because the Flagstaff there is predominantly clastic and supralacustrine (McGookey, 1958, 1960; Schneider, 1967) significant southward extension of the lake is unlikely.

5) The Lower Member and all but the upper unit of the Middle Member of the Flagstaff pinch out at Spring City and Mount Pleasant, north of Ephraim (Gill, 1950). The complete section reappears at Fairview (Pashley, 1956). Near Indianola is the site of major clastic input and delta-building that tapered off only during late Flagstaff time (Khin, 1956; Mase, 1957). The presence of the Lower Flagstaff shoreline just east of U.S. 89 is affirmed by Davis (1967).

6) Whether the Flagstaff strata in the Cedar Hills and the southern Wasatch Mountains include Lower Flagstaff lacustrine beds is conjectural, but it is not impossible (see Schoff, 1937, and Metter, 1955).

7) The lowest "Flagstaff" on the east front of the Gunnison Plateau is Katherman's (1949) irregular unit A. Gill (1950) correlates this with his unit 5 (the upper unit of the Middle Member) on the basis of the thin, clastic-rich, and unfossiliferous zone at its base, and its succession by Upper Flagstaff fossils. The earliest lacustrine beds in the eastern Gunnison are very low in the section at Wales Canyon, in "North Horn" (Birsan, 1974). They pass into clastics west, north, and south (Birsan, 1974), and are
overlain by a relatively thin amount of clastics, some of which came from the east, suggesting that if there was a lake here during Lower Flagstaff time, then it was small and isolated (see the discussion of the San Pete Valley high, below).

8) The isopachs of the Uinta basin require a lengthy explanation, as they differ substantially from other interpretations. The Flagstaff beds between Thistle and Sunny-side, traced out by Spieler (1949), were always assumed to belong to the Wasatch Plateau depocenter, not the Uinta basin (see La Rocque, 1960; Davis, 1967). Because the Green River Formation's Lake Uinta was recognised to have joined and succeeded "Lake Flagstaff" in the Thistle area (Metter, 1955; La Rocque, 1960; Davis, 1967), Lake Uinta was generally thought of as almost entirely post-Flagstaff. That the Flagstaff at Thistle is entirely late Flagstaff has long been known (La Rocque, 1956), but the significance of the increasing age of the base of the Flagstaff to the east (discussed by Spieler, 1949, and Davis, 1967) has not been fully interpreted in terms of the history of Lakes Flagstaff and Uinta (cf. Swain, 1956; Davis, 1967; Weiss, 1969). Through subsurface work, Ryder et al. (1976) have shown that the Flagstaff Formation in the Uinta basin is nowhere older or thicker than at Soldier Summit. The beds to the southeast represent a brief early (early Lower Flagstaff) expansion of the lake. The lake extended perhaps 58 km eastward
along the Book Cliffs (see Fisher, 1936; Abbott and Liscomb, 1956; Ryder et al., 1976). These lake beds are separated from the Upper Flagstaff by some Colton Formation that is equivalent in age to the Middle Member of the Flagstaff elsewhere (Swain, 1956, in part quoting La Rocque). (Again the Middle Member lake does not extend as far as the others.) Thus a lake existed east of Thistle during the early stage of the lake. Ryder et al. show the Flagstaff along the southern edge of the basin to be a facies of the Green River Formation: the Flagstaff comprises the frequently exposed carbonate/silty mudflats that fringe the central permanent (Green River Formation) playa lake, thus making the base of the Green River Formation the same age as the early Flagstaff. Thus thicknesses of lacustrine sediments can be obtained from the data of Ryder et al. To use the data of Davis (1967) requires one further recorrelation. His facies map for the late Lower Flagstaff clearly shows the encroachment of shorelines from the west and north and progradation of the Colton floodplain from the southeast. Because Davis notes that animal and plant life are greatly diminished and that the rocks are dolomitic, because Prescott (1958) reports minor gypsum in these rocks, because the Colton progradation is probably the one dated by Swain (1956), and because the succeeding stage (Davis, 1967) encompasses the westward transgression that Ryder et al. date as Upper Flagstaff, I feel that this reconstruction depicts
the Middle Flagstaff and not just before it. Davis admits the distinction between units 1 plus 2 and 2 plus 3 are hard to make, and undoubtedly his dating was strongly influenced by La Rocque's hypothesised expanded lake of unit 2.

9) Kucera's (1954) values indicate that the Flagstaff thins to the north and east of the centre of the Wasatch Plateau, as does Gill's (1950) section at the Horseshoe. Because these sections are critical but not well detailed and because the northeast corner of the Plateau is not well known, the area requires reinvestigation.
APPENDIX 3

Measured Section, Ferron Mountain

On the recommendation of my advisers, I have included the following section of most of the Lower and Middle Flagstaff from the south-central side of Ferron Mountain. It comprises two approximately continuous sections (Ferron Top and Beaver Pond - sections 9 and 10 of Appendix 1) and is summarised diagrammatically in Figure 8. The reader is cautioned that the terms in this section are field terms and as such are frequently wrong and misleading. In particular we severely overestimated the clay content of the mudrocks and argillaceous micrites and underestimated the amount of dolomite in the Lower Member. Microscopic examination of polished slabs shows countless features that were not visible in the field; 'homogenous micrite' is a particularly sorry term in this respect. For a section with more analytic control the reader is referred to the Cove Mountain section described in McCullough (1977). To aid anyone visiting the area, our field units (White 1, Blue 1, etc.) are retained; they refer to the major light and dark coloured bands visible in the Ferron Mountain cliffs. The section starts at the summit of Ferron Mountain.

<table>
<thead>
<tr>
<th>Height from base (metres)</th>
<th>Thickness (cm)</th>
<th>Lithology of bed</th>
</tr>
</thead>
<tbody>
<tr>
<td>Top of &quot;Ferron Top&quot; section, in upper unit of Middle Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td>255.07 m</td>
<td>25 cm</td>
<td>dolomite; thin-bedded, vuggy, resistant, grey</td>
</tr>
<tr>
<td>300 cm</td>
<td></td>
<td>slope-forming dolomite and mudrock</td>
</tr>
<tr>
<td>87</td>
<td></td>
<td>dolomitic siltstone</td>
</tr>
<tr>
<td>250.95</td>
<td>557</td>
<td>gypsum; bedded, top 4m are cliff-forming</td>
</tr>
<tr>
<td>68</td>
<td></td>
<td>cover, dolomite near top</td>
</tr>
<tr>
<td>30</td>
<td></td>
<td>nodular dolomicrite</td>
</tr>
<tr>
<td>150</td>
<td></td>
<td>gypsum beds but mostly dolomite/gypsum breccias</td>
</tr>
<tr>
<td>242.90</td>
<td>790</td>
<td>relatively pure dolomicrite and mudrock; cover</td>
</tr>
</tbody>
</table>
235.00 m 1650 cm largely covered, slope forming mudrocks and dolomites. Shale reddish below 1500 cm, brown above. Thin dolomitised micrites.

Base of "Ferron Top" section

Top of 'Beaver Pond" section

Top of scree covered slopes above Ferron cliffs

218.50 220 12-13 beds of vuggy slightly calcareous dolomite, contains snail moulds and some intraclasts. Constitutes middle unit of Middle Member.

216.30 30 dolomicrite; thin, straight, evenly-spaced vertical moulds ("straws").
47 creamy white dolomite; burrows, rootlet moulds, stylolites.
10 shale parting, mudcracks
74 white dolomicrite; chert nodule rich horizon near top, layer rich in rootlets, nodules vertical.
99 dolomite; 'straws'.
154 relatively homogenous dolomicrite.
8 chert.
10 completely homogenous dolomite
25 pure dolomite; burrowed,rootlets
20
30 intraclastic 'splintery' dolomicrite.
4 blackish brown organic shale.
132 'splintery', pure homogenous dolomicrite.
14 dolomite; resistant bed, loaded with rootlet moulds.
18 organic shale.
86 white 'splintery' dolomicrite, irregularly intraclastic.
6 blackish brown fissile organic shale
37 dolomicrite, chert nodules at base

208.22 7 grey mudrock
37 dolomicrite with rootlets
5 shale parting
14 pelletal dolomicrite
50 shale parting
63 dolomite, irregularly fractured
50 shale parting
28 mottled intraclastic dolomite
10 mottled intraclastic dolomite
74 platy dolomite with massive brown pink intraclasts

205.83 134 mudrock
30 dolomicrite, fizzes sluggishly
143 mudstone
20 relatively pure dolomite, some grey clasts, fizzes sluggishly, dispersed chert
grey shales
dolomitised limestone, rootlets in centre, chert
mudrock
dark dolomite with bimodal white dol. clasts
mudrock
vertically fractured largely intraclastic dol.
red (& minor green) mudrock
silty, vuggy dolomite
grey-green and red mudrock
red mudrock
dolomitic siltstone
green mudrock
red and green mottled mudrock
sandy burrowed dolomitic mudrock
red mudrock with green shale clasts
green mudrock, mottled green-purple and purple in upper half.
green mudrock
dolostone, burrowed, solution vugs
sandy light green dolomitised limestone with clasts and fossils (fragmented)
vuggy "argillaceous" dolomite
mudrock
purple/green mottled sandstone
vuggy green brown "arg." dolomite, fizzes slightly
as above, but greenish-grey
mudrock, some calcite
red mottled mudrock
green mudrock
grey brown poly lithologic bioclastic dolomite, with root moulds.
mudrock, lower 60 cm laminated
mudrock
slightly calcareous mudrock, medium grey-green
mudrock and shale
mudrock
mudrock
poly lithologic intraclastic dolomite
dark grey brown bioclastic limestone
mudrock
shale
mudrock, calcareous at top

Base of Middle Member/Top of Lower Member

poly lithologic intraclastic micrite with fossil fragments. Many root moulds
finely granular mudrock
bioclastic poly. intramicrite, some whole snails
mudrock, calcareous at top (calichification?)
micrite, root moulds filled with clasts. Knobby top.
mudrock.
micrite with small intraclasts & some bioclasts
<table>
<thead>
<tr>
<th>Interval</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>153.87 m</td>
<td>dolomite marker bed; homogenous, very light brown some portions crystalline 'arg.' micrite with pellets mudrock faintly laminated light grey calcareous dolomite mudrock, standard grey, angular blocky granules homogenous micrite; root mould filled with intraclasts mudrock bioclastic intramicrite shale parting sparsely intraclastic (poly.) micrite mudrock homogenous micrite, sparse fossil fragments mudrock with clasts and fossils homogenous micrite, locally intraclastic (poly.) shale polyolithic intramicrite homogenous micrite shale vuggy crystalline micrite mono. intramicrite, clasts small and round shale parting micrite with small voids and dispersed spar mudrock granular spar-rich micrite light grey brown micrite with whole and broken snails mudrock poly. intramicrite thin platy limestone</td>
</tr>
<tr>
<td>143.81</td>
<td>dolomite marker bed; root moulds with intraclastic and bioclastic debris dolomicrite, burrowed, recrystallised, root moulds laminated micrite greenish mudrock mudrock mottled intramicrite with chert, scattered ostreocodes bioclastic intramicrite with root moulds</td>
</tr>
</tbody>
</table>
light brown micrite with whole snails
mainly mudrock and cover
micrite
mudrock and cover
pure, homogenous micrite
mudrock
homogenous micrite with root moulds, dolomitic, monolithologic intraclasts at top
cover
4 beds of largely homogenous micrite
intramicrite (mono.)
mono. intramicrite, perhaps a few other clasts, a few Hydrobia
micrite, locally mono. intraclastic
micrite, locally mono. intraclastic
cream coloured micrite; root moulds with intra-
clastic and some bioclastic debris
laminated poly. intramicrite, clasts dolomitic
disrupted, crystalline micrite
dolomicrite
3 beds; intradolomicrite at top, base dolomicrite
4 beds; white micrite, some clasts, root moulds at top
poly. intramicrite
light grey micrite, white dolomite intraclasts, local broken snail shells. Homogenous at top, with roots
white laminated micrite, local intra- & bioclasts
dark poly. intramicrite, sparse fossil fragments
top burrowed, base homogenous micrite
dark bioclastic micrite at top, burrows and clasts at base
homogenous micrite
mudrock
homogenous micrite, light greyish brown, solution features
shale parting
homogenous micrite, solution features
"shale parting
bioclastic intramicrite, white
homogenous micrite with whole snails, calcite-
filled 'straws' and rootlet at top
whole snails at base of white homogenous micrite
Top of Blue 4
dark-biomicrite
dark bioclastic biomicrite
Base of Blue 4 / Top of White 3
micrite; bioclastic; intraclastic at base. Many 'straws' and rootlet moulds.
many straws and burrows, top 10cm bioclastic
shale parting
micrite with rootlets and 'straws', top bioclastic and poly. intraclastic, base monolithologic.

light cream coloured chalky micrite, much spar, some snail fossils, mono. intraclasts
shale parting
grey micrite with whole snails, top half mono. intraclastic, slasts derived from bottom half of bed, 'lifted up'.
shale parting
same as 17 above, thin grey bioclastic layer on top
shale parting
mudrock
homogenous micrite, knobby top.
white monolithologic intradolo?micrite
same, large white clasts, some containing rootlets
white poly. intramicrite, some vuggy horizons
white mono. intramicrite
intramicrite, polythiologic, clasts mostly white dolomite, matrix calcareous

Base of White 3 / Top of Blue 3

medium grey micrite with sparse fossils
micrite; tiny white and black clasts in burrows
shale parting
brecciated and earlier burrowed horizontally laminated beds
biomicrite with local aggregations of rounded white biomicrite clasts. Burrows contain clasts derived from overlying bed
sparsely bioclastic limestone with an uneven base and a gradational top. Burrows rich in bioclastic debris
shale parting
pinkish micrite with 'straws', whole snails, and burrows rich in poly. clastic and bioclastic debris
biomicrite, medium grey, whole snails and some light grey intraclasts, both increasing upward
grey biomicrite with pink snail fragments
shale parting
includes pinkish brecciated intrabiomicrite (poly.)
brecciated pinkish burrowed micrite
thin yellow micrite
light pink micrite
white clasts in bluish grey micrite

Base of Blue 3 / Top of White 2

vertically platy white intramicrite; three zones - white clasts in white matrix with fossils, thin shale, white and yellow clasts in bluish matrix and on top bluish clasts in white matrix
white mono. intramicrite, sparse fossils, straws, roots
white micrite, whole fossils, rare small clasts 'straws', burrows
white mono. intramicrite, fossils, sparse rootlets, 'straws', burrows

Base of White 2 / Top of Blue 2

101.39
43 dark biomicrite, uneven top, burrows with clasts from bed above. Middle is buff white homogenous micrite with local mono. intraclasts and fossils. Burrows or root moulds into basal light biomicrite contain clasts and fragments of micrite derived from middle layer

2-8 shale

19 light grey burrow-mottled intrabioticite

5 shale

12 blue bio (poly.) intramicrite

2 shale parting

49 whole Physa fragments and very small clasts

31 light grey poly. intramicrite (clasts brown) fossils rare, grades down into

14 nodular poly. intramicrite, shaly at base

2 shale parting

11 dark grey poly. intramicrite, capped with biomicrite pink micrite in burrows

2 shale

69 intrabioticite

98.76

6 blue micrite

.5 shale parting

9 pink biomicrite with 'straws', overlain by blue layer with whole snails

12 pink overlain by an irregular thin blue biomicrite sh. parting

.5 blue biomicrite

7 "

.5 shale parting

6 blue biomicrite

.5 shale parting

10 pink biomicrite

1 shale parting

7 blue biomicrite

4-0 shale parting, irregular, impersistent

8 pink biomicrite, layered, blebby, some clasts, solution seams and other features

10-2 irregular shale parting

19 four pink over blue cycles

.5 shale parting

3 mudrock

8 variable shale and snail coquina, some pink shells

12 brown micrite with 'straws' and burrows, solution

.5 shale parting

4 dark blue grey dense bioclastic micrite

.5 shale parting

7 as 4 above

1 shale parting
dark blue biomicrite with whole snails, pink at top
shale parting
as 11 above
irregular shale parting
biomicrite, whole and crushed snails, vertical burrows
shale parting
'argillaceous' limestone
shale
'argillaceous' limestone
fissile grey green mudrock
'argillaceous' limestone
homogenous micrite, 'straws'
local shale parting, undulatory contact
biomicrite, some crushed snails
organic, highly fossiliferous shale, whole snails
undulatory top
intrabiomicrite, pink and dark fossil fragments
shale
bio (poly.) intramicrite
monolithologic intramicrite
mudrock
poly. intramicrite, some whole snails
shale
fetid biomicrite, some intraclasts
shale parting
as 20 above
mudrock
dark poly. intrabiomicrite, clasts local, root moulds
mudrock
basal 5cm shaly grades into 3 beds of biointrinsicite
base is homogenous micrite, grades up into 3
bio (poly.) intramicrite beds. 'Straws' at top
mudrock
homogenous micrite with vugs
mudrock (not blocky) with snail fragments
homogenous micrite, passing up into dark biomicrite
with a few whole snails, some vugs
shale
homogenous micrite at base, burrows and areas with
clasts above
brownish micrite with clasts and fossil fragments
filling burrows
mudrock
poly. intrabiomicrite, both fossils and clasts rare
and small, some snails whole
bio (poly) intramicrite
17 homogenous micrite
7 mudrock
35 homogenous micrite, vertical burrows at top
15 mudrock
34 small dark grey and medium brown clasts in light grey micrite
32 micrite with fine mudcracks and burrows
.5 shale parting
32 very fine grained light brown micrite with a very few whole snails
1 grey shale
3 grey micrite (weathered white), clasts in burrows mudcracks, undulatory top and base
.5 shale parting
3 white clasts in light grey laminated argillaceous? micrite
1 shale

Base of Blue 2 / Top of White 1
89.64
37 white clasts in white micrite, burrows
.5 shale parting
29 brown micrite (weathers white), whole snails, laminated near top
1 shale parting
30 homogenous brown micrite (weathers white) in basal 20cm, laminated and burrowed above
15 irregularly fractured micrite, base very irregular, grades from shale
3 shale parting, intraclastic, clasts pink, organic
375 grey with white clasts at base, grades up into vertically fractured white intramicrite, which turns light grey, fossils scarce, clasts white 3cm thick brecciated bed
23 dark organic biomicrite, some snails crushed
7 light grey shale

Base of White 1 / Top of Blue 1
84.26
45 biomicrite with a few small and tiny clasts
25 biomicrite, some light brown and black clasts, local whole snails and clams at base
8 mudrock, grading up into 25
44 blue biomicrite, some whole snails
24 shale, grades up into 44
101 5 beds, very small clasts in most, knobby tops, burrowed, whole Physa micrite, 'argillaceous' at base
27 mudrock
18 slightly intraclastic to granular biomicrite
4 highly calcareous shale
13 as 18 above
.5 shale parting
<table>
<thead>
<tr>
<th>17</th>
<th>slightly fossiliferous recrystallised micrite</th>
</tr>
</thead>
<tbody>
<tr>
<td>18</td>
<td>dark fractured micrite, 'straws', no fossils</td>
</tr>
<tr>
<td>30</td>
<td>same, burrows</td>
</tr>
<tr>
<td>14</td>
<td>slightly fossiliferous recrystallised micrite, irregular top and veins of spar. Chara</td>
</tr>
<tr>
<td>25</td>
<td>pinkish burrowed micrite, irregular top, burrows filled with dark organic intraclastic micrite</td>
</tr>
<tr>
<td>16</td>
<td>'argillaceous', bioclastic, pebble bearing (rounded and well lithified intraclasts?) micrite</td>
</tr>
<tr>
<td>5</td>
<td>dark poly. intrabioticite</td>
</tr>
<tr>
<td>1</td>
<td>shale</td>
</tr>
<tr>
<td>4</td>
<td>fetid dark poly. intrabioticite</td>
</tr>
<tr>
<td>20</td>
<td>shale, snail fragments</td>
</tr>
<tr>
<td>23</td>
<td>highly organic shale, grades up into resistant limestone</td>
</tr>
<tr>
<td>17</td>
<td>very dark bioclastic micrite with brown and black clasts</td>
</tr>
<tr>
<td>51</td>
<td>mudrock</td>
</tr>
<tr>
<td>10</td>
<td>'arg.' limestone, a few fossils, very few small white, yellow clasts. Shale parting at base</td>
</tr>
<tr>
<td>115</td>
<td>pinkish blue much fractured bioclastic, minor clasts shale parting</td>
</tr>
<tr>
<td>5</td>
<td>bio (poly.) intramicrite</td>
</tr>
<tr>
<td>16</td>
<td>platy mudrock</td>
</tr>
<tr>
<td>56</td>
<td>platy pinkish brown bioclastic</td>
</tr>
<tr>
<td>5</td>
<td>dark shale</td>
</tr>
<tr>
<td>16</td>
<td>dark 'argillaceous' micrite with snail fragments</td>
</tr>
<tr>
<td>25</td>
<td>mudrock</td>
</tr>
<tr>
<td>57</td>
<td>blue-grey micrite with large angular yellow clasts</td>
</tr>
<tr>
<td>31</td>
<td>similar to 57, small clasts</td>
</tr>
<tr>
<td>63</td>
<td>yellow intraclastic 'arg.' bioclast, grades into 31 brown shale</td>
</tr>
<tr>
<td>43</td>
<td>very shaly blue intrabioticite (pololithologic)</td>
</tr>
<tr>
<td>64</td>
<td>bioclastic poly., intramicrite, grades into shale and then into 15cm bioclastic micrite</td>
</tr>
<tr>
<td>83</td>
<td>5 layers of pololithologic biointramicrite shale parting</td>
</tr>
<tr>
<td>34</td>
<td>brown micrite with a few small intraclasts at base grades into standard brown poly. intrabioticite shale</td>
</tr>
<tr>
<td>2</td>
<td>nodular dark blue locally poly. intraclastic biomicrite, clasts up to 1 cm, basal 5cm shaly shale parting</td>
</tr>
<tr>
<td>27</td>
<td>dark blue grey locally intraclastic bioclastic micrite clasts pololithologic, 2mm max</td>
</tr>
<tr>
<td>50</td>
<td>blue grey, locally shaly bioclastic micrite</td>
</tr>
<tr>
<td>57</td>
<td>yellow &amp; grey mudrock, becomes grey above</td>
</tr>
<tr>
<td>71.51</td>
<td>mudrock, gradational from bed below</td>
</tr>
</tbody>
</table>
Base of Blue 1, beds below are distinctly yellow

71.09
20 very organic, red-mottled, fossiliferous, calcareous mudstone
5-6 grey calcareous fossiliferous mudrock (not blocky)
58 mottled mudrock
234 variegated mudrock
9 micrite, much spar
62 yellow mudrock, also variegated
10 calcareous siltstone, laterally equivalent to a 1.3m (max) sandstone lens
10 reddish mudrock
50 grey yellow mudrock
25 poly lithologic intrabioticrite
20-28 light brown poly. intrabioticrite, burrows
30 'straws', some whole snails, irregular top
20 brecciated white micrite, irregular top
.5 shale parting
17 homogenous micrite, local spar and peloids
230 yellow mudrock with several thin red layers
5 micrite with rootlets and much spar
10 mudrock
7 micrite with rootlet moulds and much spar
25 organic mudstone
18 poly. intrabioticrite, rootlet moulds, burrows
50 shale
1 as 18 and 50, whole snails present
39 shale
19 poly lithologic intramicrite
40 white nodular, 'argillaceous' poly. intramicrite
1 shale
59.69 131 micrite, whole snails, large white & yellow clasts
with ostracodes, also blue-black clasts
massive
35 highly organic fossiliferous calcareous mudrock
5 mudrock
14 monolithologic intrabioticrite
27 mudrock
9 mono. intraclastic at base, bioclastic throughout
.5 shale parting
25 as 9 above, spar throughout
.5 grey shale parting
25 'arg.' micrite with small yellow, black clasts at
base, sparry micrite above
20 mudrock
11 'arg.' micrite with large yellow, black clasts
27 mudrock
44 as 11 above, clasts smaller
56.55 130 yellow mudrock
dark biomicrite
yellow brown mudrock
'targillaceous' micrite
yellow mudrock
red shale
calcite rich sandstone, a few micrite clasts
limestone at base
grey calcareous mudrock
mottled red claystone
grey calcareous mudrock
'targillaceous' mudrock
mudrock
recrystallised homogenous micrite
mudrock
micrite with a few burrows and rootlet moulds
grey mudrock
yellow mudrock
red lithoclastic shale
dark 'targillaceous' poly lithologic intramicrite
brown mudrock
homogenous micrite at base, root moulds and
fossil fragments above, top 5cm poly. intraclastic
micrite with burrow (nonvertical)

shale
micrite
shale
'targillaceous' limestone
mudrock
sparsely fossiliferous micrite, a very few intraclasts
'targillaceous' limestone, gradational with 24
light grey calcareous mudrock
moderately even-textured micrite
shale parting
as 8 above
shale parting
massive homogenous micrite
same, with calcite-filled 'straws'
shale
fossiliferous micrite, some tiny yellow clasts
top; biomicrite with 'straws', centre; biomicrite
with some whole snails, base; massive homogenous
micrite
5 irregular beds of sparry locally intraclastic
micrites. 'straws' in topmost
mudrock
poly lithologic intramicrite at top, homogenous at base
shale parting

'argillaceous' micrite, intraclastic at top

shale parting

slightly laminated brecciated intramicrite

shale parting

cream coloured biomicrite

dark mudrock

polyolithologic intrabioticite, 'argillaceous'
mudrock

'argillaceous' polyolithologic intramicrite

dark shale

mudrock

vertically fractured homogenous micrite

yellow brown mudrock

sparry biomicrite, rootlet moulds, a few 'straws'

shale parting, red brown

micrite with spar, 'straws'

shale

mottled micrite

yellow mudrock, grades up int black mudrock

poly. intramicrite, rare fossils, capped by
thin biomicrite

shale parting

slightly fossiliferous poly. intramicrite; lens

shale parting

micrite with lots of spar

dark grey mudrock

fractured monolithologic intramicrite

grey mudrock

relatively homogenous micrite, some spar

shale

as 28 above

shale parting

homogenous micrite

shale

sparse tiny orange-yellow clasts in micrite

relatively homogenous micrite

mudrock

base micrite with faint carbonaceous partings,
top relatively homogenous with sparse fossils

mudrock

intraclastic 'argillaceous' micrite

laminated micrite with faint carbonaceous partings

shale parting

homogenous micrite

micrite, sparse ostracodes

mudrock

sparsely fossiliferous micrite

top half poly. intramicrite, base homogenous

mudrock
16-30 homogenous micrite at base, capped by 3-6cm poly. intrabiomicrite
  .5 shale parting
25.20 33 homogenous micrite
  6 mudrock
  25 homogenous micrite
  23 nodular micrite in mudrock
105 top dark grey mudrock, centre 10cm red mudrock
  base red shale clasts in brown mudrock
  23 intramicrite
  37 micrite, blocky
  .5 shale parting
  19 fractured micrite, relatively homogenous
125 mudrock
  12 sparry homogenous micrite
  40 mudrock
  7 micrite
  .5 shale parting
24 relatively homogenous micrite
  5 mudrock
  12 as 24 above
  62 mudrock
  12 homogenous micrite
  2 mudrock
  15 homogenous micrite
15 same with fossil fragments, shaly at base
  7 homogenous micrite
  1 shale parting
  12 homogenous micrite
  2 micrite
  68 mudrock
  6 relatively homogenous micrite
138 mudrock
  6 argillaceous limestone
  32 blocky micrite and mudrock
  7 'argillaceous' micrite
  96 mudrock
  7 'argillaceous' micrite
  13 mudrock
16 lens of relatively homogenous micrite
15.25 205 yellow mudrock
  23 'argillaceous' micrite
  18 shale/calcareous mudrock
  6 'argillaceous' micrite
  58 mudrock
  6 'argillaceous' micrite with shale clasts
  55 dark grey mudrock
  6 'argillaceous' micrite
  3 shale
  4 'argillaceous' micrite, as usual, no fossils
relatively homogenous micrite
same
sparry micrite with 'straws'
shale parting
relatively homogenous micrite
mudrock
homogenous micrite
mudrock
as 3 above
mudrock
homogenous micrite
mudrock
'argillaceous micrite
homogenous micrite
shale parting
'argillaceous micrite
mudrock
poly. intrablastic micrite
shale parting
sparry micrite
mudrock
homogenous micrite
shale parting
same as 19 above
'argillaceous micrite
shale parting
homogenous micrite
shale parting
homogenous micrite
mudrock
homogenous micrite
same
same
'argillaceous micrite
mudrock, grades up into argillaceous micrite
homogenous micrite
mudrock
homogenous micrite
mudrock
blocky homogenous micrite interbedded with shale
mudrock
platy sparry micrite, knobby top
brown mudrock
'argillaceous micrite
'argillaceous micrite
shale parting
'argillaceous micrite
'argillaceous micrite
homogenous micrite
107 grey and brown mudrock
36 burrowed relatively homogenous micrite knobby top
105 brown mudrock
52 relatively homogenous micrite, burrowed, knobby top.

Base of section

Section begins at lowest outcrop above willows on base of scree slope. Correlation eastward to Flagstaff Peak suggest that the base of the Flagstaff Formation is at least 30 m lower (see Figure 9)
Plate 1

left to right, from upper left

A: The supralittoral facies of the Flagstaff – a piedmont fanglomerate from Long Ridge.

B: Thinly bedded, flat-lying lacustrine limestone of the Lower Member at Beaver Pond and South Ferron sections on the south side of Ferron Mt.

C: Mudcracks filled with dolomite, in chert that has replaced limestone. Middle Member.

D: Horizon of spherical chert nodules, here in a laminated calcilutite in the lower unit of the Middle Member at Cove Mt. (Pen in lower left for scale).

E: Mudcracks in laminated calcilutite, filled by the overlying limestone prior to the dewatering and compaction of the calcilutite and the production of its laminations.

F: Mudcracks in laminated calcilutite, filled by a limestone that is distinct from the overlying bed prior to compaction and the development of lamination.

G: Mudcrack in calcilutite, filled with chert. The chert hardened before compaction – note how the overlying carbonate was impaled on the nodule as the calcilutite was compressed. Other chert mudcrack-fillings on this horizon have been pushed down through the calcilutite as well. In lower unit of the Middle Member.

H: Non-branching rootlet moulds, X1.5

I: Geodic moulds of rush? stems or leaves. These are usually straighter and further apart.

J: Caliche crusts and pseudobedding in a Middle Member dolomite. Note the unusual layers of chert pods and stringers below the hammer head. Cove Mt.
Plate 2

clockwise from upper left

A: Two thin layers of limestone that were pulled apart in situ without disturbing the surrounding beds. This must have been due to differential dewatering while the unit was still mushy.

B: Mudcracks typical of the Upper Member; megascopic, filled with fine intraclastic debris and silicified throughout. This is about modal thickness for a bedded chert.

C: Knobby or nodular limestone. It is very friable, and is due to the remobilisation of carbonate, which is caused by oscillations of the water table (Freytet, 1973). The columnar aspect is probably caused by the growth of roots. The overlying bed is one of the rare fluvial (channelform) sandstones. Lower Flagstaff, Flagstaff Peak.

D: Dendroid rootlet moulds. Lower fresh unit, Middle Member. X2.5

E: See G

F: Pedotubules formed by mobilisation of carbonate around rhizomes or roots. Paludal facies, Long Ridge.

G: G and E are the top and side views of mudcracks in and filled with dolomite. Middle Member.
Plate 3

vertically from upper left

A: Top view of silicified limestone containing chert-filled skew planes that were caused by the collapse of the walls of a large mudcrack to the right of the photo. Note the ostracode shells, which have been replaced by chert. Upper Member, X10

B: Numerous entire, partial, and crushed ostracodes; evidence of an ostracode bloom. Preserved in a Middle Member chert nodule, X27

C: Grumelous bioclastic charophytic limestone. Mollusc and ostracode shells have been dissolved, but "crumbs" are still discernible at the top of the photo. Compare to figure B, plate 5. Lower Member, X12

D: Chert/dolomite boundary. The spherules in the dark dolomite to the right are micronodules of chert that are interpreted to have been precipitated as the chert migrated toward the main chert nodule. Note their gradation into the nodule. This photo encompasses only the very edge of that nodule and shows a rind of chert that was deposited after dolomitisation, which destroyed the texture of the original limestone. Thus little texture is present in this part of the nodule (note however the partial shell in the middle left of the photo). Middle Member, X17

E: The edge of another chert nodule. The chert is to the left and has preserved the grumelous fabric of the original limestone. The clasts are not very distinct where contiguous, but are made obvious by the stellate "voids" that are filled with chert (after sparite?). The voids were formed during early diagenesis. Although not apparent in the photo, there are small aggregations of pyrite along the chert/dolomite contact. Middle Member, X15

F: Arrested incipient nodular bedding. The light-coloured pods began forming and insolubles were concentrated, but the pods are now bounded by early, poorly developed stylolites. Lower Member, X2

G: The crumblike fabric can form in situ: here, clasts are being differentiated from the matrix by recrystallisation along craze planes that originated due to internal contraction. Upper Member, X12
Vertically from upper left.

All natural size unless indicated

A: Polylithologic intraclastic limestone. All clasts are calcareous but differ in size, colour, roundness, fossil content, etc. Lower Member, X.7

B: Intraclastic limestone containing mostly white dolomite clasts in a limestone matrix. It is hypothesised that the dolomite was ripped up from the mudflats and carried into the lake by a flood. Lower Member.

C: Preservation of laminae in a chert nodule, while the texture has been destroyed in the surrounding dolomite. The boundary here appears very abrupt, but chert nodules are present in the dolomite up to a centimetre away (compare fig. D, plate ). The grey areas were silicified only after dolomitisation (note the white "islands" of remnant dolomite present therein). The apparent boundary is exceptionally dark due to aggregates of tiny crystals of pyrite at the very edge of the pure chert. Middle Member, X.7

D: Two ortho-striotubules (burrows filled with debris derived from the same bed and showing a reticulate or watch-glass structure), moderately well preserved in a chert nodule from the upper saline unit - the chert/dolomite boundary can be seen at the top of the photo.

E: A discrete para-isotubule - a burrow filled with material (red clay) from an unknown source. (Discrete implies that the burrow filling does not adhere to the rock.)

F: Intraclastic bioclastic limestone grading up into a laminated fetid bioclastic limestone. The base was deposited under conditions of fairly high energy.

G: Dark and hydrocarbon-rich, or sapropelic, bioclastic limestone. The light areas are burrows that are filled with intraclastic limestone and which were deformed while very soft. Note the many tiny and unbroken snails.

H: An unusual occurrence of well-spaced vertical metajoint planes in a rock that is atexitural due to
Plate 4 (continued)

dolomitisation but has since been marmorised (note splotchy colouring).

I: Marmorised oncolitic limestone from Birdseye Quarry.

J: Like F, but is crumb-like and contains some whole snails in its lower half, suggesting deposition in a lower energy environment.
clockwise, from upper left

A: Undissolved, "clean," unit rhombs of calcite from the Lower Member at Cove Mt. SEM photo, X 4,300.

B: Bioclastic charophytic limestone. Shells have been partly dissolved. Features elsewhere suggest that this was once a crumblike limestone that has been altered enough that the crumbs are no longer identifiable. Alteration did not include formation of stellate voids (compare figure C plate 3). Lower Member, X 12.

C: Sheet and prism cracks (horizontal and vertical respectively): debris-filled components of mudcracks that have undergone minor compaction but also extensive sidewall collapse and secondary cracking. Lower Member, X 1.2

D: Intracrystalline limestone. Clasts are either calcitic (stained dark) and derived from the underlying bed or they are dolomitic (unstained and angular). Skew planes, caused by internal contraction, can be seen in the calcite, X 1.

E: Oncolite. This is formed around a pelecypod shell. Others are formed around snail shells, pebbles, or twigs. Ones this large are possibly formed in situ by inorganic precipitation, X 0.6

F: Abraded clay fragments from red shale; lower unit of the Middle Member, Cove Mt. SEM Photo, X 5,100

G: Polyphased nodule in a dark, fetid, bioclastic limestone. Lower Member. X 1.

H: Greatly dissolved limestone with well-developed stylolites and stylolite-bound meta-isotubules (burrows filled randomly with debris from the overlying bed). X 1.
Plate 6

clockwise from upper left

A: Completely silicified algal (oolitic/pisolitic) limestone. Most oolites are irregular and are formed on ostracodes, algal fragments, and grains of calcite and quartz. X 35

B: Secondary cracking: failure of the walls of a prism crack. Cracks are filled with spar. X 35

C: Crumbl like limestone with degraded crumbs but prominent calcite-filled "stellate voids." Lower Member, X 35

D: Craze planes fully developed and preserved in a Middle Flagstaff chert nodule. Pyrite and pseudomorphous hematite are concentrated near the chert/dolomite boundary (not shown). Cracks filled with clear chert. Etched in boiling NaOH. X 35