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Sequence stratigraphy and biostratigraphy of the lower member of the Deep Spring Formation: Implications for the Neoproterozoic-Cambrian boundary in the Basin and Range Province, western United States

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SEQUENCE STRATIGRAPHY AND BIOSTRATIGRAPHY OF THE LOWER MEMBER OF THE DEEP SPRING FORMATION: IMPLICATIONS FOR THE NEOPROTEROZOIC-CAMBRIAN BOUNDARY IN THE BASIN AND RANGE PROVINCE, WESTERN UNITED STATES

by

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A thesis submitted in partial fulfillment of the requirements for the degree of

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in

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ABSTRACT

The Deep Spring Formation of the southern Basin and Range Province provides information valuable in locating and correlating the Precambrian-Cambrian boundary in western North America. This study provides a sequence-stratigraphic analysis of the lower member of the Deep Spring Formation and a revised Neoproterozoic-Cambrian biostratigraphy for the southern Basin and Range Province.

Sequence-stratigraphic analysis of the lower Deep Spring Formation revealed three sequence boundaries. Because the formation represents a mixed carbonate-siliciclastic depositional system that does not fully conform to conventional models for homogenous systems, a modified sequence-stratigraphic model is proposed. The proposed model includes early highstand slumps, a feature interpreted to be unique to mixed systems. Early highstand slumps of both carbonate and mixed carbonate and siliciclastic sediment form at the onset of the Highstand Systems Tract. During this time, carbonate production resumed on the shelf following a lag in carbonate deposition that had resulted from siliciclastic sediment being deposited across the shelf during relative sea-level lowstand and transgression. This lag in sedimentation resulted in an oversteepening of the shelf during the Lowstand Systems Tract.
and Transgressive Systems Tract that facilitated the down-slope deposition of the early highstand slumps.

This study also identified the small shelly fossil *Cloudina* in the lower Deep Spring Formation. *Cloudina* was previously unrecognized at this stratigraphic level in the Basin and Range. This Neoproterozoic fossil is associated in the lower Deep Spring Formation with the shelly fossils *Nevadatubulus* and *Sinotubulites*, which traditionally are believed to be Cambrian in age. Limestone beds containing the shelly fossils are found stratigraphically lower than the shale and siltstone strata that contain Late Proterozoic trace fossils. *Phycodes pedum*, the trace fossil that signals the beginning of the Cambrian is not found in the lower Deep Spring Formation, but it is present in the upper member of the Deep Spring Formation. Therefore, the lower member is interpreted as Neoproterozoic in age and not Cambrian, as previously reported. This information, combined with the occurrence of a lowermost Cambrian body fossil from the middle member reported by Signor and others (1994), suggests the Neoproterozoic-Cambrian boundary lies somewhere between the last occurrence of *Cloudina* in the lower member and the first occurrence of Cambrian body fossils in the middle member of the Deep Spring Formation.
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I wish to thank my family for their constant support. In particular, I would like to thank them for raising me ignorant to the fact that some people believe men are more capable than women, and that science is a “man’s profession” – by the time I realized that belief was out there, it was already too late to stop me. Also, you instilled in me, by example, a self confidence (or, at least a stubbornness) that has served me well.
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People have asked me what I am going to do with this monster now that it is complete. One copy is going on my bookshelf to collect dust. One copy is going to my father, who first introduced me to geology, and is the only other person I know who took as long to complete his Master's degree as I did. The final copy is going to be soaked in lighter fluid and used to start a barbeque in celebration of its completion. Burgers anyone!?!
CHAPTER 1

INTRODUCTION

The Precambrian-Cambrian boundary represents an extraordinary time in earth history. Profound biologic, tectonic, climatic, and chemical changes combined to produce a metazoan radiation event that radically changed the earth's ecological and sedimentary systems (Knoll, 1991; Brasier, 1992a, 1992b, 1992c; Cowie, 1992; Knoll and Walter, 1992). In order to truly understand these events, we must first recognize their signatures in the rock record and their temporal relationships. Definition of the Precambrian-Cambrian boundary has been at the center of a geologic debate for many years, and only recently has a stratotype section been defined (Landing, 1992) and internationally approved (Landing, 1994). Proposed criteria for boundary identification include stratigraphic, paleontologic, and geochemical data. All of these criteria, however, are rarely present in any one locality that contains the boundary. Nonetheless, it is necessary to apply all of these techniques, where possible, if global correlation of events recorded in the rocks near this boundary is to take place.

The Deep Spring Formation is a mixed carbonate-siliciclastic unit that crops out in the southern Basin and Range Province, along the California and Nevada
border (Figure 1). It may represent late-rift or early-drift deposition (Levy and Christie-Blick, 1991) on the Cordilleran margin during late Neoproterozoic and early Cambrian time. In this study, a sequence-stratigraphic analysis of the Deep Spring Formation revealed three sequence boundaries. These sequence boundaries provide information useful in regional correlation of these pre-trilobite-bearing rocks. Together with new paleontological findings reported herein, they may have global stratigraphic implications for recognition of physical and biological events recorded in terminal Proterozoic strata. In addition, this analysis has provided new insight into the response of a mixed carbonate-siliciclastic system to changes in relative sea level.

Although many studies include sequence-stratigraphic analyses of pure carbonate (see papers in Crevello and others, 1989; Wilgus and others, 1989; Tucker and others, 1990) and pure siliciclastic systems (Posamentier and Vail, 1988; Van Wagoner and others, 1988; Walker, 1990), relatively few have been concerned with the sequence stratigraphy of mixed systems (Mount, 1984; Dolan, 1989; Yose and Heller, 1989; Acker and Stearn, 1990). The present study of the Deep Spring Formation provides a model that may be used for comparison in future studies of mixed carbonate-siliciclastic systems. The model developed in this study of the Deep Spring Formation elucidates a depositional features that is proposed to be unique to mixed systems: the occurrence of minor carbonate slump deposits that signal the base of the Highstand Systems Tract, refered to as early highstand slumps.
Figure 1. Location map of measured sections (*) of the lower member of the Deep Spring Formation. MD-Mt. Dunfee; MM-Magruder Mountain; HR-Hines Ridge; LR-Loretta Road.
During this study, the small shelly fossil *Cloudina* was recognized for the first time in the lower member of the Deep Spring Formation. The world-wide occurrence of this Neoproterozoic fossil makes it an excellent candidate for global correlations (Grant, 1990). The co-occurrence of *Cloudina* with a small shelly fauna previously defined as Cambrian requires revision of the accepted Precambrian-Cambrian biostratigraphy.

Although a stratotype section for the Precambrian-Cambrian boundary has been selected in Newfoundland (Landing, 1992), several problems still exist. Correlation from the stratotype section to temporally equivalent strata globally is complicated by the general sparsity of paleontological data, the predominance of siliciclastic rocks, and poor chemostratigraphic results from the Newfoundland succession. If global correlation of the Precambrian-Cambrian boundary is to take place, it is critical to recognize reference sections that include useful sedimentologic, paleontologic, and geochemical data that can be tied back to the stratotype section. Sequence-stratigraphic analysis and the revised biostratigraphy of the Deep Spring Formation, when combined with chemostratigraphy of Corsetti (1993) and Corsetti and Kaufman (1994), provides valuable information that may be useful for correlation of the California-Nevada sections to the stratotype.

This thesis focuses on the depositional history of the lower Deep Spring Formation. After a review of existing sequence-stratigraphic concepts and controls, new ideas are proposed that may be useful in understanding mixed carbonate-siliciclastic systems. Using this sequence-stratigraphic framework, the lithofacies
of the lower member of the Deep Spring Formation are described, and their depositional environments are interpreted in order to present a model for deposition. This model is then combined with the new information regarding the fauna of the Deep Spring Formation to define a higher resolution stratigraphy for terminal Proterozoic strata in the western United States. Finally, the implications of these new findings are discussed in regard to global correlation and Precambrian-Cambrian boundary issues.

**Geologic Setting**

**Tectonic History**

The Cordilleran passive margin of North America was initiated during the late Proterozoic and continued through Devonian time (Armin and Mayer, 1983; Levy and Christie-Blick, 1989; 1991). It has long been argued that a series of rifting events separated one or more continents or microcontinents from the western margin of North America during the late Proterozoic (Stewart, 1972, 1976; Sears and Price, 1978; Bond and Kominz, 1984; Bond and others, 1983; Bond and others, 1985; Hoffman, 1991; Moore, 1991; Dalziel, 1991; Dalziel and others, 1994). Work by Hoffman (1991) and Moore (1991) suggested that the Australia-Antarctic shield began rifting away from the western edge of Laurentia as early as 1200 million years ago. The western margin of Laurentia underwent at least one additional phase of rifting, the timing of which is poorly constrained.
Subsidence analyses of stratigraphic successions in the western United States suggest the edge of the continent in that area began drifting near the end of the Proterozoic (Armin and Mayer, 1983; Bond and others, 1983), or Early Cambrian (Levy and Christie-Blick, 1991), forming the Cordilleran miogeocline: a passive margin similar to that found on the modern Atlantic coast of North America. The subsidence analysis presumed the Neoproterozoic-Cambrian boundary was about 560-570 million years ago (Ma) (Bond and others, 1985; Levy and Christie-Blick, 1991), but recent studies show the boundary nearer to 544 Ma (Bowring and others, 1993). This change in radiometric age does not significantly effect the analysis (Levy and Christie-Blick, 1991).

Regardless of the radiometric age assigned to the Proterozoic-Cambrian boundary, field evidence from the Death Valley region suggests rifting did not begin until the stratigraphic level of the upper Kingston Peak Formation (Walker and others, 1986; Heaman and Grotzinger, 1992). The onset of drift deposition in the Death Valley region corresponds with either the Stirling Quartzite or perhaps the upper Wood Canyon Formation (Levy and Christie-Blick, 1991). Therefore, the Deep Spring Formation, which is believed to be equivalent to the lower Wood Canyon Formation (Fedó and Cooper, 1990), was deposited either near the end of the rift-drift transition or the onset of drift. This suggests that tectonic activity, such as block faulting, could account for some of the relative sea-level changes recognized in this study.
Rates of subsidence on the Cordilleran passive margin were relatively rapid in the late Proterozoic. Levy and Christie-Blick (1991) assumed 250 to 350 m/my$^6$ in their tectonic subsidence analysis of the southern Great Basin. However, these rates were based on an assumed Proterozoic-Cambrian boundary age of 560 Ma. The new age of 544 Ma, presented by Bowring and others (1993), would result in an even greater rate of subsidence. These rates decreased exponentially through the early part of the Devonian until crustal shortening began in the Late Devonian (Armin and Mayer, 1983; Levy and Christie-Blick, 1989; 1991). Thus, a westward-thickening wedge of sediment was deposited across western Utah and Nevada towards California from the late Proterozoic through the Devonian, with the rate of deposition decreasing through time (Stewart and Poole, 1974). Classically, this wedge was believed to have been deposited uniformly across the broad continental shelf (Stewart, 1972; Stewart and Poole, 1974; Stewart, 1976). Recent work, however, has revealed that east-west trending faults cut the continental shelf and controlled the distribution of sediment in the Middle Cambrian (Kepper, 1981; Rees, 1986) and the Early Silurian (Hurst and others, 1985). The origin of these faults and whether or not they were fundamental crustal structures that may have been active in the Precambrian is still unknown. The recognition in this study, however, of possible fault activity in the lower Deep Spring suggests that they may have been. Wright and others (1976), suggested that a fault-bounded basin, the Amargosa "aulacogen", was active during the Late Proterozoic in the
southern Basin and Range. The presence and orientation of this basin is still being debated, and it may have been oriented parallel to the margin (Levy and Christie-Blick, 1991; Heaman and Grotzinger, 1992). Thus, if faults across the passive margin were active or if rifting were continuing during deposition of the Deep Spring Formation the configuration of the local depositional margin may have been complex and fault movement may account for relative sea-level changes.

Since the Devonian, sediment deposited on the Cordilleran passive margin has undergone numerous periods of contractional and extensional deformation that complicate palinspastic reconstruction of the area. Levy and Christie-Blick (1989) have provided the most widely accepted reconstruction to date of the Basin and Range Province; their reconstruction will be used in this thesis.

Stratigraphy

The Deep Spring Formation crops out in eastern California and western Nevada (Figure 1) and maintains a generally uniform thickness of 500-550 meters over the entire area. It overlies the Reed Dolomite and is overlain by the Campito Formation (Signor and others, 1987). Kirk (1918, in Nelson, 1962) divided the Deep Spring into three members. The lower member, on which this study focuses, is composed mainly of limestone with minor dolomite, and quartzitic sandstone that contains varying amounts of lime mud. The percentage of carbonate increases northward as the thickness of the member decreases.
slightly (Albers and Stewart, 1972). The middle member of the formation contains quartzite overlain by limestone, and is less carbonate-rich than the lower member (Albers and Stewart, 1972). The upper member is composed of a basal dark quartzite overlain by a massive dolomite. Although the formation is lithologically variable, this three-member division is distinguishable throughout the outcrop areas by the quartzite units at the base of the middle and upper members (Albers and Stewart, 1972).

The majority of early regional lithologic correlations and interpretations were developed during the 1960's and 1970's (Nelson, 1962; Albers and Stewart, 1962; 1972; McKee and Moiola, 1962; Stewart, 1972; Stewart and Poole, 1974; Stewart and Suczek, 1977). More recently, Mount and Signor (1989) have completed studies on the sedimentology, stratigraphy and paleontology of the Lower Cambrian strata in this area, which indicate the Deep Spring Formation is Proterozoic to Cambrian in age. As will be discussed in greater detail in Chapter 4 in this thesis, the lower Deep Spring Formation is Neoproterozoic in age, based on the occurrence of *Cloudina*. Although *Cloudina* has not been recognized in the rocks of the Death Valley region south of the study area, Ediacaran fauna were identified in the lower member of the Wood Canyon Formation (Horodisky and others, 1994). *Cloudina* has been found closely associated with Ediacaran fauna in Brazil, Namibia, and China (Grant, 1990), suggesting that the Deep Spring Formation and the Wood Canyon Formation are biostratigraphically correlative. Lithologic correlations with rocks
in Death Valley further suggest that the Deep Spring is probably equivalent to the lower member of the Wood Canyon Formation (Fedo and Cooper, 1990), which represents nearshore marine deposits of dominantly siliciclastic and lesser carbonate sediments (Prave and others, 1991).

Although many studies have involved the Deep Spring Formation, few have concentrated on the lower member. Extensive Mesozoic and Cenozoic faulting and alteration by volcanic activity in its outcrop area have resulted in a limited number of complete sections of the lower member. Mount and Rowland (1981) described the entire formation as a shallowing-upward sequence, representing a peritidal carbonate bank, that is capped by a subaerial erosional surface. Several workers (Gevirtzman, 1983; Greene, 1982; Dienger, 1983; Gevirtzman and Mount, 1986) interpreted parts of the formation as a shallow-shelf deposit, which is consistent with the classic, although possibly erroneous, regional interpretation of a passive-margin setting (cf. Stewart and Poole, 1974). This study builds on this previous work and specifically addresses the lower member of the formation.

Paleogeography

Stratigraphic sections measured in this study are located at Mt. Dunfee (MD) on the outskirts of Goldpoint, NV; Magruder Mountain (MM) near the California-Nevada border; and Hines Ridge (HR) and Loretta Road (LR) located in the Inyo Mountains outside of Bishop, California (Figure 1). Exact locations are provided on topographic maps in Appendix A. The palinspastic
reconstruction of the Basin and Range by Levy and Christie-Blick (1989) provides the best available information on the original geographic position of these mountain ranges (Figure 2). Palinspastically restored ranges attain a more north-south orientation than is presently the case. Although this reconstruction was utilized during model development and in the stratigraphic cross-section and block diagrams, all directional references in the text are to present day settings.

In addition, the locations of palinspastically restored mountain ranges do not entirely agree with the sedimentological findings of this study. Measured sections tend to display shallower-water features in the Inyo Mountain sections at Hines Ridge and Loretta Road, deeper-water settings at Magruder Mountain, and the deepest-water features at Mt. Dunfee. However, the palinspastic reconstruction places Magruder Mountain further outboard than Mt. Dunfee (Figure 3) (Levy and Christie-Blick, 1989). The sedimentological findings of this study suggest that sedimentology provides a more refined placement than the limited structural controls available.

The palinspastically restored mountain ranges also suggest the orientation of the western margin of North America during the time of deposition of the Deep Spring Formation was not entirely north-south as it is today. The deepening-to-the-north trend indicated by this study suggests a more east-west
Figure 2. Palinspastic reconstruction of the mountain ranges of the Basin and Range Province from Levy and Christie-Blick (1989). MD-Mt. Dunfee; MM-Magruder Mountain; HR-Hines Ridge; LR-Loretta Road.
Figure 3. Palinspastic reconstruction of the mountain ranges from this study showing proposed margin orientation (depositional strike). Sedimentary structures suggest shallowest-water deposition at Hines Ridge (HR) and Loretta Road (LR), deeper-water deposition at Magruder Mountain (MM) and deepest-water deposition at Mount Dunfee (MD). Reconstruction from Levy and Christie-Blick (1989). Note that the structurally restored location of Magruder Mountain does not agree with the sedimentologic implications of this study.
orientation to the margin in the southern Basin and Range (Figure 3). A similar margin orientation is suggested by the study of the Wood Canyon Formation (Fedo and Cooper, 1990), which indicates thinner, cratonic deposits to the south and thicker, basinal deposits to the north. This proposed margin orientation is also similar to the orientation of the southern end of the $^{67}\text{Sr}/^{86}\text{Sr} = 0.706$ isopleth (Figure 2), which is believed to represent the westernmost limit of Precambrian basement rocks (Levy and Christie-Blick, 1989). The Middle Cambrian fault mentioned previously (Rees, 1986) is also oriented roughly parallel to the $^{67}\text{Sr}/^{86}\text{Sr} = 0.706$ isopleth in southern Nevada and to the proposed depositional strike of the margin. Thus, deepening to the northwest is not a unique feature to the Deep Spring Formation and may represent a regional bend in the western margin of the United States prior to the Devonian.

Methods

Four stratigraphic sections of the lower member of the Deep Spring Formation were measured using a Jacob staff, and they were described and sampled in detail. Orientation of sedimentary structures indicative of paleocurrent directions were measured using a Brunton compass. These data were sparse and given the extent of post-depositional faulting they could not be correlated between sections. Therefore, no rose diagrams are presented because of the poor quality of the data. Rock specimens were cut and polished, and thin sections prepared for descriptive analysis. Thin section analysis included descriptions of the carbonate and terrigenous grains, matrix, and

Photographs throughout the text are labeled with a distinctive location code. This code describes the section location, year it was measured, section designation and height in meters above the base of the section. For example, MD91I24 is read as follows: Mt. Dunfee, measured in 1991, section I, 24 m above the base of the section. Abbreviations for section locations are as follows: HR is Hines Ridge; LR is Loretta Road; MM is Magruder Mountain; and MD is Mt. Dunfee. Locations for each measured section are in Appendix I.
CHAPTER 2

SEQUENCE-STRATIGRAPHIC CONCEPTS AND CONTROLS

Controls on Deposition

Carbonate, siliciclastic, and mixed depositional systems each respond to changing environmental conditions and exhibit sedimentological features that reflect those changing conditions. Three main factors that must be considered in any sequence-stratigraphic analysis are subsidence, eustasy, and rate of sedimentation. These factors are intricately interrelated with one another (Heller and others, 1993), and this study, as discussed in the following chapters, suggests that this is especially true in mixed carbonate-siliciclastic systems. Locally, other factors such as climate, drainage patterns, tectonism, and oceanic conditions may also leave an imprint on marine sedimentation and thus on the sequence stratigraphy (Suttner and others, 1981; Mack, 1984; Suttner and Dutta, 1986; Nelson, 1988; Read, 1989; Cecil, 1990; Fulthorpe, 1991).

Subsidence

The subsidence history of the Cordilleran miogeocline has been modeled and subsidence histories developed by "backstripping" the thickness of
sedimentary rocks. This method quantitatively removes the subsidence produced by nontectonic processes, such as sediment loading (Armin and Mayer, 1983; Bond and others, 1983; Bond and Kominz, 1984; Bond and others, 1985; Levy and Christie-Blick, 1991). The graph of tectonic subsidence versus time yields a tectonic subsidence curve that can then be compared to similarly modeled subsidence for other passive margins. The shape of the curve may be explained by thermal models such as McKenzie's (1978) uniform stretching model. This graph is then used to interpret the thermal component of subsidence and its decay over time (Allen and Allen, 1990). Subsidence curves for the Cordilleran miogeoclinal are exponential in form (Levy and Christie-Blick, 1991) (Figure 4). These data suggest that after rifting near the end of the Proterozoic or even earliest Cambrian time, cooling of the lithosphere was the main cause of subsidence in the western Cordillera. As discussed in Chapter 1, the data also indicated that the rate of subsidence was very rapid near the Precambrian-Cambrian boundary (Levy and Christie-Blick, 1991), now thought to be about 544 Ma (Bowring and others, 1993).

Eustasy

Eustasy is a very complicated consideration in sequence-stratigraphic analysis. It involves cyclic changes of absolute global sea level. The interpretation of eustasy, however, is generally based on local evidence which reflects relative sea-level change, that is compared and contrasted between
Figure 4. Subsidence curves for the Cordilleran miogeoclone from Levy and Christie-Blick (1991). The Inyo Mountain Curve considers the sections included in this study. \( \beta \) represent the stretching factor.
continents. Vail and others (1977) described cycles of global sea-level change. In their work, Vail and others (1977) recognized various scales or “orders” of sea-level cycles. The larger scale cycles are typically asymmetric, exhibiting gradual rises and rapid falls, whereas the smaller scale cycles exhibit rapid rises in sea level. They recognized first-order cycles that are of long duration, on the order of several hundred million years. Second-order cycles range from 10 to 80 million years in duration. Third-order cycles are typically 1 to 10 million years in duration. Fourth-order cycles typically last less than 1 million years. Embedded in the fourth-order cycles are higher frequency Milankovitch cycles (Hays and others, 1976; Grotzinger, 1986; Mitchum and Van Wagoner, 1991; Vail and others, 1991). Haq and others (1987), working at outcrop scale, suggested that cycles displayed much less severe sea-level falls than did the cycles that Vail and others (1977) recorded from seismic sections. The resultant sea-level curves produced by both groups showed smaller scale sea-level changes superimposed on large-scale sea-level rises and falls; even this picture is probably oversimplified because of the occurrence of even smaller scale sea-level changes associated with fourth-order cycles or smaller.

The causes of changes in sea level remain questionable. First-order cycles of sea-level change appear to be a result of long-term changes in volume of the ocean basins, which may result from changing sea-floor spreading rates (Pitman, 1978). Smaller scale cycles (second-order cycles and smaller) are probably due to tectonic factors (Watts, 1982) or glacio-eustasy (Clark and
others, 1978). Milankovitch cycles are believed to result from variations in the earth's orbital elements (Hays and others, 1976; Mitchum and Van Wagoner, 1991) and are believed to effect fourth- and fifth-order cyclicity (Goldhammer and others, 1987; Mitchum and Van Wagoner, 1991).

Rate of Sedimentation

Rates of sedimentation vary between carbonate and siliciclastic systems. This variability is due in part to in situ carbonate production primarily on platforms versus siliciclastics that are transported to the depositional site. Consequently, carbonate and siliciclastic systems respond differently to sea-level fluctuations, and the resulting changes in accommodation space. It should be noted that climate, tectonism, and latitude may also influence sedimentation rates.

The productivity of carbonate systems reflects both organic and inorganic processes that are dependent on a complex interaction between climate, water depth and temperature, as well as organisms, availability of nutrients and relationship to the photic zone (Wilson, 1975; Tucker and Wright, 1990). Warm waters of sub-tropical to tropical settings with deep water upwelling along platforms provide the best environment for abundant carbonate production. In cooler climates, carbonate production tends to be limited (Nelson, 1988). Rainfall and other climatic conditions, together with platform and basin morphology, play a role in controlling terrigenous sediment dispersal into carbonate-production zones (Sarg, 1988). The organic component of carbonate
production can be adversely effected by murky waters resulting from terrigenous input and positively influenced by abundant nutrients and sunlight. Therefore, carbonate platforms tend to develop best in nutrient-rich, clear, warm waters within the photic zone (Wilson, 1975). As a result, during rising relative sea level, production tends to be high (Dolan, 1989). During lowstands, the width of the carbonate platform is reduced, thus, carbonate production is greatly reduced (Kendall and Schlager, 1981; Dolan, 1989).

Siliciclastic systems, however, are dependent on terrigenous sediment supply, which is in turn dependent on climate and tectonism (Vail and others, 1991; Boggs, 1987); as a result, their response is nearly opposite that of carbonates. In humid environments and in tectonically active areas, siliciclastic sediment supply is high (Boggs, 1987). However, in arid environments and in tectonically stable areas, siliciclastic sediment supply tends to be lower (Boggs, 1987). As noted earlier, during highstands, siliciclastic sediment tends to be trapped nearshore; during falling relative sea level, it is transported seaward (Jervey, 1988; Dolan, 1989). Unlike carbonate systems, that tend to respond directly to available accommodation space in otherwise productive environments, siliciclastic systems respond dominantly to the supply of extrabasinal sediment. Therefore, a lower supply of siliciclastic sediment will result in a low volume of sedimentation or accumulation regardless of available accommodation space.
Mixed carbonate-siliciclastic systems reflect the intricate interplay between both systems, the type of sediment supplied, subsidence and eustasy. This response is best described and interpreted in terms of facies migration patterns. During falling relative sea level, siliciclastic sediment migrating across the shelf displaces carbonate production. However, when relative sea level rises, siliciclastic sediment becomes trapped landward (Jervey, 1988), allowing a suitable environment for carbonate production to thrive (Schlager, 1981; Read and others, 1986; Dolan, 1989).

In a carbonate system, a sedimentation lag often takes place between the time sea level begins to rise and the time when the carbonate factory reestablishes (Schlager, 1981; Read and others, 1986). When carbonate production reestablishes itself, it often initially lags behind but eventually produces sufficient sediment for accumulation rates to match and to exceed the rate of increase in accommodation space (Schlager, 1981; Soreghan and Dickinson, 1994). This production and accumulation pattern is referred to as catch-up deposition. After the carbonate factory has become well established, keep-up deposition takes over, where the rate of carbonate production and accumulation equals or exceeds the rate of sea-level rise, commonly resulting in progradation (Schlager, 1981; Soreghan and Dickinson, 1994).

Sequence-Stratigraphic Concepts

Sequence stratigraphy is the study of repetitive stratigraphic patterns that reflect alternating periods of onlap and offlap and provide information about
sequential changes in sea level. The concept was suggested by Vail and others (1977) as an outgrowth of work on seismic stratigraphy and has since grown into the primary analytic tool used in the description of both carbonate and siliciclastic systems (Posamentier and Vail, 1988; Sarg, 1988; Van Wagoner and others, 1988; Walker, 1990). Although mixed carbonate-siliciclastic systems are well documented, they have had only limited sequence-stratigraphic analysis (Mount, 1984; Dolan, 1989; Yose and Heller, 1989). One reason for this situation may be that the two systems differ drastically in their response to sequence-stratigraphic controls (Dolan, 1989). Studies that attempted such analyses of mixed systems (Yose and Heller, 1989; Srinivasan and Walker, 1993) typically forego the use of sequence-stratigraphic nomenclature because it is often incompatible with the deposits being studied. In this study of the Deep Spring Formation, however, an attempt was made to use traditional sequence-stratigraphic terms where appropriate. Therefore, in this study, the terminology of Sarg (1988) is used during times of carbonate deposition, and that of Vail and others (1977) and Posamentier and Vail (1988) is used for siliciclastic-dominated periods. Generic diagrams displaying a pure siliciclastic and a pure carbonate sequence-stratigraphic framework appears in Figures 5 and 6, respectively. Where existing terminology was insufficient, new terms were defined to describe features unique to mixed systems. One new term, _early highstand slumping_ is proposed for this study to describe a feature interpreted to be unique to mixed systems.
Figure 5. Generalized sequence-stratigraphic block diagram displaying the sequence-stratigraphic framework for a siliciclastic system (Vail, 1987).
SEQUENCE STRATIGRAPHY DEPOSITIONAL MODEL
SHOWING SURFACES, SYSTEMS TRACTS AND LITHOFACIES

Figure 6. Generalized sequence-stratigraphic block diagram displaying the sequence-stratigraphic framework for a carbonate system (Sarg, 1988).

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The following discussion provides an overview of accepted, general sequence-stratigraphic concepts for both pure carbonate and pure siliciclastic systems. General features and responses of mixed systems also are discussed, and are based on this study of the Deep Spring Formation unless otherwise noted. They are presented here to illustrate the similarities and differences among the three types of depositional systems.

As defined by Vail and others (1977), a sequence is a succession of genetically related strata that is bounded above and below by unconformities, known as sequence boundaries, or their correlative conformities. Posamentier and Vail (1988) redefined the original use of the terms Type 1 and Type 2 sequence boundaries, indicating that a Type 1 sequence boundary is marked by an unconformity that records a relative fall of sea level and an abrupt basinward shift in facies, accompanied by fluvial incision, often in the form of an incised valley. A Type 2 sequence boundary is an unconformity that forms in response to decelerating and then accelerating relative sea-level rise. Type 2 sequence boundaries do not display the dramatic evidence of relative sea-level fall as seen in the Type 1 boundary. They do indicate relative sea-level fall, however, and commonly have correlative conformities. Correlative conformities form within basins and are surfaces that are correlative in time with the sequence boundary, but across which there is no depositional hiatus.

Sequence boundaries form because of the sedimentary response to changes in eustasy, subsidence, and sediment supply. Depositional sequences
are bounded above and below by sequence boundaries or their correlative conformities and are subdivided into three systems tracts: lowstand, transgressive, and highstand. Systems tracts are defined by their relative position within a given sequence and by stacking patterns of parasequences and parasequence sets within the systems tract (Van Wagoner and others, 1988). As defined by Van Wagoner (1985), parasequences are conformable, genetically related successions of rock that are bounded by marine flooding surfaces, which are surfaces that display evidence of abrupt rise in relative sea level. Parasequence sets are conformable, genetically related successions of parasequences that are bounded at the base by marine flooding surfaces that are typically more readily apparent in the rock record than the marine flooding surfaces bounding parasequences (Van Wagoner and others, 1988). The tops of parasequence sets are often coincident with either systems tract boundaries or sequence boundaries (Van Wagoner and others, 1988).

Carbonate parasequences are often composed of upward-shallowing meter-scale cycles (Osleger and Read, 1991), therefore, most meter-scale cycles are commonly considered equivalent to parasequences (Mitchum and Van Wagoner, 1991). Carbonate cycles typically rest on a marine flooding surface and shoal to sea level. They may be the result of either internal effects (autocyclic), external effects (allocyclic), or a combination of the two (Osleger and Read, 1991). The effect of autocyclic events within meter-scale cycles can leave a record different from that which would be predicted by looking at the
impact of allocyclic effects only. Therefore, it is often difficult to sort out the record of relative sea-level change preserved in parasequences.

Most of the recognizable meter-scale cycles within the lower Deep Spring Formation contain a subtidal deposit that shoals upward to a peritidal cap. Many of these parasequences display a typical succession of bedding that is associated with a change in relative sea level, such as a gradual thickening of beds upsection during a sea-level rise. Some of the parasequences, however, do not progress as predicted by carbonate models. These irregularities in bed thickness and stacking patterns are attributed to autocyclic fluctuations that have been recorded in the rock record but are superimposed upon allocyclic events (cf. Osleger and Read, 1991). Thus, parasequence sets are sometimes used instead of parasequences to decipher relative sea-level change, and ultimately systems tracts designations.

The lowermost system tract in a Type 1 sequence is the lowstand systems tract (LST), which is deposited at the time of falling relative sea level through the sea level minimum. In Type 2 sequences, in both siliciclastic and carbonate systems, the deposits that accumulate during lowstand are referred to as a shelf margin systems tract (SMST) (Sarg, 1988; Vail and others, 1991). The lower boundary of the LST or SMST is marked by a sequence boundary indicated by either an exposure surface or the base of the correlative lowstand fan or wedge (Vail and others, 1991). In carbonate systems, this debris wedge may be composed of sediment that is shed off the exposed shelf (allochthonous debris
wedge), or the upper slope (autochthonous debris wedge) (Sarg, 1988). In siliciclastic systems, the submarine fan is composed of terrigenous sediment that bypassed the shelf during relative sea-level lowstands (Jervey, 1988; Posamentier and Vail, 1988) and of collapsed and redeposited slope sediments (Vail and others, 1991).

The upper boundary of the lowstand systems tract is the first significant marine flooding surface, called the transgressive surface, which is the surface the first sediment is deposited on as sea level begins to rise (Vail and others, 1991). The transgressive systems tract (TST) directly overlies this surface. The TST is in turn capped by another significant marine flooding surface, called the maximum flooding surface, which denotes the beginning of the highstand systems tract. In a siliciclastic system, the first transgressive deposits are typically sands that migrate back across the formerly exposed shelf as the shoreline moves toward the craton (Posamentier and Vail, 1988). As a result of carbonate lag-times (Schlager, 1981; Read and others, 1986), transgressive systems tracts in carbonate systems may be very thin to non-existent, in which case the maximum flooding surface may be coincident with either the sequence boundary or the transgressive surface.

Based on this study of the Deep Spring Formation, an interpretation has been made as to the features present during the LST and TST of a mixed carbonate and siliciclastic system. The LST in the lower Deep Spring Formation is marked at its base by an exposure surface and incised-valley-fill deposit. In
the deeper parts of the shelf, a lowstand-prograding wedge was deposited during the LST. Shelf sands of the TST rest directly on the incised-valley-fill deposit. The surface between the incised-valley-fill deposit and the transgressive sands is the transgressive surface. The transgressive sands may have been fluvial deposits that were trapped in topographic lows formed during the time of exposure and then reworked on the shelf as relative sea level rose during the transgression. As carbonate sedimentation began during the TST, minor fluvial input continued, thus the fluvial siliciclastic sediments were interbedded with shallow-water carbonates. These transgressive deposits are capped by the maximum flooding surface. Also during transgression, in deeper parts of the shelf, a depositional lag took place, forming a starved shelf basin that resulted in substantial relief between the shallow and deep shelf. The duration of the depositional lag was probably longer in the deeper areas of the shelf than the shallower areas, due to the time it took for carbonate production to migrate from the shallow shelf to the deeper shelf. Therefore, in the deeper shelf, no deposition is recorded during the transgressive systems tract, thus, the transgressive surface is coincident with the maximum flooding surface.

In this study of the Deep Spring Formation, no distinction between Type 1 or Type 2 sequence boundaries was made. Although the recognition of a karst surface in Sequence Boundary A suggests a Type 1 sequence boundary, for the most part, the limited number of outcrops of the lower Deep Spring Formation and the poor preservation of distinctive features in those outcrops that were
available made it difficult to determine how dramatic the relative sea-level falls were during the time of Deep Spring Formation deposition. The distinction between Type 1 and Type 2 sequence boundaries is not critical to interpretations presented in this study, as this study does not attempt to determine the magnitude of relative sea-level rise and fall. Thus, sequence boundaries are discussed with no reference to type distinction, and Type 1 terminology is used throughout this report.

In general sequence-stratigraphic models, transgressive systems tracts deposits are overlain directly by highstand systems tracts (HST) deposits. In carbonate systems, the highstand is marked initially by early catch-up deposition (Soreghan and Dickinson, 1994). Thus, parasequences reflect deposition in progressively shallower waters even though relative sea level is rising. During the late highstand systems tract, keep-up deposition takes place on the shelf (Soreghan and Dickinson, 1994). It is during late highstand time that pure carbonate systems produce enough sediment to result in progradation and the development of relief between the platform and the basin. Consequently, sediment slumps off the shelf into the basin during late highstand. The abundant shedding of sediment from the shelf during highstand is typical of carbonate systems (Droxler and Schlager, 1985; Dolan, 1989; Mullins, 1983). Conversely, coarse siliciclastic sediments on passive margins tend to be trapped landward during sea-level highstands (Jervey, 1988; Vail and others, 1991).
Thus, only thin, fine-grained deposits, which are referred to as condensed sections, develop on the outer shelf and basins (Loutit and others, 1988).

Mixed systems are seemingly more complicated than either pure carbonate or pure siliciclastic systems. Rocks from the Deep Spring Formation provide evidence that during highstands of relative sea level, siliciclastic deposition became trapped landward, allowing carbonate production to reestablish in areas formerly dominated by siliciclastic sedimentation. Catch-up carbonate sedimentation on the shelf, combined with the significant relief produced during the transgressive system tract resulted in Early Highstand Slumping into the basin: a feature interpreted in this study as unique to mixed systems. These mixed-system slumps differ from carbonate-system slumps because they formed during early highstand catch-up deposition instead of late highstand keep-up deposition.

In general sequence-stratigraphic models, highstand systems tracts are overlain by a sequence boundary or correlative conformity. The overlying lowstand systems tracts of the next depositional sequence are initiated as relative sea level continues to fall (Vail and others, 1991). The sequence boundary is marked by the fall of relative sea level and is characterized by a basinward shift in facies. In pure carbonate systems, the locus of carbonate sedimentation migrates basinward until the platform is eventually exposed and the area of carbonate production is greatly diminished (Kendall and Schlager,
1981; Dolan, 1989). In pure siliciclastic systems, the sands again bypass the shelf forming the lowstand systems tract (Vail and others, 1991).

Features from both pure siliciclastic and pure carbonate systems that are associated with the formation of sequence boundaries are recognized in the mixed system of the lower Deep Spring Formation. These features, however, are the result of a local tectonic event that impacted relative sea-level change and not eustatic events. This local tectonic event will be discussed in more detail in Chapter 3.
CHAPTER 3

DEPOSITIONAL ENVIRONMENTS AND SEQUENCE STRATIGRAPHY

This section describes the lithofacies of the lower Deep Spring Formation, interprets their depositional environments, and places the facies migration patterns within a sequence-stratigraphic framework. Table 1 presents a summary of lithofacies of the lower Deep Spring Formation, including general descriptions of the lithofacies and interpretations of the depositional environments. Correlation of facies among measured sections was complicated because outcrops of the lower Deep Spring Formation are located in a tectonically extended terrane and in some areas have undergone greenschist-grade contact metamorphism. The presence of unconformities and syndepositionally deformed beds provided stratigraphic markers on which many lithologic correlations were based. Nevertheless, many correlations, especially between the shallow-water and deep-water facies, are inevitably model dependant. The facies stacking patterns, when interpreted in a sequence-stratigraphic framework, reveal three sequence boundaries within the lower Deep Spring Formation (Figure 7 and Figure 8). Facies patterns also suggest
Table 1. Summary of lithofacies features.

<table>
<thead>
<tr>
<th>Lithofacies</th>
<th>Color and Bed Thickness</th>
<th>Rock Types</th>
<th>Constituents</th>
<th>Diagnostic Sedimentary Features</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>I. Limestone conglomerate</td>
<td>Red weathering matrix, light-gray clasts, 0-0.6 m thick</td>
<td>Intraformational limestone conglomerate</td>
<td>Parallel-laminated limestone clasts, mixed lime-mud and quartz-silt matrix</td>
<td>Channel-like morphology; no apparent clast imbrication; fills dissolution cavities</td>
<td>Incised valley fill</td>
</tr>
<tr>
<td>II. Parallel-laminated sandstone</td>
<td>Red-brown; 0-10 m thick</td>
<td>Quartz arenite, shale, siltstone; wackestone and packstone at HR and LR; only quartz arenite at MM</td>
<td>Medium-grained quartz with overgrowths, quartz silts and muds, neomorphosed carbonate cement</td>
<td>Parallel laminae, low angle x-bedding, ball-and-pillow structures, parting lineations, ripples</td>
<td>Nearshore to shallow ramp</td>
</tr>
<tr>
<td>III. Clotted and intraclastic limestone</td>
<td>Medium to light gray, light-gray intraclasts; 15-60 m thick</td>
<td>Lime mudstone; intraclastic wackestone; sandstone at LR; siltstone at LR and HR</td>
<td>Intraclasts and peloids in lime mud</td>
<td>Parallel laminae often occurring with x-lamina; structureless peloidal (clotted) beds</td>
<td>Tidal flat to subtidal lagoon</td>
</tr>
<tr>
<td>IV. Bioclastic limestone</td>
<td>Light to dark gray, dark-gray bioclasts; 0-15 m thick</td>
<td>Bioclastic, peloidal grainstone; bioclastic wackestone</td>
<td>Small shelly fossils in lime mud; small shelly fossils and peloids, hematite and glauconite replacement at top of lithofacies</td>
<td>Internal homogeneity replaces parallel laminae; beds thicken upsection; bioclastic-rich lenses</td>
<td>Innershelf to upper ramp</td>
</tr>
<tr>
<td>V. Cross-bedded sandstone</td>
<td>Red-brown to gray; 5-17 m thick</td>
<td>Quartz arenite; interbedded siltstone and shale at MD and HR; interbedded lime mudstone at LR</td>
<td>Very coarse- to fine-grained quartz with overgrowths, neomorphosed carbonate cement</td>
<td>Tabular cross-bedding; bedding and cross-bedding sets thicken upsection parallel-laminated sandstone at HR; hummocks and loading structures at MD</td>
<td>Middle to innershelf</td>
</tr>
<tr>
<td>VI. Ooid limestone</td>
<td>Light gray; 0-20 m thick</td>
<td>Oolitic grainstone</td>
<td>Recrystallized carbonate ooids</td>
<td>Relict large-scale cross-bedding</td>
<td>Ooid Shoal</td>
</tr>
</tbody>
</table>
Table 1. (Continued) Summary of lithofacies features.

<table>
<thead>
<tr>
<th>Lithofacies</th>
<th>Color and bed Thickness</th>
<th>Rock Types</th>
<th>Constituents</th>
<th>Diagnostic Sedimentary Features</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>VII. Intraclastic limestone</td>
<td>Medium to light gray; 0-30 m thick</td>
<td>Wackestone to packstone</td>
<td>Intralasts in lime mud</td>
<td>Thick-bedded, coarsening of grains upsection, no diagnostic sedimentary structures visible</td>
<td>Below to within storm wave base</td>
</tr>
<tr>
<td>VIII. Contorted limestone</td>
<td>Dark gray; 0-7 m thick</td>
<td>Lime mudstone to packstone</td>
<td>Lime mud, minor intralasts</td>
<td>Highly to slightly contorted bedding, ball- and pillow structures</td>
<td>Ramp below storm wave base</td>
</tr>
<tr>
<td>IX. Shale and siltstone</td>
<td>Red-brown siltstone and dark brown shale; 0-1 m thick</td>
<td>Siltstone and shale</td>
<td>Quartz silt and mud, glauconite and hematite at base</td>
<td>Parallel laminae, horizontal traces and resting marks</td>
<td>Below storm wave base</td>
</tr>
<tr>
<td>X. Contorted limestone-sandstone</td>
<td>Red-brown siltstone and gray limestone; 1-3 m thick</td>
<td>Sandstone, siltstone, and lime mudstone</td>
<td>Very fine-grained to silt-sized quartz and lime mud</td>
<td>Highly-contorted bedding, boudinage structures</td>
<td>Ramp below storm wave base</td>
</tr>
<tr>
<td>XI. Dolomitized allochem conglomerate</td>
<td>Buff to orange, dark-gray allochems, green-black shale; 0-25 m thick</td>
<td>Packstone, siltstone, shale, and dolomite</td>
<td>Dolomite, lime mud and intralastic clasts</td>
<td>Coarsening of grains and thickening of beds upsection; shale rare upsection</td>
<td>Turbidites below to possibly within storm wave base</td>
</tr>
</tbody>
</table>

HR-Hines Ridge; LR-Loretta Road; MD-Mt. Dunfee; MM-Magruder Mountain
Figure 7. Stratigraphic columns showing the rock type and lithofacies present in the Lower Deep Spring Formation. Lithofacies correspond to those described in Table 1. Patterns used above are the same used in all tables within Chapter 3. LR-Loretta Road, HR-Hines Ridge, MM-Magruder Mountain, MD-Mount Dunfee.

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Southwest

Northeast

100 m

LR HR MM MD

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Figure 8. Stratigraphic columns showing the sequence-stratigraphic framework of the Lower Deep Spring Formation including systems tracts and bounding surfaces. Colors correspond to the block diagrams that appear throughout the text. LR-Loretta Road, HR-Hines Ridge, MM-Magruder Mountain, MD-Mount Dunfee.
the lower Deep Spring Formation represents deposition on a ramp to distally steepened ramp (sensu Read, 1985).

In the Deep Spring Formation, the combined effects of rapid subsidence, third- and fourth-order eustatic events, and variable sedimentation rates resulted in abundant accommodation space. The rate of subsidence was thought to be initially rapid on the Cordilleran continental margin, but decreased exponentially through the early Paleozoic (Levy and Christie-Blick, 1991). However, as will be discussed in this chapter, faulting may have played a major role in the development of Sequence Boundary B. Faulting during development of Sequence Boundary B suggests that the time of deposition of the lower Deep Spring Formation was probably closer to the rift-drift transition than to the onset of drift.

Other changes in the rate of subsidence resulting from local tectonic events could not be resolved, although they might be present. Similarly, tectonic versus eustatic control on relative sea-level changes could not be elsewhere independently distinguished. Thus, relative sea level is explicitly used or implied throughout the following discussion. Relative sea-level change is displayed as an over-simplified sinusoidal curve on figures in the text. Sea-level changes were probably not symmetric nor of equal magnitude, as shown by the sea-level curves produced by Watts (1982) and Haq and others (1987). However, use of a sinusoidal curve to represent sea-level change is an accepted practice to simplify geologic responses to sea-level change (sensu Posamentier and Vail,
1988: Posamentier and others, 1988; Sarg, 1988). The effects of higher frequency sea-level change are recorded in the parasequences and parasequence sets that comprise the depositional sequences of the lower Deep Spring Formation (Tables 2 through 7).

The thicknesses of systems tracts in the lower Deep Spring Formation are on the order of tens of meters, which suggests third-order or even fourth-order relative sea-level cycles (Vail and others, 1977; Posamentier and others, 1988). Although parasequences and parasequence sets are recognizable in the lower Deep Spring Formation, diagenesis and discontinuous exposure make logging them difficult. The sequences recognized in the lower Deep Spring Formation were superimposed on a first-order sea-level rise that took place during the terminal Proterozoic and early Cambrian (Vail and others, 1977; Sloss, 1979). The first-order sea-level rise was probably related to the breakup of Laurentia and the subsequent reduction of volume in the ocean basin due to development of a new spreading center associated with rifting (Hays and Pitman, 1973; Bond and others, 1984).

Rates of sedimentation are interpreted to have been highly variable during deposition of the lower Deep Spring Formation. Based on previous studies and models of Neoproterozoic rocks throughout the Basin and Range (Cowie, 1971; Scotese and others, 1979), the Deep Spring Formation was deposited in a warm, sub-tropical to tropical setting, probably similar to modern environments of Eastern Mexico or Belize as described by Ward and others.
(1985) and James and Ginsburg (1979), respectively. This setting allowed prolific carbonate production when siliciclastic input was low. However, the siliciclastic sediment supply was periodically abundant, which resulted in a mixed carbonate-siliciclastic depositional system and intermingling carbonate and siliciclastic facies. The siliciclastic sediment source was probably the same as the source that supplied the time-equivalent Wood Canyon Formation, which, in part, as Prave and others (1991) discussed, may represent fluvial deposition.

**Sequence Boundary A**

The contact between the basal Deep Spring Formation and the underlying Reed Dolomite is here interpreted as a sequence boundary, designated Sequence Boundary A (Figures 7, 8, 9, and 10). The best evidence supporting this interpretation is present at the Hines Ridge section, where the top of the Reed Dolomite displays dissolution features (Figures 10a, 10b). This surface is overlain by a conglomerate composed of Reed Dolomite clasts that forms the limestone-conglomerate lithofacies (Lithofacies I Table 1; Figures 10c, 10d).

**Dissolution Surface: Description**

At Hines Ridge, the upper surface of the Reed Dolomite, which directly underlies the Deep Spring Formation, displays an irregular dissolution surface (Figures 10a, 10b, 10d). This surface displays a highly irregular geometry along the top of the Reed Dolomite (Figure 10a), and, locally, thin, elongate cavities.
Figure 9. **Sequence Boundary**: Interpretive block diagram of deposition during development of Sequence Boundary A and the lowstand prograding wedge deposited on the underlying Reed Dolomite in a shelf environment.
Figure 10a. **Sequence Boundary A:** Cross-sectional view of minor dissolution features at the top of the Reed Dolomite (arrow). The red, quartz-rich carbonate sediment filling the features is similar in composition to the matrix of the limestone conglomerate in Figures 10c and 10d. Location HR91101; lens cap is approximately 6 cm in diameter.

Figure 10b. **Sequence Boundary A:** Close-up of irregular pockets at base of "channel" feature with infill of the limestone-conglomerate lithofacies of Incised-Valley-Fill A. Location HR91101; scale in inches, approximately 15 cm long.

Figure 10c. **Incised-Valley-Fill Conglomerate:** Incised-valley-fill conglomerate clasts. Note the parallel laminae in some of the clasts. Location HR91101; bar is approximately 5 cm.

Figure 10d. **Incised-Valley-Fill Conglomerate:** Limestone conglomerate filling "channel" feature at the top of the Reed Dolomite. Note pockets in the irregular base (lower arrow). The conglomerate is overlain by the parallel-laminated sandstone of the Transgressive Systems Tract (upper arrow). Location HR91101; hammer head is approximately 20 cm long.
that extend several centimeters down into the Reed Dolomite are present (Figures 10b, 10d). The connectedness of some cavities to the upper surface of the Reed Dolomite is not visible. The cavities are filled with a red-brown weathering carbonate mudstone containing abundant fine-grained quartz sand and silt (Figure 10b).

Outcrops displaying the irregular surface and associated cavities are limited to rare occurrences at the Hines Ridge section. The most dramatic outcrop face that contains these cavities is located on the eastern flank of Hines Ridge and displays a channel-like morphology (Figure 10d). This face is only a maximum of about 0.5 m high and extends laterally about 6.5 m before it becomes covered. The Reed Dolomite crops out again on the western flank, where the upper surface is highly irregular, but no cavities were found underlying the Reed Dolomite-Deep Spring Formation contact (Figure 10a).

Although difficult to see in outcrop, the sequence boundary is most easily discerned on the eastern flank of Hines Ridge by its association with the overlying limestone-conglomerate lithofacies (Lithofacies 1, Table 1; Figure 10d). The western-flank exposures of the surface display no conglomerate fill. Instead, the surface is overlain by a 2-cm-thick layer composed of the red-brown, carbonate mud and fine-grained quartz sand and silt similar to that filling the cavities and comprising the matrix of the conglomerate (Figure 10a).
Dissolution Surface: Depositional Interpretation

The highly irregular and pocket-like nature of the features at the base of the "channel" suggests they are dissolution features formed during subaerial exposure following deposition of the Reed Dolomite. In other formations, similar irregular features that are filled with siliciclastic sediment have been interpreted as paleokarst (Evans and Hine, 1991; Pelechaty and others, 1991). These irregular features at the base of the channel and the similarity of the fill material to the underlying Reed Dolomite suggest a large dissolution pocket or karst (sensu Pelechaty and others, 1991) and perhaps not a scoured channel.

Dissolution Surface: Sequence-Stratigraphic Interpretation

The dissolution surface reflects subaerial exposure of the carbonate platform, and the development of an unconformity that is interpreted as a sequence boundary. In Figure 8 and elsewhere, this sequence boundary is referred to as Sequence Boundary A. It is directly overlain by the Lowstand Systems Tract in the form of an incised-valley-fill conglomerate and lowstand wedge.

Lowstand Systems Tract A

Incised-Valley-Fill Conglomerate: Description

Directly overlying Sequence Boundary A at Hines Ridge is the limestone-conglomerate lithofacies (Lithofacies I, Table 1; Figure 10c, 10d). Exposure of
this lithofacies is limited only to an outcrop at Hines Ridge. Although a similar lithology is present in float at Magruder Mountain, no intact exposures were found there. The conglomerate at Hines Ridge directly overlies the well-exposed dissolution cavities in the Reed Dolomite described previously (Figure 10d). The thickness of the conglomerate varies from 0.2 m at its western-most outcrop limit to 0.6 m at its eastern outcrop limit: a lateral distance of 6.5 m.

The conglomerate is clast supported and is composed of clasts of Reed Dolomite in a matrix of red-brown-weathering carbonate mud and fine-grained quartz sand and silt (Figures 10c, 10d). The carbonate matrix is the same material that is filling the underlying cavities in the Reed Dolomite. The conglomerate clasts are well rounded to subangular and range in size from granule to boulder (3 cm to 30 cm) (Figure 10c). In available two-dimensional exposures, clasts appear elongate to equant in shape. In limited three-dimensional exposures, however, the subequant clasts are actually elongate with their long axes nearly perpendicular to the outcrop surface. Some of the clasts have very irregular shapes (Figures 10c, 10d). Although most clasts dip less than 10°, many others show considerable dip, and some stand vertically. Because of the lack of good three-dimensional exposure, however, and the variable clast shape, reliable data regarding clast orientation was unobtainable.

The composition of the clasts is similar to that of the underlying Reed Dolomite. The Reed Dolomite, therefore, probably acted as a local source for the conglomerate. Primary parallel lamination is preserved within many of the
clasts (Figure 10c). Those clasts without parallel laminae contain visible clots of ferroan dolomite and ferroan calcite mud with dolomite cement between clots, or they are homogenous with no discernable textures. Some of the clasts contain minor amounts (about 1%) of quartz silt. Typically, clasts have irregular boundaries outlined by dolomite rhombohedra and iron-staining indicating stylotization and dissolution.

The limestone-conglomerate lithofacies is directly overlain by the parallel-laminated-sandstone lithofacies (Figure 10d; Lithofacies II, Table 1). The clasts provide relief at the top of the conglomerate, but they are not truncated by an erosional surface: thus, the contact with the overlying sandstone is irregular but conformable.

Incised-Valley-Fill Conglomerate:
Depositional Interpretation

Due to the limited exposure of this lithofacies, it is difficult to determine the exact origin or depositional mechanism of the "channelized" conglomerate. If the clasts were deposited concurrently with the matrix, then the conglomerate may represent a lag of coarse material deposited in a physiographic low. Because the physiographic low appears to be a paleokarst feature, which must have formed during exposure, the clasts may have been transported fluvially at lowstand or deposited during the onset of transgression. Evans and Hine (1991) attributed similar features to karst-controlled fluvial deposition during subaerial exposure. In their interpretation, fluvial sediments that were deposited during
lowstand were reworked and redistributed during transgression. A similar mechanism may explain the origin of the conglomerates at the base of the Deep Spring Formation. An alternative possibility is that both the clasts and the matrix represent cavern-fill deposits that were subsequently exhumed as the subaerial unconformity continued to develop during lowstand. Examples of similar exhumed paleokarsts are described by Desrochers and James (1988).

Incised-Valley-Fill Conglomerate: Sequence Stratigraphic Interpretation

Regardless of the exact mechanism of conglomerate deposition, the conglomerate rests directly on the subaerially exposed unconformity of Sequence Boundary A. It is therefore considered part of the Lowstand Systems Tract. It is interpreted as an incised-valley-filling conglomerate deposited in the physiographic lows created during exposure of the carbonate platform.

Lowstand Prograding Wedge A

To the palinspastic south of Hines Ridge at Mt. Dunfee (Figures 5, 7 and 8), the contact between the Reed Dolomite and the Deep Spring Formation is interpreted as Sequence Boundary A. Although poorly exposed, the two formations appear conformable, but the basal Deep Spring Formation displays an abrupt basinward shift in facies relative to the upper Reed Dolomite. These basal strata are the intraclastic-limestone lithofacies (Lithofacies VII, Table 1; Figures 7, 8, and 11a), and they are interpreted as a lowstand prograding wedge that accumulated on the shelf (Figure 9).
Figure 11a. **Lowstand Prograding Wedge**: Intraclastic limestone at Mt. Dunfee. Note the small intraclasts that display some iron staining. On a fresh surface the clasts are buff colored. Location MD911129; 6-cm-diameter lens cap.

Figure 11b. **Transgressive Surface**: Cross-sectional view through reworked surface marking reworking at the top of the lowstand prograding wedge (intraclastic-limestone lithofacies) at Mt. Dunfee. This surface is directly overlain by dark-gray, syndepositional slump beds. Upper surface of location MD911130; 6-cm-diameter lens cap.
Lowstand Prograding Wedge: Description

The intraclastic-limestone lithofacies crops out only at the Mt. Dunfee section. It is a 30-meter-thick succession of thick-bedded, coarsening-upward light-gray wackestone to grainstone (Lithofacies VII, Table 1; Table 2: Figure 11a). The lithofacies directly overlies the Reed Dolomite at Mt. Dunfee. Because of extensive faulting, this lowermost part of the Deep Spring Formation crops out only on the northern side of the canyon that is located immediately south of the canyon containing the primary Mt. Dunfee measured section (Appendix A). No sedimentary structures other than thick bedding were observed. Bedding is laterally continuous and ranges in thickness from 0.2 m to 0.65 m.

The limestone intraclasts are often iron-stained on weathered surfaces (Figure 11a) but are off-white on fresh surfaces. The intraclasts coarsen from a maximum of 2 mm at the base to a maximum of 10 mm at the top of the succession. The larger clasts are rare, except in a few beds in the upper meter of the lithofacies where they are common. Additionally, the clasts increase in sphericity upsection.

The uppermost bed of this lithofacies is 35 cm thick. The lower 15 cm of the bed, like all of the underlying beds of the lithofacies (Figure 11a), is a packstone to wackestone and contains abundant lime mud as a matrix. The top 20 cm of the uppermost bed, however, is a laterally continuous, buff to reddish-brown, dolomitized grainstone (Figure 11b). The change from the underlying
packstone and wackestone at the base of the uppermost bed to the grainstone at its top is gradational. The clast size remains fairly consistent throughout the uppermost bed of Lithofacies VII (Table 1), however, the clasts in this bed are generally larger than the clasts in the underlying beds, averaging 7 to 10 mm in size. The clasts are limestone intraclasts, which are similar to those found in the underlying beds of the lithofacies (Lithofacies VII, Table 1; Figure 11a). The clasts display secondary iron staining, but on fresh surfaces they are off-white like the underlying clasts. Aside from the dolomitized grainstone at the top of the lithofacies, the transition to the overlying lithofacies at Mt. Dunfee is easily identified by the change from thick-bedded, light-gray, intraclastic limestone (Lithofacies VII, Table 1) to dark-gray, thin-bedded, highly contorted limestone (Lithofacies VIII, Table 1).

Lowstand Prograding Wedge:
Depositional Interpretation

The light-gray color of the grains suggests derivation from a shallow-water platform (Wilson, 1975, p. 26). The mechanism of deposition, however, is complicated by the lack of observable sedimentary structures. Two possible scenarios may explain the lack of sedimentary structures: basinal deposition by submarine debris flows or shelfal accumulation with post-depositional homogenization of beds by bioturbation. Debris flows are a common mode for transporting sediment off the shelf. Because debris flows are cohesive flows, no internal stratification is formed within them (Cook and Mullins, 1983). Repetitive
successions of submarine debris flows typically result in interbedded turbidite flows; none of which are present in Lithofacies VII (Table 1). Perhaps a more reasonable explanation for the lack of internal structure is homogenization of the beds by bioturbation. The action of burrowing organisms commonly destroys any existing internal structures (Droser and Bottjer, 1986). During the Neoproterozoic, extensive bioturbation was limited to the shallow-subtidal zone, although rare traces are found in deeper water deposits (Fedonkin, 1985).

Regardless of the exact mechanism of deposition, the stratigraphic position and light color of clasts suggests that this lithofacies may represent reworked, partially lithified sediment eroded from the Reed Dolomite and transported into a shallow-shelf basin located just offshore during exposure of the platform. The extent of bioturbation suggests this shallow-shelf basin provided an environment that was still shallow enough to allow the organisms to flourish.

Lowstand Prograding Wedge: Sequence-Stratigraphic Interpretation

Allochthonous wedges of debris shed from the shelf into the basin are commonly associated with exposure surfaces (Sarg, 1988). The intraclastic deposits in the lower Deep Spring Formation are much smaller than the wedges of debris described by Sarg (1988) and the exact geometry of the deposit is uncertain. However, the mode of origin may be similar to that discussed by Sarg (1988) because of its lateral association with Sequence Boundary A at the top of
the Reed Dolomite at the Hines Ridge locality. Allochthonous debris wedges form as relative sea level falls and exposes the shelf. As more of the shelf becomes exposed, carbonate sand eroded from the exposed area is transported off the shelf and into the basin, forming a wedge of sediment. Similar features are also associated with reefs, where debris shed from the reef is deposited in deeper water in the form of a debris wedge (Sellwood, 1981; Franseen, 1988; Pomar, 1991). Thus, debris wedges often have a shallow-water appearance (i.e., intraclastic and light gray in color) because of the shallow-water origin of the lime sands, even though they are actually deeper water deposits.

In the lower Deep Spring Formation, megabreccias are absent, the intraclastic-limestone lithofacies is relatively thin and displays evidence of bioturbation, and at Mt. Dunfee it directly overlies the Reed Dolomite with no evidence of exposure. This suggests that sediment was not transported off the shelf, but instead, was deposited in a shallow-shelf basin seaward of an exposed carbonate platform. As relative sea level continued to fall, the exposed area on the shelf extended further seaward. The increase in intraclast size and clast sphericity upsection may be attributed to increasing proximity to the source, and a higher energy environment that developed as sea level fell.

The top of the intraclastic-limestone lithofacies (Lithofacies VII, Table 1; Figure 11b) displays evidence of a change in either primary or secondary depositional energy levels. The lack of mud in the upper 20-cm-thick grainstone of the intraclastic lithofacies (Lithofacies VII, Table 1) at Mt. Dunfee suggests the
sediment was well washed. This sorting could have resulted from either a primary high-energy depositional regime or a secondary winnowing of the mud. Mud is abundant in all the underlying beds, however, suggesting at least episodic relatively low-energy deposition for the majority of this lithofacies and, as discussed with the lowstand prograding wedge, the interpreted bioturbation suggests a moderately slow rate of deposition. In addition, mud is abundant in the lower 15 cm of the uppermost bed. Mud is absent only from the upper 20 cm of the uppermost bed of this lithofacies and the change from mud-rich to mud-depleted is gradational from the base to the top of the bed. This gradational change suggests that energy levels changed after initial deposition of the uppermost bed of the lithofacies and not before or during deposition. The increase in energy may result from an increase in wave action associated with a lowering of wave base, which suggests a fall of sea level. However, the overlying deposit suggests that sea level rose prior to its deposition, as will be discussed in the "Early Highstand Slumps" section. The upper surface of this bed may represent a depositional hiatus; such surfaces often display evidence of reworking during time of nondeposition (Mullins and Nuemann, 1979; Tucker and Wright, 1990). Therefore, this succession of strata is interpreted to have formed during a time when sea level reached its minimum, thus lowering wave base to winnow the mud from the upper strata. Sea level began rising soon after and this is recorded at Mt. Dunfee as a time of nondeposition prior to the deposition of the deeper early-highstand slumps.
Early Transgressive Systems Tract A

The parallel-laminated-sandstone lithofacies (Lithofacies II, Table 1) directly overlies the incised-valley-fill conglomerate where exposed at the Hines Ridge section, and it directly overlies Sequence Boundary A at Hines Ridge, Loretta Road and Magruder Mountain (Figures 7 and 8). This lithofacies is interpreted as the initial marine inundation of the platform and initial siliciclastic deposition of Transgressive Systems Tract A (Figure 12). Further seaward, at Mt. Dunfee, the lithofacies is not present. At that section, no depositional record of the Transgressive Systems Tract is present, but the transgressive surface is interpreted to directly overlie the reworked upper strata of the lowstand prograding wedge. These strata are capped by a depositional hiatal surface representing the time of deposition of Transgressive Systems Tract A on the shelf.

Transgressive Sandstone Deposit: Description

The parallel-laminated-sandstone lithofacies (Figure 13; Lithofacies II, Table 1) is limited in extent, being represented only at Hines Ridge, Loretta Road, and Magruder Mountain. This lithofacies varies in exposed thickness from approximately 9 m to 2 m. It is composed of sandstone, shale, siltstone and limestone that are variably associated in three subfacies (Table 2): (a) a shale-and-siltstone subfacies; (b) a quartz-arenite subfacies; and, (c) a sandstone-and-limestone subfacies. Not all subfacies are present at all sections.
Figure 12. Transgressive Systems Tract: Interpretive block diagram of deposition during the Transgressive Systems Tract A. During the early part of the transgression, the shoreline sands migrate across the previously exposed shelf but are trapped by the karst topography; carbonate production is very limited during this time as it reestablishes.
Figure 13a. **Transgressive Systems Tract:** Siltstone and shale beds (Lithofacies II, Table 1) of the basal Deep Spring Formation that directly overlie the incised-valley-fill conglomerate at the top of the Reed Dolomite at Hines Ridge. Location lower HR91I02; hammer is approximately 40 cm long.

Figure 13b. **Transgressive Systems Tract:** Parallel-laminated sandstone at Hines Ridge. Location upper HR91I05; bar is approximately 10 cm.

Figure 13c. **Transgressive Systems Tract:** Photomicrograph of thick-section of stained parallel-laminated quartz arenite in cross-polarized light. Staining is for presence of ferroan calcite; pink is calcite, blue is ferroan calcite. Location HR91I05; bar is approximately 1 mm.

Figure 13d. **Transgressive Systems Tract:** Low-angle cross-bedding of the transgressive sandstone at Hines Ridge. Note how some of the beds pinch out. Location middle HR91I03; 40-cm-long hammer.

Figure 13e. **Transgressive Systems Tract:** Ball-and-pillow features in the transgressive sandstone beds at Hines Ridge. These features display a folded internal geometry. Arrow lies parallel to the fold axis of the feature. Location lower HR91I03; bar is approximately 10 cm.

Figure 13f. **Transgressive Systems Tract:** Wave ripples in transgressive sandstone at Hines Ridge (above pencil). Location lower HR91I02; 15-cm-long pencil.
The shale-and-siltstone subfacies (Subfacies IIa, Table 2) crops out only at Hines Ridge where it composes the lower 2.5 m of the parallel-laminated lithofacies. It contains parallel-laminated shale, dark-brown to black siltstone, and minor, thin, red-brown sandstone beds (Figure 13a). The shale constitutes 2/3 of the subfacies and is present in lenticular beds 20 to 50 cm thick. The siltstone beds that compose the remaining 1/3 of the subfacies are lenticular to laterally continuous and 10 to 20 cm thick. The rare sandstone beds are typically irregular and lenticular and are usually 10 cm thick or less. The siltstone beds and rare sandstone beds display abundant internal parallel laminae, common low-angle, tabular cross-stratification, common ball-and-pillow structures, and rare ripples and parting lineations (Figure 13).

Low-angle, planar cross-stratification is preserved as truncated, tabular-to wedge-shaped sets that are continuous for several meters before pinching out laterally (Figure 13d). Sets range in thickness from 10 cm to 50 cm. Thinner sets tend to lie between parallel-laminated sandstone beds, whereas thicker sets are present in co-sets up to a meter thick. The bounding surfaces of the cosets are typically inclined a few degrees. Some of the cross-stratification appears to be bi-directional in nature and indicates a general paleoflow direction to the north, with opposing flow to the south. All other cross-stratification indicates a general northerly paleoflow direction. However, too few paleocurrent indicators were present to provide a good statistical analysis of paleoflow.
TABLE 2. Summary of Early Transgressive Systems Tract A

<table>
<thead>
<tr>
<th>Lithofacies/ Subfacies</th>
<th>Distance above base of lithofacies</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Loretta Road</td>
</tr>
<tr>
<td>Lithofacies I</td>
<td>N/A</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Lithofacies II Subfacies</th>
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<th>0-2.5 m</th>
<th>N/A</th>
<th>N/A</th>
</tr>
</thead>
<tbody>
<tr>
<td>a) shale and siltstone</td>
<td>N/A</td>
<td>2.5-5.5 m</td>
<td>0-1.5 m</td>
<td>N/A</td>
</tr>
<tr>
<td>b) quartz arenite</td>
<td>N/A</td>
<td>2.5-5.5 m</td>
<td>0-1.5 m</td>
<td>N/A</td>
</tr>
<tr>
<td>c) sandstone and limestone</td>
<td>10 m</td>
<td>5.5-9.0 m</td>
<td>N/A</td>
<td>N/A</td>
</tr>
</tbody>
</table>

| Lithofacies VII          | N/A | N/A | N/A | 0-30 m |

Parasequences sets (n+1 order) and Subfacies within Early Transgressive Systems Tract A

Letters to the left of sections correspond to subfacies designations shown above. Numeral to the right of sections correspond to lithofacies designations shown in Table 1 and Figure 7. Patterns in sections correspond to rock types shown in Figure 7. Arrows to the right of sections represent n+1 order relative sea-level change.
Ball-and-pillow features are typically 1 m long and 20 to 30 cm thick (Figure 13e). They are composed of siltstone and sandstone in shale. Their internal structure is similar to an isoclinal recumbent fold with two sub-parallel limbs roughly parallel to bedding. The axes of the folds are oriented northwest. The plunges of some folds are to the northeast, while others are to the southwest. Deformed parallel laminae are preserved locally within the folds.

Wave ripples are rare in Subfacies IIa (Figure 13f). Where present, they have an average height of 5 cm and a wavelength of 20 cm. In profile, the ripples are symmetric and exhibit fairly sharp crests and rounded troughs. Their internal structure, however, is more complex, and generally indicate flow towards the north.

The shale-and-siltstone subfacies (Subfacies IIa, Table 2) of Lithofacies II (Table 2) at Hines Ridge is directly overlain by a 3-m-thick succession of quartz-arenite subfacies (Subfacies IIb, Table 2) containing abundant parallel-laminated sandstone (Figure 13b) interbedded with minor siltstone and shale. The sandstone increases upsection volumetrically until the siltstone and shale are almost absent. The sandstone is a dark-brown to red weathering, carbonate-cemented, quartz arenite (Figure 13b). It is present in continuous to slightly lenticular beds ranging from 0.2 to 1.2 m thick. Abundant parallel laminae (Figure 13b) and rare parting lineations are the only sedimentary structures present in this subfacies.
At Magruder Mountain, Lithofacies II (Table 1) is limited to a 1.5-m-thick exposure of the parallel-laminated sandstone (Subfacies IIb, Table 2) with individual beds ranging in thickness from 5 cm to 10 cm. Although the exposure at Magruder Mountain is believed to be the quartz-arenite subfacies, both the lower and upper surfaces of the outcrop are covered; thus it is difficult to tell which subfacies it belongs to because they all contain some of this lithology. It is assigned to Lithofacies II (Table 1) because it is a sandstone-dominated lithofacies that is found below Lithofacies III (Table 1) and therefore appears to be stratigraphically equivalent to Lithofacies II (Table 1) at Loretta Road and Hines Ridge. The quartz-arenite subfacies is not present at Loretta Road or Mt. Dunfee.

The quartz arenite is composed of subrounded to subangular, medium sand- to silt-sized quartz and rare (less than 1%) plagioclase grains (Figure 13c). The quartz grains have quartz overgrowths, and some show calcite replacement. Thus, their original detrital shape can rarely be determined. The sandstone contains ferroan calcite and dolomite, which become more abundant upsection, as well as an iron-rich matrix (Figure 13c).

The parallel laminae are only a few millimeters thick, are slightly undulose, and are present in beds that range from decimeter to meter scale (Figure 13c). Thickness of the individual laminae is closely related to grain size, such that thicker laminae contain coarser grains. Microscopically, these sandstone beds display crudely alternating coarse- and-fine laminae. A single
lamina differs from those with which it is intercalated in terms of grain size, packing, sorting, thickness or a combination of features. Parting lineations are often present within these parallel-laminated beds and indicate a roughly north-south paleoflow direction. Too few paleocurrent indicators were present for statistical representation of the paleoflow direction.

Directly overlying the quartz-arenite subfacies at Hines Ridge is a 3.5-m-thick succession of the sandstone-and-limestone subfacies (Subfacies IIc, Table 2) of Lithofacies II (Table 1). This subfacies contains the same parallel-laminated sandstone described in the quartz-arenite subfacies (Figure 13b), but it is interbedded with limestone beds and only rare siltstone and shale beds. The limestone is very poorly preserved wackestone to packstone. Bed thickness increases upsection from 0.2 m at the base of the subfacies to 0.8 m at the top. Internal structures, other than rare parallel laminae, are absent. The Loretta Road section is the only other section in which the sandstone-and-limestone subfacies is found (Table 2). There, the limestone beds are highly altered by contact metamorphism and often contain actinolite. The sandstone beds of Subfacies IIc are lenticular, and they thin from 1 m at the base of the subfacies to 0.4 m at the top.

Limestone is present, and even becomes dominant in the upper part of the sandstone-and-limestone subfacies of Lithofacies II (Table 1). Nevertheless, the parallel-laminated-sandstone lithofacies (Lithofacies II, Table 1) is distinguishable from the overlying clotted-and-intraclastic-limestone lithofacies.
(Lithofacies III, Table 1) because of the abundance of siliciclastic sediment in Lithofacies II (Table 1) and the rarity of siliciclastic sediment in the overlying Lithofacies III (Table 1).

Transgressive Sandstone Deposit: Depositional Interpretation

A variety of sedimentary structures are present in the parallel-laminated-sandstone lithofacies (Lithofacies II, Table 1) providing information on the environment of deposition for this lithofacies. The most common sedimentary structure preserved in the lithofacies is parallel laminae. Proposed mechanisms for the formation of parallel laminae call on either suspension settling of sediment, deposition at high flow regimes, or a combination of the two (Lombard, 1963; Sanders, 1965; Kuenen, 1966; Smith, 1971; Reineck and Singh, 1972; McBride and others, 1975; Bridge, 1978; Allen, 1982; 1984; Cheel and Middleton, 1986; Bridge and Best, 1988; 1990; Paola and others, 1989; Cheel, 1990a; 1990b; Arnott, 1993).

As in the parallel-laminated-sandstone lithofacies (Lithofacies II, Table 1), parting lineations are often recorded associated with parallel-laminated beds (Allen, 1982; Tucker, 1982; Boggs, 1987). This association reflects the high bed-shear stresses, typically associated with very shallow water depths, required for the formation of both features (Allen, 1982). The parting lineations are aligned parallel to the flow direction and indicate unidirectional flow (Tucker, 1982). The parallel laminae in this lithofacies are similar to the laminae
attributed to combined flow regimes described by Arnott (1993) in that they are
only a few grain diameters thick, slightly irregular, and contain no mud. Because
these laminae are associated with parting lineations indicating unidirectional
flow, they are most likely a result of combined oscillatory and unidirectional flow
(Arnott, 1993) in very shallow water.

Cross-stratification similar to that recorded in Lithofacies II is well known
in sandstones and siltstones. It forms as a result of migrating bedforms and
typically represents the avalanche face of the bedform (Allen, 1982). The exact
bedform is a consequence of particle size and flow regime, which is in turn
controlled by flow velocity, water depth, and acceleration due to gravity (cf. Blatt
and others, 1980; Allen, 1982). The low-angle, planar cross-stratification in
Lithofacies II (Table 1) suggests that water depth increased or flow velocity
decreased, or both, from that in which the parallel-laminated beds were
deposited. The interbedding of these two stratification types indicate a shifting
of environments related to small scale sea-level fluctuations.

Ball-and-pillow features form as a result of the gravitational instability of
liquidized sands and muds, causing the denser sand to "sink" into the underlying
mud, displacing water and sediment (Allen, 1982). These structures are
typically dish-shaped features with upturned edges. However, rod-like and s-
shaped examples also have been described from disrupted turbidites, shallow-
marine, tidal, and deltaic deposits (Allen, 1982). Considering the associated
sedimentary structures in Lithofacies II (Table 1), the ball-and-pillow structures
in the lower Deep Spring Formation probably formed in a shallow-marine environment.

The co-occurrence of alternating coarse-and-fine parallel laminae, truncated sets of low-angle planar cross-stratification, and primary current lineations requires a high Froude number (Allen, 1982). A high Froude number is achieved in environments with a high-flow velocity and very shallow water. The beach to near-shore depositional environment meets these criteria, and these features commonly co-exist in such deposits (Elliot, 1981; Allen, 1982; Tucker, 1982; Inden and Moore, 1983; Leckie and Krystinik, 1989; Vilas and others, 1991). The lack of inverse grading in the sandstone and lack of other sedimentary features typical of beach deposition (Elliot, 1981; Allen, 1982; Tucker, 1982; Inden and Moore, 1983), however, suggest the beach deposition is not preserved in this lithofacies. The parallel-laminated-sandstone lithofacies (Lithofacies II, Table 1) is therefore interpreted as recording deposition in a high-energy nearshore environment.

As shown in Table 2, the stacking pattern of the subfacies that form the parasequence sets within Lithofacies II are interpreted to record deposition during a relative rise in sea level. This interpretation is based on the gradual change upsection from siliciclastic-dominated facies to carbonate-dominated facies, rather than the stacking patterns of individual parasequences, which cannot be accurately determined.
Transgressive Sandstone: Sequence-Stratigraphic Interpretation

The combination of features from both the shallow shelf and deeper shelf basin leads to an overall interpretation of events taking place during the transgressive systems tract. As relative sea level began to rise, siliciclastic sands that were probably trapped shoreward of the exposed carbonate platform began a limited seaward migration over the eroded platform, possibly due to redistribution of fluvial sediment by marine processes similar to those described by Evans and Hine (1991). These redistributed sands produced the parallel-laminated-sandstone lithofacies (Lithofacies II, Table 1) at the Hines Ridge, Loretta Road, and Magruder Mountain sections. Limited redistribution of fluvial sediment and the potential for entrapment of these sediments in the paleokarst surface of the exposed platform may explain the lack of siliciclastic deposits further seaward at the Mt. Dunfee section. Seaward of the siliciclastic sediment, sea-level lowstand is marked by the top of the lowstand prograding wedge being reworked into the grainstone. As sea level continued to rise during the transgressive systems tract, the area of potential carbonate production increased as the shelf was inundated by marine waters. However, the establishment of a productive carbonate-producing community typically lags behind inundation (Read, 1985). Therefore, sediment supply to the Deep Spring Formation deep shelf was extremely limited. This limited sediment supply, combined with rapid subsidence, resulted in a distally steepened ramp and subsequent starved shelf basin. Evidence for the low sedimentation rate in the...
basin is the marine hiatal surface, that is interpreted as coincident with the top of the reworked strata, at the top of the lowstand prograding wedge at Mt. Dunfee. Because no deposit representing the transgressive systems tract is present, the transgressive surface is interpreted as the same surface as the marine hiatal surface.

**Late Transgressive Systems Tract A**

Sandstones of the early transgressive systems tract (Lithofacies II, Table 1) are overlain directly by upward-shoaling parasequence sets that thicken and deepen upward and are composed of the clotted-and-intraclastic-limestone lithofacies (Lithofacies III, Tables 1 and 3) at the Inyo Mountain sections (Figure 7). During the late transgressive systems tract (Figure 14), carbonate production resumed on the platform and siliciclastic sedimentation no longer dominated the environment.

**Late Transgressive Systems Tract A: Description**

The clotted-and-intraclastic-limestone lithofacies (Figure 15: Lithofacies III, Tables 1 and 3) crops out only at Hines Ridge, Magruder Mountain and Loretta Road, where it directly overlies the parallel-laminated-sandstone lithofacies (Lithofacies II, Table 1). Lithofacies III (Table 1) varies in thickness from section to section, generally thinning in a northeasterly direction from approximately 70 m at Loretta Road to only 52 m at Magruder Mountain (Figure 7). Lithofacies III contains three subfacies (Table 3): (a) an intraclastic-
TABLE 3. Summary of Late Transgressive Systems Tract A

<table>
<thead>
<tr>
<th>Subfacies</th>
<th>Distance above base of lithofacies</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Loretta Road</td>
</tr>
<tr>
<td>a) intraclastic wackestone and lime mudstone</td>
<td>0-9 m</td>
</tr>
<tr>
<td></td>
<td>18-38 m</td>
</tr>
<tr>
<td>b) lime mudstone and siltstone</td>
<td>9-18 m</td>
</tr>
<tr>
<td>c) intraclastic wackestone and sandstone</td>
<td>38-43 m</td>
</tr>
</tbody>
</table>

Systems Tracts (n order)
Relative Sea Level
Falling Rising

Parasequence sets (n+1 order) and Subfacies within Late Transgressive Systems Tract A.

Maximum Flooding Surface A

Letters to the left of sections correspond to subfacies designations shown above. Numeral to the right of sections correspond to lithofacies designations shown in Table 1 and Figure 7. Patterns in sections correspond to rock types shown in Figure 7. Arrows to right of section represent n+1 order relative sea-level change.

Arrows represent n order relative sea-level change.

? Top assumed in covered interval

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Figure 14. **Late Transgressive Systems Tract**: Interpretive block diagram of deposition during Late Transgressive Systems Tract A. As sea level continued to rise during the late transgression, the area of limited carbonate production expanded shoreward and carbonate production increased.
Figure 15a. Late Transgressive Systems Tract: Photomicrograph of stained clotted limestone in plane polarized light. Note the peloidal structure within some of the clots. Location HR91124; bar is approximately 1 mm.

Figure 15b. Late Transgressive Systems Tract: Wavy parallel-laminated limestone at Hines Ridge. Note the slightly irregular nature of the laminae. Location HR91108; scale is approximately 15 cm long.

Figure 15c. Late Transgressive Systems Tract: Cross-stratified intraclastic limestone at Hines Ridge, which overlies a parallel-laminated limestone; arrow at contact. HR91140; bar is approximately 20 cm.
wackestone-and-lime-mudstone subfacies, which is present at Hines Ridge, Loretta Road, and Magruder Mountain: (b) a lime-mudstone-and-siltstone subfacies, which is present at Hines Ridge and Loretta Road; and, (c) an intraclastic-wackestone-and-sandstone subfacies, which is present only at Loretta Road.

At all locations where Lithofacies III is present, the intraclastic-wackestone-and-lime-mudstone subfacies (Subfacies IIIa, Table 3) is volumetrically the most abundant. The distribution of the lithologies within this subfacies is shown in Figure 7 and Table 3. At Hines Ridge, this subfacies comprises the lower 32 m of Lithofacies III. At Loretta Road, this subfacies is present in the lower 9 m of Lithofacies III, from 18 m to 38 m above the base of Lithofacies III. The subfacies comprises the entire 33 meters of Lithofacies III at Magruder Mountain. The subfacies also comprises the upper 20 m of Lithofacies III at Hines Ridge, the upper 28 m of Lithofacies III at Loretta Road, and the upper 7 m of Lithofacies III at Magruder Mountain, but is considered part of the Highstand Systems Tract and is shown on Table 4 instead of Table 3.

Although lime mudstone and intraclastic wackestone are the most common rock types within this subfacies, the lime mudstone of the upper 20 m of Lithofacies III of the Hines Ridge section has undergone extensive neomorphism and is now structureless dolomite (Figure 7).

Bedding in the lime mudstone of the intraclastic-wackestone-and-lime-mudstone subfacies (Subfacies IIIa, Table 3) typically is 0.3 to 0.6 m thick and
laterally continuous, and often has slightly undulatory contacts with associated beds. Internally, beds appear either structureless or parallel laminated in outcrop. The lime mudstone that appears structureless in outcrop, however, displays a clotted texture that is visible microscopically. In thin-section, individual clots of microspar (0.5 to 6 mm) are distinguishable with calcite spar between, and some of the larger clots contain peloidal structures within them (Figure 15a).

The laminae of the parallel-laminated lime mudstone of Subfacies IIIa (Table 3) are most obvious on the weathered surface of the outcrop. The laminae are typically less than 10 mm thick and are commonly parallel and wavy (Figure 15b). Microscopically, the laminated lime mudstone appears either structureless because of recrystallization or it contains 1 mm couplets of alternating very fine-grained quartz sand (0.125 mm in diameter) and microspar.

The intraclastic nature of the intraclastic wackestone of Subfacies IIIa (Table 3) is most obvious on a weathered outcrop surface. It usually crops out as a light-gray wackestone containing 5-mm-long white intraclasts. The intraclasts are recrystallized, irregular in shape and often display a clotted texture. These intraclastic-wackestone beds display rare tabular cross-lamination in centimeter-scale sets (Figure 15c); more commonly, they are homogenous with no discernable internal structure. Where cross-laminated, the intraclasts are aligned parallel to the cross-laminae.
The lime-mudstone-and-siltstone subfacies (Subfacies IIIb, Table 3) of Lithofacies III is present at the Hines Ridge section 32 m to 43 m above the base of Lithofacies III, and at the Loretta Road section 9 m to 18 m above the base of Lithofacies III (Figure 7). This subfacies contains the same lime mudstone described previously in Subfacies IIIa, as well as siltstone. The dark brown siltstone is typically present in lenticular, parallel-laminated beds that are about 5 cm thick.

The intraclastic-wackestone-and-sandstone subfacies (Subfacies IIIc, Table 3) of Lithofacies III is present at the Loretta Road section 38 m to 43 m above the base of Lithofacies III (Figure 7). It is present at no other measured sections. It contains the same intraclastic wackestone and siltstone described previously from Lithofacies III (Table 1). In addition, this subfacies contains a red-brown, parallel-laminated, quartz arenite. This sandstone is present in slightly undulatory, lenticular beds that are typically 0.3 to 0.5 m thick. The parallel laminae are only a few millimeters thick and slightly undulose.

Late Transgressive Systems Tract A:
Depositional Interpretation

The variation between lime mudstone and intraclastic wackestone in Subfacies IIIa (Table 3) is interpreted as the depositional record of the change from tidal-flat deposition to shallow-subtidal deposition, respectively. The abundance of thin-bedded lime mudstone in this subfacies suggests a moderately low-energy environment. The clotted texture of the lime mudstone
and the irregular wavy nature of some of the parallel laminae may be a result of cryptmicrobial binding of sediment, as suggested for similar rocks by Goldhammer and others (1993). However, no microbial organisms could be identified. Cryptmicrobial laminae may represent periods of deposition in a relatively low-energy, shallow-water environment, such that bedforms could not develop, and in which microbial organisms could bind the sediment, such as on a tidal flat. The existence of parallel laminae indicates that the type or the size of the material deposited varied over time. This variation may be a result of energy fluctuations, binding of material by organisms, or both (Flügel, 1978). Cryptmicrobial binding can take place in a variety of different calm-water environments. However, no larger cryptmicrobial structures, such as stromatolitic mounds, are present. The water depth was probably too shallow for such structures to form. Therefore, the lime mudstones of Subfacies IIIa (Table 3) are interpreted as having been deposited on a tidal-flat environment. Thin-bedding, lime mudstone, clotted textures and microbial binding are all features common to the tidal-flat environment (Tucker and Wright, 1990).

Intraclasts typically are formed as a result of the erosion and redeposition of partially lithified or cryptmicrobially bound sediments (see Folk, 1959, 1962; and Dunham, 1962 for a more complete discussion of intraclasts). The intraclastic wackestone of Subfacies IIIa (Table 3) may have originated from the break-up of parallel-laminated beds with which these intraclastic beds are often interbedded. Although intraclasts may also form from desiccation of sediment,
no indication of subaerial exposure was observed. Therefore, high-energy, subtidal events are called upon in this interpretation to produce enough velocity to erode the intraclasts and to form the cross-laminated intraclastic beds without evidence of subaerial exposure. These beds are interpreted as subtidal lagoon deposits because no evidence of subaerial exposure was seen. Lagoons that have unrestricted circulation typically display similar features (Tucker and Wright, 1990).

The combination of features in the intraclastic-wackestone-and-limestone-mudstone subfacies (Subfacies IIIa, Table 3) of the clotted-and-intraclastic-limestone lithofacies (Lithofacies III, Table 1) indicates variations between low-energy, shallow-water (cryptmicrobial binding) and high-energy, deeper water (cross-bedding) deposition. The change from shallow-water, tidal-flat deposits to subtidal-lagoon deposits is repeated cyclically within Lithofacies III (Table 1; Table 3). This subfacies is interpreted as being deposited in an environment that varied periodically between a tidal flat and shallow subtidal lagoon because of small-scale relative sea-level change. An individual couplet of basal lagoon deposits overlain by tidal-flat deposits is here interpreted as a parasequence. All of the parasequences within Lithofacies III shallow upsection. Multiple parasequences of varying thickness, designated by arrows to the right of the columns on Table 3, that are composed of Subfacies IIIa lithologies stack together into parasequence sets that are capped by siliciclastic deposits of either the lime-mudstone-and-siltstone subfacies (Subfacies IIIb, Table 3) or the
intraclastic-wackestone-and-sandstone subfacies (Subfacies IIIc, Table 3). A repetitive succession of Subfacies IIIa parasequences overlain by a siliciclastic cap (Subfacies IIIb or IIIc, Table 3) is interpreted as an upward-shallowing parasequence set. These siliciclastic beds record the dispersal of siliciclastic sediments across the tidal flat at the time of lowest relative sea level. Parasequences sets within depositional sequence A appear to thicken upsection (Table 3). This stacking pattern indicates that accommodation space was increasing on the shelf, suggesting a relative rise in sea level corresponding to the Transgressive Systems Tract.

Although the parasequence sets thicken upsection, a great deal of variability is seen in the thicknesses of parasequences that make up the parasequence sets. As seen in Table 3, thicknesses of lagoonal and tidal-flat lithologies within an individual cycle vary upsection. In some cases, the intraclastic limestones of the lagoon are much thicker than the clotted limestones of the tidal flat; in an overlying succession, the two facies are almost the same thickness. These variations in thickness are probably due, in part, to the autocyclic nature of the depositional environments (cf. Wilkinson, 1982; James, 1984; Pratt and James, 1986), such as the natural variation in size of the tidal-flat or lagoon environment and distribution of sediments within the environment. The thickness of an individual tidal-flat deposit may be controlled by the extent and direction in which it prograded — thicker deposits may have prograded further, whereas thinner deposits may have prograded less, thus allowing for
thicker lagoon deposits. In addition, the progradation may have been in such a direction that the outcrop in which it is seen represents only the edge of the deposit, resulting in a much thinner deposits than would be seen if the section represented the thickest part of the depositional environment. Regardless of the exact reason, the variations in thickness of individual rock types is probably more a result of autocyclic mechanisms than of eustatic changes. Although autocyclicality may account for the variable thickness of the beds within the parasequences, it cannot alone account for thickness of the actual parasequences, which are ultimately controlled by relative sea-level change (cf. Grotzinger, 1986; Osleger, 1991; Goldhammer and others, 1993). However, the effects of autocyclicality on the deposits, when superimposed on the effects of sea-level change that control parasequence deposition, can influence the thickness of the parasequences. As a result, a clear record of relative sea-level change is not apparent until the stacking pattern of the parasequence sets is recognized (Table 3).

Late Transgressive Systems Tract A: Sequence-Stratigraphic Interpretation

During the late transgression, sea level continued to rise, as indicated by the parasequence-set stacking pattern (Table 3) present within the Deep Spring Formation. Siliciclastic deposition that was dominant across the Inyo Mountain sections during the early transgressive systems tract migrated landward again as relative sea level rose. The landward migration of siliciclastics allowed
carbonate-sediment production to begin to reestablish on the shallow shelf and continue through the highstand systems tract. The deeper water Mt. Dunfee section continued to experience a depositional hiatus during deposition of the transgressive systems tract in shallow parts of the basin. As a result of continued subsidence and no deposition during this time, the Mt. Dunfee section deepened, creating a slope, which had a significant impact on later deposition. This impact will be discussed in the "Early Highstand Slump" section.

**Highstand Systems Tract A**

The transgressive systems tract at the Inyo Mountains is overlain by the uppermost parasequences of the clotted-and-intraclastic-limestone lithofacies (Lithofacies III, Table 1). The transgressive surface/maximum flooding surface at Mt. Dunfee is directly overlain by the contorted-limestone lithofacies (Lithofacies VIII, Table 1). The contorted-limestone lithofacies is in turn overlain by the lime-mudstone-and-bioclastic-wackestone subfacies (Subfacies IVa) and the bioclastic-and-peloidal subfacies (Subfacies IVb) of the bioclastic-limestone lithofacies (Lithofacies IV, Table 4). This succession is interpreted as the record of initial deposition during Highstand Systems Tract A (Figure 16). The presence of the contorted-limestone lithofacies resting on the maximum flooding surface has prompted the proposal of the term *early highstand slumps* for these features.
Highstand Carbonate Parasequences at the Inyo Mountains

At the Inyo Mountain sections, the uppermost upward-shallowing parasequence set, consisting of 30 m of clotted-and-intraclastic-limestone lithofacies (Lithofacies III, Table 4) at Loretta Road and 20 m of Lithofacies III at Hines Ridge, is here interpreted as the highstand systems tract (Figure 16) because it is the thickest parasequence set underlying the sequence boundary. Because of the poor exposure at Magruder Mountain, it is unclear exactly where, or even if, the change from transgression to highstand occurs. If present, the change probably takes place in the covered portion of the section; thus, the upper 7 m of Lithofacies III are here considered part of the highstand systems tract.

The uppermost parasequence set of Lithofacies III is believed to rest on the maximum flooding surface because its thickness is greater than underlying parasequence sets, signaling accumulation at a time of maximum accommodation (Table 4). However, Sequence A is truncated by Sequence Boundary B. As a result, the upper portion of Highstand Systems Tract A, where parasequence sets should thin and prograde seaward (cf. Mitchum and Van Wagoner, 1991) is not present within the lower Deep Spring Formation. The cause of truncation in Sequence A will be discussed in the “Sequence Boundary B” section. For the purposes of this study, these parasequence sets are considered part of Highstand Systems Tract A. This interpretation is based on the thickening of the parasequence sets and the assumption that these thicker
TABLE 4. Summary of Highstand Systems Tract A

Lithofacies III: Clotted and Intraclastic Limestone
Lithofacies IV: Bioclastic Limestone
Lithofacies VIII: Contorted Limestone

<table>
<thead>
<tr>
<th>Lithofacies IV Subfacies</th>
<th>Distance above base of lithofacies</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Loretta Road</td>
</tr>
<tr>
<td>a) lime mudstone and bioclastic wackestone</td>
<td>N/A</td>
</tr>
<tr>
<td>b) bioclastic and peloidal limestone</td>
<td>N/A</td>
</tr>
<tr>
<td>Lithofacies III: Subfacies a) intraclastic wackestone and lime mudstone</td>
<td>43-73 m</td>
</tr>
<tr>
<td>Lithofacies VIII</td>
<td>N/A</td>
</tr>
</tbody>
</table>

Systems Tract (n^th order)
Relative Sea Level  Falling    Rising

Parasequence sets (n+1 order) and Subfacies within Highstand Systems Tract A.

Letters to the left of sections correspond to subfacies designations shown above. Numeral to the right of sections correspond to lithofacies designations shown in Table 1 and Figure 7. Patterns in sections correspond to rock types shown in Figure 7. Arrows to right of sections represent n+1 order relative sea-level change.

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Figure 16a. **Highstand Systems Tract**: Interpretive block diagram of deposition during the Early Highstand Systems Tract. Note the Early Highstand Slumps, which are interpreted as unique features of mixed carbonate-siliciclastic systems.

Figure 16b. **Highstand Systems Tract**: Interpretive block diagram of deposition during the Early Highstand Systems Tract. As sea level approaches its maximum, carbonate production establishes across the entire shelf. Light blue circles represent occurrence of small shelly fossils.
parasequence sets are temporally equivalent to the Highstand Systems Tract A deposits at Mt. Dunfee. Because a complete passive-margin type depositional sequence is not preserved, it is impossible to determine if these parasequence sets are truly the thickest parasequence sets. These parasequence sets could alternatively represent late transgressive systems tract deposits that might have been overlain by still thicker parasequence sets if a complete passive-margin type depositional sequence (sensu Van Wagoner and others, 1988) were preserved here.

Early Highstand Slumps: Description

At the deeper water Mt. Dunfee section, the reworked bed and hiatal surface of the transgressive surface is directly overlain by the dark-gray contorted-limestone lithofacies (Lithofacies VIII, Table 4) (Figure 7). These contorted limestones are interpreted to represent sediment that was deposited on the slope that subsequently slumped down the distally steepened ramp that formed because rates of sea-level rise outpaced carbonate accumulation rates in this region.

The contorted-limestone lithofacies (Lithofacies VIII, Table 1) is composed of slightly to highly contorted, thinly interbedded dark-gray lime mudstone to intraclastic packstone (Figure 17a). This lithofacies is 7 m thick and crops out only in the Mt. Dunfee section (Figure 7, Table 4). The most contorted limestones are present at the base of the unit, and limestone beds become less contorted toward the top of the lithofacies. The disrupted bedding
Figure 17a. **Early Highstand Slumps:** Highly contorted limestone beds interpreted as sediment slumped toward the shelf basin. At Mt. Dunfee, these slumps directly overlie a depositional hiatus. Note the varying styles of contortions (A); and ball-and-pillow-like features (B) within the slumps. Location MD91I40; 40 cm long hammer.

Figure 17b. **Highstand Systems Tract:** Dark bioclastic grainstone that is present within the lighter lime mudstone of the bioclastic-wackestone subfacies at Mt. Dunfee. Dark ellipses (arrow) are small shelly fossils. Location MD91I05; bar is approximately 5 cm.

Figure 17c. **Highstand Systems Tract:** Peloidal bioclastic limestone at Mt. Dunfee. Red coloration is due to hematite replacement of grains. Location MD91I25; bar is approximately 0.5 m.

Figure 17d. **Highstand Systems Tract:** Plane-polar photomicrograph of stained hematite (A) and glauconite (B) replacement of small shelly fossil or peloid. Location MD91I25; bar is approximately 0.25 mm.
is visible in 0.5- to 1-m-thick packages that exhibit varying styles of deformation. These styles include parallel, wavy bedding; concentric, elliptical to cuspate folds that are often recumbent; and ball-and-pillow-like features that weather out in rounded, 5- to 10-cm-thick pods. Many of the folds in the lower beds appear to have ruptured along the axial surfaces. Nearly planar surfaces truncate the tops of some folded packages and form the base of overlying folded packages; these surfaces may represent shear planes along which the folded packages traveled during emplacement. Other disrupted beds are overlain conformably by flat lying, undisturbed beds. Lithofacies VII! (Table 1) is directly overlain at Mt. Dunfee by the bioclastic-limestone lithofacies (Lithofacies IV, Table 1), which is described below.

**Early Highstand Slumps: Depositional Interpretation**

Slump deposits are well known in both modern (Coniglio, 1986; Coniglio and James, 1990; Kenter, 1990) and ancient (Hurst and others, 1985; Eberli, 1987; Gibling and Stuart, 1988) slope settings. Although many of the slump beds described in the literature are of a much larger scale than those of the Deep Spring Formation, small-scale slumps are presently forming off the eastern coast of North America (Knebel and Carson, 1979). Regardless of the size, the mode of origin is believed to be similar.

A variety of features have been cataloged to identify syndepositional slumped beds. The most common features in slumped beds is the presence of
deformed beds between undisturbed beds (Rupke, 1981; Allen, 1982; Cook and Mullins, 1983; Enos and Moore, 1983). These deformed beds can range from cohesive but angularly discordant blocks to contorted masses (Enos and Moore, 1983) in a variety of sizes (Rupke, 1981). The upper surface of the folds may be eroded or possibly wavy. Because slump masses move elastically or elastically and plastically, relict bedding is often preserved (Cook and Mullins, 1983), as it is in the Deep Spring Formation. Because slump folds generally occur between undisrupted beds, folding is interpreted as syndepositional and not the result of later tectonic deformation (sensu Rupke, 1981; Cook and Mullins, 1983; Enos and Moore, 1983; Elliott and Williams, 1988).

Soft-sediment deformation typically takes place as a result of translational movement of partially lithified sediment. Failure along a shear plane on a depositional slope is one of the most common ways that partially lithified sediment moves downslope (Cook and Mullins, 1983). Although these features can form on gentle slopes, they are much more common on steeper slopes (Allen, 1982; Enos and Moore, 1983). Slumps that form on gentle slopes are typically a result of progressive downslope movement of material that begins moving on a steep slope and later overrides a lower-gradient slope and causes slumping of underlying sediments on the lower-gradient slope (Rupke, 1981). Rapid deposition, fine grain size, and a lack of intergranular friction (Allen, 1982; Enos and Moore, 1983), as well as differing sediment porosity (Nelson and Lindsley-Griffin, 1987), can also induce slumping. These mechanisms, however,
are not common in carbonate environments because of the rapid cementation rates of carbonate sediments. An exception arises where cementation rates differ enough between beds to allow for the formation of shear planes (Coniglio and James, 1990).

Initial sediment movement may be triggered by earthquakes, storms, oversteepening, or increased pore-fluid pressure (Allen, 1982). Gas generation by microbial processes has also been suggested as a cause of shear-plane development (Nelson and Lindsley-Griffin, 1987). These mechanisms only work, however, if a slope already exists. Whether earthquakes or storms acted as a triggering mechanism, the slumps in the Deep Spring Formation indicate that a depositional slope steep enough to allow for down-slope slumping had developed. As discussed in Transgressive Systems Tract A, this slope was probably in the form of a distally steepened carbonate ramp.

**Early Highstand Slumps: Sequence-Stratigraphic Interpretation**

Typically, highstand shedding is minimal in carbonate environments until sediment production nears its maximum, when shelf sediment fills available accommodation space and basinward progradation dominates (Droxler and Schlager, 1985; Dolan, 1989; Mullins, 1983), typically during the late highstand systems tract. Slumps in the Deep Spring Formation, however, are thought to be a result of the interplay between carbonate and siliciclastic facies during the Early Highstand Systems Tract. While sands were being deposited and
carbonate production was being re-established on the platform during the late transgressive systems tract, the shelf basin was starved. Carbonates of the late transgression accumulated rapidly on the ramp, while subsidence continued to deepen the shelf basin. The combined effect resulted in further development of relief between the carbonate ramp and the shelf basin much earlier than usually occurs in pure carbonate systems. As a result of the relief (i.e., distally steepened ramp), sediment was shed off the platform and slumped downslope. Therefore, this study proposes the term early highstand slumps for these features, and they are interpreted to be a unique product of a mixed carbonate-siliciclastic system. These slump deposits are overlain by the bioclastic-limestone lithofacies (Lithofacies IV), indicating carbonate production and deposition eventually was established at Mt. Dunfee. This carbonate deposition suggests sedimentation was able to outpace subsidence, prograde seaward, and reduce the relief in the Mt. Dunfee area.

Highstand Systems Tract at Mt. Dunfee:
Description

The bioclastic-limestone lithofacies (Lithofacies IV, Table 1) is only present at Mt. Dunfee and contains a lime-mudstone-and-bioclastic-wackestone subfacies (Subfacies IVa) and a bioclastic-and-peloidal-limestone subfacies (Subfacies IVb; Table 4). The lime-mudstone-and-bioclastic-wackestone subfacies comprises the lower 18 m of Lithofacies IV at Mt. Dunfee (Figure 7). The subfacies contains 0.2 m-thick beds of the lime mudstone similar to that
described previously for Lithofacies III, that is interbedded with a bioclastic wackestone to grainstone. The bioclastic wackestone is light- to medium-gray and appears to be a lime mudstone in outcrop, because the fossil-fragment allochems are rarely recognizable without the aid of a microscope. In a few beds, however, the fossil debris is discernable in the field within dark gray to black lenses (5 cm) of grainstone that are present within thin beds (10 to 30 cm) of wackestone (Figure 17b). In thin section, the bioclastic fragments are typically whole and deformed to broken small shelly fossils composed of recrystallized calcite. They are very faint and are difficult to recognize in thin section. Use of Dravis' (1991) white-card technique aided greatly in the recognition and identification of the fossils, which are discussed in more detail in Chapter 4.

At Mt. Dunfee, the lime-mudstone-and-bioclastic-wackestone subfacies (Subfacies IVa) is gradationally overlain by the bioclastic-and-peloidal-grainstone subfacies (Subfacies IVb; Table 4) of Lithofacies IV. Three sedimentological aspects change upsection in Lithofacies IV: (1) the abundance of small shelly fossils and peloidal grains increases; (2) parallel-laminated limestone beds are replaced by internally homogenous, undulose beds; and, (3) bedding thickness increases with the occurrence of the bioclastic-and-peloidal-grainstone subfacies. The bioclastic-and-peloidal-grainstone subfacies (Lithofacies IV, Subfacies IVb, Table 1) is composed of a 10-m-thick bioclastic-peloidal grainstone (Figure 17c). The shells in the bioclastic-peloidal grainstone
are whole and undeformed to slightly deformed small shelly fossils. The associated peloids are 1 to 2.5 mm in diameter, hematite-replaced, and elliptical. They have abundant euhedral magnetite crystals and common glauconite (Figure 17d). Hematite, magnetite, and glauconite have replaced most of the small shelly fossils associated with these peloids; replacement becomes more abundant upsection. Bedding thickness in this lithofacies is about 0.5 m. Bedding is highly undulose on all planes, suggesting a hummocky nature to the beds (Figure 17c). However, no internal structures are preserved.

Previous studies of the lower Deep Spring Formation have identified small shelly fossils at all outcrop locations (Gevirtzman, 1983). In this study, an abundance of fossils was found only at Mt. Dunfee within the bioclastic-limestone lithofacies (Lithofacies IV, Table 1). Small shelly fossils were identified at the Loretta Road section, but the fossil-bearing beds are rare and are poorly preserved.

Highstand Systems Tract at Mt. Dunfee: Depositional Interpretation

The presence of broken shells that accumulated in lenses with only minor amounts of mud is indicative of reworking or transport and concentration of the shells (Tucker, 1982; Grant, 1990). The shells may have been transported by periodic storms because the accumulation of broken-shell lenses are typical of storm-lag deposits (Aigner, 1985).
The increase in small-shelly-fossil accumulations and thickening of bedding in the bioclastic-limestone lithofacies (Lithofacies IV, Table 1) are interpreted to record upsection deepening. The undulose, slightly mounded nature of the bedding may be remnant hummocky cross stratification, which, if present, would also support an upsection increase in accommodation space. For sediment of approximately the same size, the change from upper flow regime plane beds to hummocky cross stratification requires either an increase in oscillation period or decrease in orbital speed of the waves impinging on the environment (Southard, 1991). An increase in accommodation space could account for either of these requirements. Because the shells are generally undamaged, however, they probably underwent only minor, if any, transport (Grant, 1990). Therefore, energy levels must have been high enough to winnow most mud that was deposited, but insufficient to damage the shells.

Highstand Systems Tract at Mt. Dunfee: Sequence-Stratigraphic Interpretation

The upward thickening and deepening nature of the of the parasequence (Table 4) that makes up Lithofacies VIII and IV at Mt. Dunfee is indicative of early highstand deposition as relative sea level continued to rise prior to the sea-level fall that takes place during the late highstand systems tract. No Late Highstand Systems Tract A is recognized in the lower Deep Spring Formation because Systems Tract A is truncated by Sequence Boundary B.
**Sequence Boundary B**

As relative sea level reached its maximum rate of fall, a second sequence boundary (Sequence Boundary B) formed in the lower member of the Deep Spring Formation (Figure 7). Sequence Boundary B is defined by the contact between Highstand Systems Tract A and the overlying Lowstand Systems Tract B described below. However, as stated previously, no evidence for the Late Highstand Systems Tract A is present anywhere in the lower Deep Spring Formation. Two possibilities can explain the absence of the late highstand systems tract: either, 1) the late highstand systems tract was deposited and then eroded during formation of the Sequence Boundary B unconformity, or 2) the late highstand systems tract was never deposited. Either case is unusual if the lower Deep Spring Formation were deposited on a steadily subsiding passive margin. This suggests that local phenomena were controlling the development of accommodation along the Deep Spring margin. Faulting is a likely local phenomenon that could account for either possibility. Because the timing of breakup is uncertain along the Cordilleran continental margin, the lower Deep Spring Formation probably represents deposition at the end of the rift-to-drift transition or immediately following the onset of drift (Levy and Christie-Blick, 1991), as discussed in the “Tectonic History” section of Chapter 2. If faulting did occur, it would add additional support to the idea that the onset of drift took place higher in the stratigraphic section (Levy and Christie-Blick, 1991). Faulting of Middle Cambrian and Early Silurian age has been recognized in the

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The recognition of older fault activity in the lower Deep Spring Formation suggests that all of this fault activity may indicate the presence of fundamental crustal structures that were active as early as the Late Neoproterozoic. If faults were periodically active along the Cordilleran continental margin from Neoproterozoic through Early Silurian time, the margin may have been much more tectonically active than originally thought.

If faulting did take place during deposition of the lower Deep Spring Formation, the relative motion along the fault could impact the stratigraphic record. If faulting were to lower the platform, the carbonate platform would have probably drowned. There is no evidence, however, of drowning of the carbonate platform. Therefore, fault motion probably raised the lower Deep Spring platform instead of lowering it. However, the lower Deep Spring Formation does not contain evidence of subaerial exposure prior to deposition of the siliciclastic sediment of the lowstand systems tract. If the platform was raised high enough to expose it, the exposure period was not long enough to allow significant erosion or karstification to take place that could account for the absence of the late highstand systems tract. Thus, it is reasonable to conclude that faulting must have raised the platform, but not high enough to expose it subaerially.

Sequence Boundary B is directly overlain by siliciclastic sediments, indicating a lowering of relative sea level and this is manifested by a basinward shift in facies. The sudden change from carbonate deposition to siliciclastic
deposition suggests that the lower Deep Spring platform experienced a sudden fall in relative sea level that resulted in a basinward shift in facies, perhaps as a result of upward motion along a marginal fault. The return of siliciclastic sedimentation is interpreted as a basinward shift in facies that signals the initiation of the Lowstand Systems Tract B. Deposition of siliciclastic sediment would have smothered the carbonate platform that was producing limestone of Highstand Systems Tract A, resulting in the early termination of deposition during the highstand systems tract.

Although the lower Deep Spring Formation shows no evidence of exposure, shallower areas of the platform not present in the study area were probably raised out of the submarine environment and subaerially exposed. The presence of iron- and glauconite-replaced grains at the top of the bioclastic-limestone lithofacies (Lithofacies IV, Table 1), suggests that a period of slow deposition or non-deposition took place prior to deposition of the lowstand siliciclastic sediment. This hiatus may be correlative to the time of exposure and subsequent non-deposition on the shelf that is not recognizable in the study area.

Although possible faulting resulted in a basinward shift of facies and consequently produced a sequence boundary, it should be noted that this boundary is probably not regionally extensive. As a result, Sequence Boundary B is probably of little use in global correlations of unconformities and sea-level change.
Lowstand Systems Tract B

The cross-bedded-sandstone lithofacies (Lithofacies V, Table 1), which overlies the clotted-and-intraclastic-limestone lithofacies (Lithofacies III, Table 1; Figure 7) in the Inyo sections at Hines Ridge, Loretta Road, and Magruder Mountain, is interpreted in the following discussion as Lowstand Systems Tract B, which was deposited over the shallower part of the ramp (Figure 18). At Mt. Dunfee, the shale-and-siltstone lithofacies (Lithofacies IX, Table 1), overlies the bioclastic-limestone lithofacies (Lithofacies IV, Table 1) of Highstand Systems Tract A (Figure 7). These shales and siltstones are interpreted as some of the earliest siliciclastic sediments of the lowstand systems tract to reach the deeper part of the ramp, probably as a result of bypass sedimentation. If these interpretations are correct, then the contact between the limestone (Lithofacies III and IV, Table 1) and the overlying siliciclastic rocks (Lithofacies V and IX, Table 1) must, by definition, be a sequence boundary.

Early Lowstand Bypass Sedimentation:
Description

Interbedded, burrowed, dark-brown shale and reddish-brown siltstone comprise the shale-and-siltstone lithofacies (Lithofacies IX, Table 1; Figure 19a). This lithofacies is present only at the Mt. Dunfee section and directly overlies the iron- and glauconite-replaced peloids and small shelly fossils of the bioclastic-limestone lithofacies (Lithofacies IV, Table 1). Glauconite and iron are also present at the base of the shale-and-siltstone lithofacies (Lithofacies IX, Table 1).
Figure 18. **Lowstand Systems Tract**: Interpretive block diagram of deposition during the Lowstand Systems Tract B. Irregular yellow lines represent the presence of trace fossils.
Figure 19a. **Early Lowstand Bypass Sedimentation:** The shale-and-siltstone lithofacies (Lithofacies IX) at Mt. Dunfee interpreted as the initial siliciclastic deposits of Lowstand Systems Tract B. Siliciclastic sediments bypassed the shelf and were deposited in basinal regions at Mt. Dunfee. Location MD911I30; scale is approximately 25 cm long.

Figure 19b. **Lowstand Sandstone Deposit:** Loading structures of siltstone into shale in the siltstone-and-shale subfacies (Subfacies Va) of the cross-bedded sandstone lithofacies (Lithofacies V, Table 1) at Mt. Dunfee (arrow). Location MD91I131; scale is in inches, approximately equal to 15 cm in length.

Figure 19c. **Lowstand Sandstone Deposit:** Small hummocks in the siltstone-and-sandstone subfacies (Subfacies Vb) of the cross-bedded sandstone lithofacies (Lithofacies V, Table 1). Note the draping nature of the uppermost laminae (A) and the undulatory base (B) (arrows). Location MD91I132; scale is in inches, approximately equal to 15 cm in length.

Figure 19d. **Lowstand Sandstone Deposit:** Thick succession of cross-bedded sandstone (Subfacies Vc) at Mt. Dunfee. Note the general increase in bedding thickness upsection. Location MD91I138; scale is approximately 25 cm long.
1). Because of difficulty in obtaining thin sections from the shale, it is difficult to
tell if the glauconite and iron is authogenic or, more likely, reworked from the
underlying beds. Lithofacies IX (Table 1) is directly overlain by the siltstone-
and-shale subfacies (Subfacies Va) of the cross-bedded-sandstone lithofacies
(Lithofacies V, Table 1) at Mt. Dunfee.

The shale beds of Lithofacies IX (Table 1) are dark brown, parallel
laminated and 2 cm to 5 cm thick. The light-brown siltstone beds are also
parallel laminated and typically less than 5 cm thick. The entire lithofacies is
only 1 m thick and crops out as a soft weathering slope between Lithofacies IV
and Lithofacies V (Table 1).

Rare trace fossils are usually preserved as molds in the siltstone and
casts in the shale. All of the trace fossils were found in float. This slope,
however, was the only place that trace fossils were found in the entire lower
Deep Spring Formation. They are horizontal traces or resting marks and include
*Palaeophycus* or *Planolites, Scolicia, Protopalaeodictyon*, and a *Bergaueria*-like
trace. All of the forms and their biostratigraphic significance are discussed in
greater detail in Chapter 4.

**Early Lowstand Bypass Sedimentation:**
Depositional Interpretation

The presence of glauconite, which commonly forms during quiet water
depositional hiatuses, may be evidence for a period of non-deposition (Jenkyns,
1981). However, the glauconite in the shale is probably reworked. The
presence of horizontal traces and resting marks generally indicates a low-energy, marine depositional environment of moderate water depth (Seilacher, 1967; Johnson, 1981). Depending on grain size and water depth, parallel-laminated beds may form either as a result of high flow velocities or suspension settling (Allen, 1982; 1984). Given that the dominant grain size is clay to silt and that horizontal traces are present, it is unlikely that these beds were deposited under high velocities. Therefore, the parallel-laminated beds, in part, represent deposition in calm waters below fair-weather wave base. The horizontal traces and resting marks, however, are very rare in this succession. The low density of traces suggests that calm-water conditions favorable for their formation and preservation may have been only periodic, and that deposition was not very deep, probably only slightly below fair-weather wave base.

Lowstand Sandstone Deposit: Description

The cross-bedded-sandstone lithofacies (Lithofacies V, Table 1) (Figure 19b) is present at all measured sections of the lower Deep Spring Formation. It contains mainly cross-bedded sandstone with minor amounts of siltstone, shale, parallel-laminated sandstone, and lime mudstone at some locations that combine in four subfacies (Table 5): (a) a siltstone-and-shale subfacies, which is present only at Mt. Dunfee; (b) a siltstone-and-sandstone subfacies, which is present at Hines Ridge and Mt. Dunfee; (c) a sandstone subfacies, which is present at Loretta Road, Mt. Dunfee and Magruder Mountain; and, (d) a sandstone-and-limestone subfacies, which is present only at Loretta Road, and
TABLE 5. Summary of Lowstand and Transgressive Systems Tract B

<table>
<thead>
<tr>
<th>Lithofacies V Subfacies</th>
<th>Distance above base of lithofacies</th>
<th>Loretta Road</th>
<th>Hines Ridge</th>
<th>Magruder Mountain</th>
<th>Mt. Dunfee</th>
</tr>
</thead>
<tbody>
<tr>
<td>a) siltstone and shale</td>
<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
<td>0-1.25 m</td>
<td></td>
</tr>
<tr>
<td>b) siltstone and sandstone</td>
<td>N/A</td>
<td>0-12 m</td>
<td>N/A</td>
<td>1.25-2 m</td>
<td></td>
</tr>
<tr>
<td>c) sandstone</td>
<td>0-1 m</td>
<td>N/A</td>
<td>0-17 m</td>
<td>2-15 m</td>
<td></td>
</tr>
<tr>
<td>d) sandstone and limestone</td>
<td>1-6 m</td>
<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
<td></td>
</tr>
<tr>
<td>Lithofacies IX</td>
<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
<td>0-1 m</td>
<td></td>
</tr>
</tbody>
</table>

Parasequence sets (n+1 order) and Subfacies within Lowstand and Transgressive Systems Tract B.

Arrows represent nth order relative sea-level change.

Letters to the left of sections correspond to subfacies designations shown above. Numeral to the right of sections correspond to lithofacies designations shown in Table 1 and Figure 7. Patterns in sections correspond to rock types shown in Figure 7. Arrows to right of section represent n+1 order relative sea-level change.
is considered part of Transgressive Systems Tract B. Subfacies Vd will be discussed in the Transgressive Systems Tract B section. Although Lithofacies V (Table 1) crops out again higher in the section, thicknesses of subfacies reported in this section refer only to those sandstones that are here interpreted as Lowstand Systems Tract B.

The basal 1.25 m of Lithofacies V (Table 1) at Mt. Dunfee contains the shale-and-siltstone subfacies (Subfacies Va, Table 5; Figure 19b). The shale beds of this subfacies are typically 2 to 5 cm thick and are parallel laminated. The shales often display small-scale loading structures that form when the overlying silts sink into less dense muds (Figure 19b). These features resemble flame structures, except that these "flames" bend in opposing directions.

Beds of siltstone are typically 5 cm thick at the base of the subfacies and thicken upward to about 10 cm. Siltstone beds are parallel laminated at the base of the section and display hummocks toward the top of the section where shale beds are progressively less abundant and siltstone beds become interbedded with sandstone to form the siltstone-and-sandstone subfacies (Subfacies Vb, Table 5; Figure 19c).

The siltstone-and-sandstone subfacies (Subfacies Vb, Table 5) is 0.75 m thick at Mt. Dunfee and first crops out 1.25 m above the base of Lithofacies V (Table 5). At Hines Ridge, Subfacies Vb comprises the entire 12 m of Lithofacies V. Subfacies Vb (Table 5; Figure 19c) contains dark-brown, parallel-laminated siltstone beds that are typically 0.1 to 0.2 m thick and lenticular in
nature. These siltstone beds are interbedded with sandstone beds similar to those described in the sandstone subfacies except that they display abundant hummocky cross-stratification. The hummocky siltstone beds display undulatory upper and lower surfaces (Figure 19c). Laminae at the crests of the hummocks fan into the trough in a draping fashion and are slightly discordant with the underlying lamina. Heights range from 2 to 7 cm with wavelengths of 10 to 25 cm. At Mt. Dunfee, hummocks become less abundant upsection where tabular cross-beds of the overlying sandstone subfacies (Subfacies Vc, Table 5; Figure 19d) predominate.

The sandstone subfacies (Subfacies Vc, Table 5; Figure 19d) comprises the upper 13 m of the cross-bedded-sandstone lithofacies (Lithofacies V, Table 1) at Mt. Dunfee, the entire thickness of Lithofacies V at Magruder Mountain, where the lithofacies is 17 m thick, and the basal 1 m of Lithofacies V at Loretta Road. Subfacies Vc displays planar cross-bedded sandstone that is composed of very coarse to fine-grained quartz with overgrowths and minor (>1%) feldspar and secondary micas cemented by neomorphosed calcite. The sandstone is typically gray to buff in color on fresh surfaces due to the abundance of calcite cement and weathers reddish-brown.

The beds are laterally continuous and thickness of bedding varies from 0.5 to 2 m and generally increases upsection. Thicknesses of sets and co-sets of cross-strata also increases upsection. Sets of cross-strata are tabular and range from 3 cm thick at the base of the lithofacies to 65 cm thick near the top of
the lithofacies, with co-sets varying from 0.75 to 2 m thick, respectively. Laminae within cross-beds vary from 1 to 5 mm thick. Angle of foreset dip varies from 10° to 69°, generally to the northwest and southeast, although only a few reliable measurements were taken. Crossbedding is the only sedimentary structure recognized in the sandstones at the Loretta Road, Magruder Mountain, and Mt. Dunfee sections.

Lowstand Sandstone Deposit: Depositional Interpretation

As discussed previously, cross-bedding represents the avalanche face of bedforms as they migrate. Tabular cross-bedding, in particular, represents the migration of straight-crested bedforms. It occurs in aeolian, fluvial, lacustrine and shallow-marine environments. However, because of the abundance of calcite cement, the variety of grain sizes and shapes, and the lack of fluvial depositional features, such as scouring, channelization and trough cross-stratification, and the lack of other eolian features, such as distinctive eolian stratification (cf. Kocurek and Dott, 1981), the sandstones in the Deep Spring Formation are interpreted as marine deposits.

Hummocky cross-stratification is commonly associated with the waning flow stages of storm surges. In particular, it is interpreted as the product of dominantly oscillatory flow (Dott and Bourgeois, 1982; Southard and others, 1990) but can form under conditions of superimposed slight unidirectional flows (Nottvedt and Kreisa, 1987; Arnott and Southard, 1990). Hummocky cross-

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stratification has been recognized from a variety of shallow-marine environments, from the tidal flat to innershelf (Dott and Bougeois, 1982), often within the transition zone of fair-weather and storm wave base (Krassay, 1994). The change upsection in the lower Deep Spring Formation from interbedded siltstone and shale to hummocky fine-grained siltstone and sandstone to tabular cross-bedded sandstone suggests a change from a calm depositional environment with periodic high-energy storm events to a dominantly high-energy environment (de Raaf and others, 1977). In addition, similar shoaling hummocky cross-stratified successions have been described by Myrow (1992). In the Deep Spring Formation, this change is interpreted to represent shoaling conditions. Similar upward-shallowing sandstone successions were described by Davis and Byers (1989), Dirks and Norman (1992), and McCormick and Grotzinger (1993). The increasing thickness upsection of beds, and sets and cosets of cross-stratification suggests that relative water depth was shallowing as these sandstones were being deposited (sensu de Raaf and others, 1977). The occurrence of large-scale cross-bedding indicates a high-energy environment, which is common across shallow siliciclastic shelves, and may represent migrating marine sand dunes (Dirks and Norman, 1992).

**Lowstand Systems Tract: Sequence-Stratigraphic Interpretation**

The shale-and-siltstone lithofacies (Lithofacies IX, Table 1) at Mt. Dunfee, is interpreted to record initial basinal deposition of sediment that bypassed the
shallow shelf following a sudden basinward shift in facies resulting from faulting. At Mt. Dunfee, the cross-bedded-sandstone lithofacies (Lithofacies V, Table 1) gradationally overlies the shale-and-siltstone lithofacies (Lithofacies IX, Table 1), and the preserved sedimentary structures within this succession indicate shallowing of water depths upsection. Although parasequences are not readily recognized within these beds, evidence is strong for an overall relative sea-level fall based on the stacking pattern of parasequence sets. Therefore, this upward-shallowing succession of siliciclastic sediment is here interpreted as an upward-shoaling parasequence set. Sedimentary structures are not as well preserved at sections other than Mt. Dunfee. Lithofacies V (Table 1) is interpreted to represent upsection-shallowing parasequence sets at all lower Deep Spring Formation localities. These parasequence sets that stack in an upward-shoaling pattern and are correlative across the platform are typical of lowstand systems tract sedimentation and reflects falling relative sea level.

**Transgressive Systems Tract B**

When sea level rose sufficiently during the transgression to allow carbonate sedimentation to return to the shelf, thin carbonate rocks where deposited in the lower Deep Spring Formation. These thin limestones are the only deposits in the study area that can be attributed to Transgressive Systems Tract B. Thus, this time is interpreted as one of minimal deposition and subsequent relative deepening and steepening of the ramp. Directly overlying Lithofacies V (Table 1) at Mt. Dunfee is a 0.5 m thick package of thinly
interbedded shale, siltstone, and limestone beds. These beds are directly overlain by the contorted-limestone-and-sandstone lithofacies (Lithofacies X, Table 1). The limestone beds are the least common and most poorly preserved beds. They are typically 2-to-3-cm-thick, grainy beds with no internal structure preserved. These limestone beds represent a return to carbonate deposition following the dominance of siliciclastic sedimentation during the lowstand systems tract. The shale and siltstone beds interbedded with the limestone beds are much more common. The shale and siltstone beds are typically 5 to 10 cm thick and often display hummocky cross-stratification. The initial basal limestone bed within the sandstone-and-limestone subfacies (Subfacies Vd, Table 5) of Lithofacies V may also signal the transgressive rise in sea level that cut off the terrigenous siliciclastic sediment supply and allowed carbonate production to reestablish on the shelf. At the Loretta Road section, the upper portion of Lithofacies V (Table 1) is 5 m thick. It consists of lime mudstone beds that are interbedded with cross-bedded sandstone in the sandstone-and-limestone subfacies (Subfacies Vd, Table 5). The limestone beds are typically undulatory but continuous beds of 0.3 to 0.5 m-thick, light-grey lime mudstone. These lime mudstones have a sugary texture, possibly due to recrystallization, and no sedimentary structures were discernable.

Carbonate systems commonly experience a lag following siliciclastic deposition before carbonate production resumes (Schiager, 1981; Read and others, 1986). Therefore, these carbonate sediments probably represent
deposition during the late transgressive systems tract following a sedimentary lag during the early transgressive systems tract. A depositional hiatus during most of the transgressive systems tract would result in deepening across the entire platform. This deepening, combined with the return of carbonate sedimentation, produced enough relief to allow the formation of slumps during Early Highstand Systems Tract B. This relief could have been only a minor steepening of a few degrees of the ramp and was probably not significantly steeper than that which took place during Transgressive Systems Tract A. Although distal steepening of the ramp did take place during the development of Transgressive Systems Tract A, the steepening during Transgressive Systems Tract B was seemingly more spatially extensive, possibly due to the widespread deposition of siliciclastic sediment during Lowstand System Tract B or to faulting that formed Sequence Boundary B. Although siliciclastic sedimentation was widespread, siliciclastic sedimentation rates are much lower than carbonate production rates, as discussed in Chapter 2, and, thus, rate of sea-level rise could easily outpace sediment accumulation. Therefore, the accommodation space would be greater across the shelf following Lowstand Systems Tract B, allowing for a wider distribution of early highstand slump deposits, as discussed below.

**Early Highstand Systems Tract B**

The contorted-limestone-and-sandstone lithofacies (Lithofacies X, Table 1) directly overlies the cross-bedded-sandstone lithofacies (Lithofacies V,
Lithofacies X (Table 1) is interpreted to record the return of carbonate sedimentation during Early Highstand Systems Tract B (Figure 20a) and subsequent slumping down the distally steepened ramp. Because the early highstand slumps are the first evidence of carbonate sedimentation at most measured sections, the source of carbonate production is believed to be located outside of the study area.

**Early Highstand Slumps: Description**

The contorted-limestone-and-sandstone lithofacies (Lithofacies X, Table 1) is composed of highly-contorted, interlaminated to thinly interbedded light-gray lime mudstone and brown very fine-grained quartz sandstone and siltstone. Although this lithofacies is present in all measured sections, the style of slumping varies from large roll-over folds in the southeast at Mt. Dunfee, to highly contorted beds in the northwest in the Inyo Mountains.

At Mt. Dunfee, the slumps crop out as a 3.5-m-thick unit with large, recumbent folds up to 0.5 m thick each (Figure 20b). In outcrop, these folded rocks are dark brown and rather homogenous in composition. Microscopically, however, alternating sandstone and siltstone laminae are discernable and the grains are aligned with the folded bedding. The slumps are tightly folded with interlimb angles typically between 30° and 40°. Bedding is continuous and of fairly equal thickness, although the siltstone beds occasionally thicken in the nose of the fold.
Figure 20a. **Early Highstand Systems Tract:** Interpretive block diagram of deposition during the Early Highstand Systems Tract B. Note the occurrence of slump features, which are interpreted as unique features of mixed carbonate-siliciclastic systems.

Figure 20b. **Early Highstand Slumps:** Large roll-over folds at Mt. Dunfee interpreted as slump beds. Location MD91II04; scale is approximately 25 cm long.

Figure 20c. **Early Highstand Slumps:** Contorted slumps at Hines Ridge. Note the difference in style of slumping as compared to Mt. Dunfee (Figure 20b). Location HR91I96; 40 cm long hammer.
In the northwest, at the Inyo Mountain sections, Lithofacies X (Table 1) crops out as a 1- to 3-m-thick bed of highly contorted, very fine-grained quartz sandstone to siltstone and lime mudstone. Most of the sandstone and siltstone beds are discontinuous, appearing as pods of brown siliciclastic rocks within the light-gray carbonates (Figure 20c). These contorted beds do not display large folds like the deposits in the southeast, but instead contain boudinage-like features. Individual siliciclastic pods average 20 cm in length, although size and shape are highly variable. Carbonate pods are up to 0.3 m in diameter and may represent crudely preserved beds that were subjected to extensive boudinage.

The cross-bedded-sandstone lithofacies (Lithofacies V, Table 1) directly overlies the contorted-limestone-and-sandstone lithofacies (Lithofacies X, Table 1) at Loretta Road, Hines Ridge, and Magruder Mountain. At Loretta Road and Hines Ridge, Lithofacies V (Table 1) is 5 m thick and 3 m thick, respectively. At these two sections, Lithofacies V (Table 1) is represented by the sandstone-and-limestone subfacies (Subfacies Vd, Table 6). At Magruder Mountain, the lower 7 m of Lithofacies V (Table 1) contains the sandstone subfacies (Subfacies Vb, Table 6). This subfacies is subsequently overlain by a very poorly preserved, 11-m-thick succession of Subfacies Vd (Table 6).

At the Mt. Dunfee section, the contorted-limestone-and-sandstone lithofacies (Lithofacies X, Table 1), is overlain by a 5-m-thick succession of poorly preserved, undulatory, lime-mudstone beds. These lime-mudstone beds are believed to be closely related to the beds of contorted limestone and
sandstone that underlie them, and may represent the undisturbed beds that are typically associated with syndepositional slump formations.

**Early Highstand Slumps: Depositional Interpretation**

The mechanism of formation for these slumps is believed to be similar to those in Early Highstand Systems Tract A. However, the major difference between these slumps and the first set (Lithofacies VIII, Table 1) is the presence of siliciclastic beds within these slumps. As discussed previously, the sandstones deposited during Transgressive Systems Tract A were limited in their distribution and never reached the deeper areas of the ramp. Siliciclastic sediments were dominant across the shelf during deposition of Lowstand Systems Tract B, prior to the second set of slumps. Thus, the second set of slumps contain a mixed composition.

The differences in slumping style between the shallower water Inyo Mountain sections and deeper water Mt. Dunfee may be explained by their location in relation to the paleoslope (cf. Naylor, 1981; Myrow and Hiscott, 1991). Although the orientation of the paleoslope is unknown, the Inyo slumps look like slumps that typically form on the upper ramp where deformation is limited because only minor movement of the sediment takes place due to the short transport distance and lack of momentum. The Mt. Dunfee slumps, however, look like slumps that underwent abundant folding, indicating larger-scale movement than that which occurred in the Inyo slumps. This abundant
**TABLE 6. Summary of Early Highstand Systems Tract B**

<table>
<thead>
<tr>
<th>Lithofacies/ Subfacies</th>
<th>Distance above base of lithofacies</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Loretta Road</td>
</tr>
<tr>
<td>Lithofacies X</td>
<td>2.5 m</td>
</tr>
<tr>
<td>Lithofacies V Subfacies</td>
<td>b) sandstone</td>
</tr>
<tr>
<td></td>
<td>d) sandstone and limestone subfacies</td>
</tr>
</tbody>
</table>

**Systems Tract (n\(^{th}\) order)**  
**Relative Sea Level**  
**Falling**  
**Rising**

Parasequence sets (n+1) and Subfacies within the Early Highstand Systems Tract

Letters to the left of sections correspond to subfacies designations shown above. Numeral to the right of sections correspond to lithofacies designations shown in Table 1 and Figure 7. Patterns in sections correspond to rock types shown in Figure 7. Arrows represent n+1 order relative sea-level change.
deformation suggests these slumps were deposited further down the ramp following transport of a greater distance and perhaps greater momentum. The presence of the interbedded shale, siltstone, and limestone beds below the Mt. Dunfee slumps also suggests they were deposited in deeper water than the Inyo slumps.

The presence of subfacies of Lithofacies V (Table 1) above Lithofacies X (Table 1) suggests that siliciclastic sediment was locally being distributed across the shelf. The amount of sand, however, decreases laterally. The greatest volume of sand is seen in the most shoreward area at the Loretta Road section. Sand content gradually decreases toward the most basinward Inyo Mountain section at Magruder Mountain (Figure 7). Almost no siliciclastic sediments are interbedded with limestone at Mt. Dunfee. This suggests that the coarser siliciclastic sediment was becoming trapped shoreward, indicating a relative rise in sea level (Table 6). In addition, the amount of siliciclastic sediment decreases upsection, also indicating a relative sea-level rise (Table 6).

Early Highstand Slumps: Sequence Stratigraphic Interpretation

As with the slumps of Early Highstand Systems Tract A, the slumps of Early Highstand Systems Tract B are interpreted to be a product of early carbonate ramp deposition following a period of siliciclastic dominance during lowstand and transgression. As a result, they are unique signals of a mixed carbonate-siliciclastic system. During deposition of the Early Highstand
Systems Tract B (Figure 20a), an area of carbonate production was reestablished on the shallow shelf outside of the study area providing carbonate sediment that is present in the slump deposits. Part of this carbonate production area may be represented by the limestone beds within Subfacies Vd below Lithofacies X at the Loretta Road section. Carbonate slump features of the contorted-limestone-and-sandstone lithofacies are similar to those in Highstand Systems Tract A. These slumps, however, differ from those in Tract A, in that they were emplaced across the entire study area. This difference seemingly is due to the overall deepening of the shelf, and subsequent increase in accommodation space, that took place because of the relatively slower sedimentation rate of siliciclastic sediments during Lowstand Systems Tract B, and the depositional hiatus of Transgressive Systems Tract B as opposed to the higher carbonate sedimentation rates that took place during Early Highstand Systems Tract A.

**Late Highstand Systems Tract B**

The ooid-limestone lithofacies (Lithofacies VI, Table 1) was deposited directly on the cross-bedded-sandstone lithofacies (Lithofacies V, Table 1) that overlies the slump features in the Inyo Mountains, and is interpreted as an ooid shoal. At Magruder Mountain, this lithofacies overlies Lithofacies III (Table 1). Because ooid formation requires nearly continuous wave or current action (Tucker and Wright, 1990), this lithofacies is believed to have formed during the late highstand systems tract as sea level was beginning to fall (Figure 21). At
Mt. Dunfee, however, the clotted-and-intraclastic-limestone lithofacies (Lithofacies III, Table 1) is overlain by the dolomitized-allochem-conglomerate lithofacies (Lithofacies XI, Table 1) that is interpreted as upward-shallowing deposit of the carbonate-dominated highstand systems tract.

Ooid Shoal: Description

The ooid-limestone lithofacies (Lithofacies VI, Table 1) is present only in the shallow-water sections at Loretta Road, Hines Ridge, and Magruder Mountain (Figure 7). It thins to the northeast, ranging in thickness from 17 m at Loretta Road to only 7 m at Magruder Mountain (Figure 7). The lithofacies displays continuous beds of abundant recrystallized calcite ooids (Figure 22a). In outcrop, the grainstone appears light gray with lenses of pisolithic or oncolitic grains. Relict meter-scale cross-bedding is present although difficult to recognize because of recrystallization. The grainstone is composed of ooids whose diameters range from 0.25 to 0.5 mm and typically appear as circular "ghosts" within larger crystals of equant calcite (Figure 22b). The ooids no longer retain any internal structure. The grains, however, are pure calcite with no apparent siliciclastic grains as nuclei. Stratigraphically, this is the uppermost lithofacies of the lower Deep Spring Formation at the shallow-water Inyo Mountain sections. It is directly overlain by the basal sandstone of the middle Deep Spring Formation at Hines Ridge, Loretta Road, and Magruder Mountain. However, the ooid grainstone (Lithofacies VI, Table 1) is temporally equivalent
Figure 21. **Late Highstand Systems Tract**: Interpretative block diagram of the Late Highstand Systems Tract B.
Figure 22a. **Ooid Shoal**: Thick ooid grainstone at Hines Ridge that represents an ooid shoal. Location HR91I110; field assistant for scale.

Figure 22b. **Ooid Shoal**: Photomicrograph of recrystallized ooid ghost (arrow). Note the lack of internal structure within the ooids. Location HR91I110; bar is approximately 0.1 mm.

Figure 22c. **Highstand Deposits**: Dolomitized allochem conglomerate at Mt. Dunfee. Note the irregular nature of the dark gray alloches in the orange matrix indicating dissolution of clast or stylotization. Location MD91III25; scale in divisions of 5 cm.

Figure 22d. **Highstand Deposits**: Lace-work pattern of silty, dolomitized beds within the limestone indicating dissolution of limestone and precipitation of ferroan dolomite. Location MD91III31; bar is approximately 5 cm.
to the deeper water dolomitized-allochem-conglomerate lithofacies (Lithofacies XI, Table 1) at Mt. Dunfee, which is described below.

Ooid Shoal: Depositional interpretation

Ooids are a well known constituent in both ancient and modern carbonate environments. They typically form in well agitated, warm, marine waters that are supersaturated with respect to CaCO$_3$, although they have been documented from non-marine environments (see Ginsburg and James, 1974; Milliman, 1974; Bathurst, 1975; and Tucker and Wright, 1990, for more detailed explanations of the formation of ooids). Agitation is necessary to allow the fairly even coating of CaCO$_3$ to precipitate around the nuclei. Internally, ooids may have a variety of structures that may be controlled by a combination of salinity, primary mineralogy, and the energy regime in which they formed (Richter, 1983).

Because Deep Spring Formation ooids have been recrystallized, they no longer provide information on the environment of origin. The lack of terrigenous nuclei, however, indicates formation in a pure carbonate environment. The lack of carbonate mud in the grainstone indicates a high-energy depositional environment regardless of the environment of formation (Dunham, 1962). If the ooid deposit faces open water that is at least 10-m deep, wave energy is usually enough to remove any mud (Halley and others, 1983).

As stated above, steady agitation is necessary for the formation of ooids. This agitation typically takes place in areas of constant wave or tidal activity, most commonly in tidal channels or shoals that fringe carbonate platforms. In
the Bahamas, modern ooids form most readily at water depths of 2 to 5 m, which seems to provide maximum sediment motion (Hine, 1983). Ooid shoals can also migrate actively during relative changes in sea level if they do not become stabilized by marine flora (Halley and others, 1983; Hine, 1983). Features such as light color, abundant ooids, medium- to large-scale crossbedding, lack of fossils, thickness and extent of the deposit, and facies associations are used to recognize ooid shoals (Sellwood, 1981; Halley and others, 1983). Because of its (1) thickness; (2) cross-stratification, although poorly preserved in the lower Deep Spring Formation; (3) association with marine sediments; (4) pure carbonate composition regardless of its association with siliciclastic sediment, which suggests migration; and (5) lateral association with deep-water deposits seaward (Lithofacies XI, Table 1; discussed below) and inferred shallow-marine deposits landward, the ooid grainstone facies appears to represent an ooid shoal.

Highstand Deposits: Description

The dolomitized-allochem-conglomerate lithofacies (Lithofacies XI, Table 1) includes mottled dark-gray and orange limestone interbedded with thin-bedded limestone and shale. Upsection, the thinner beds gradually give way to thicker, coarser, more mottled limestone beds. Therefore, it is divided into an allochem-rich-limestone-and-shale subfacies (Subfacies XIa, Table 7), and an allochem-rich-limestone subfacies (Subfacies XIb, Table 7).
TABLE 7. Summary of Late Highstand Systems Tract B

<table>
<thead>
<tr>
<th>Lithofacies XI: Dolomitized Allochem Conglomerate</th>
<th>Distance above base of lithofacies</th>
</tr>
</thead>
<tbody>
<tr>
<td>Subfacies</td>
<td>Loretta Road</td>
</tr>
<tr>
<td>a) allochem-rich limestone and shale</td>
<td>N/A</td>
</tr>
<tr>
<td>b) allochem-rich limestone</td>
<td>N/A</td>
</tr>
<tr>
<td>Lithofacies VI</td>
<td>0-17 m</td>
</tr>
</tbody>
</table>

Systems Tract (n<sup>th</sup> order) Relative Sea Level Falling Rising

Parasequence sets (n+1) and Subfacies within Late Highstand Systems Tract B.

Letters to the left of sections correspond to subfacies designations shown above. Numeral to the right of sections correspond to lithofacies designations shown in Table 1 and Figure 7. Patterns in sections correspond to rock types shown in Figure 7. Arrows represent n+1 order relative sea-level change.
The allochem-rich-limestone-and-shale subfacies comprises the lower 10 m of Lithofacies XI (Table 1) and contains an interbedded succession of allochem-rich limestone, shale and some dolomite. Beds are typically less than 0.3 m thick. Microscopically, the allochem-rich beds contain disorganized intraclastic and grainy material at the base of the bed that sometimes grades into a parallel-laminated zone at the top of the bed. These disorganized to graded beds are capped by 3- to 10-cm-thick, planar greenish-black siltstone and shale. Rarely, the siltstone caps display a rippled or hummocky upper surface. Although this tripartite succession of homogenous allochem-rich limestone, parallel-laminated lime mudstone, and parallel laminated-siltstone is present within some beds, it is more common for one of the components to be missing. Small shelly fossils are present as grains within these beds. Poor preservation and extreme secondary alteration of the beds, however, makes confirmation and identification difficult.

The allochem-rich limestone subfacies (Subfacies Xlb, Table 7), which overlies Subfacies XIa (Table 7) and makes up the remaining 16 m of the lithofacies, is distinguished from Subfacies XIa (Table 7) by the thickening of bedding (from 0.1 m to 0.6 m), the increase in the size and number of allochems (from 3 mm to 50 mm), and the decrease in the amount of shale and siltstone in the upper subfacies; these upsection changes are gradational. The mottled appearance is a result of clasts of dark-gray limestone set within the orange limestone (Figure 22c). The orange sediment acts as a matrix for the large dark-gray grains, which contain smaller intraclastic grains. These intraclasts are
similar to the intraclasts found at the disorganized base of the graded beds of Subfacies XIa (Table 7). Because of the extent of secondary alteration of the beds, it is difficult to tell how, or if, the orange and dark gray grains were originally related. These mottled beds typically grade into alternating orange and brown, planar, thickly laminated to very thinly bedded limestones. The brown laminae often contain dolomite and minor amounts of siliciclastic silt; thus, they stand in relief due to differential weathering. These laminae repeatedly display a "lacework" pattern such that the brown laminae connect at various intervals down through the orange laminae (Figure 22d). As bedding thickens, the brown dolomite laminae become less common and the limestone clasts become coarser.

Highstand Deposits: Depositional Interpretation

Based on the few sedimentary structures preserved within Lithofacies XI, it is interpreted to represent an upward-shallowing succession. Although lithologically similar to the thin-bedded shale and siltstone beds directly below the slump deposits, the fewer shale beds and the abundance of limestone beds indicates a shallower depositional environment. Although shallower, the presence of structureless limestone beds interbedded with thin-bedded, hummocky shale and siltstone beds suggests the succession was not within fair-weather wave base all the time. Instead, it is interpreted to represent deposition in an environment within the transition between fair-weather and storm wave base. In addition, the crude cyclic nature of bedding in Subfacies XIa that
alternates between limestone beds and siliciclastic beds is here interpreted as a succession of parasequences. The gradational change in the parasequence stacking pattern from Subfacies XIa to Subfacies Xlb (Table 7) manifests itself as an upsection increase in the amount of the thicker and coarser mottled beds, and an upsection decrease in siliciclastic sediment. This change in parasequence stacking pattern is interpreted as an upward-shoaling succession that is evidence of falling relative sea level. The increase in grains upsection, part of which are small shelly fossils, also suggests moderately shallow water depths.

Ooid Shoal and Highstand Deposits: Sequence-Stratigraphic Interpretation

As relative sea level was beginning to fall during the late highstand systems tract, an ooid shoal developed where wave action was sufficient to agitate shallow areas. This shoal was part of a highly productive carbonate platform that developed during Late Highstand Systems Tract B. Initial deposition at Mt. Dunfee was below fair-weather wave base, but was often within storm wave base. The later coarser and thicker limestone deposits at Mt. Dunfee suggest upward shoaling, just as the ooid shoal does, indicating a relative fall in sea level.

Sequence Boundary C

The third and final sequence boundary recognized in this study occurs at the contact between the lower and middle members of the Deep Spring
Formation (Figure 23a). The best evidence for Sequence Boundary C is a highly irregular dolomitized zone at the top of the ooid-limestone lithofacies (Lithofacies VI, Table 1; Figure 23b) and pockets of red cement (Figure 23c). The dolomitized zone has a highly undulatory contact with the ooid-limestone lithofacies, thus the zone varies in thickness from about 0.25 m to 1.5 m. Due to its irregular, diagenetically altered appearance, this zone is interpreted as a dolomitization front that formed during the sea-level lowstand. The zone is then overlain directly by the sandstones of the middle member of the Deep Spring Formation, which probably represent a lowstand or transgressive systems tract. The top of the lower member at Mt. Dunfee is fault bounded, therefore no evidence of the sequence boundary is present there.
Figure 23a. **Sequence Boundary:** Interpretive block diagram of Sequence Boundary C.

Figure 23b. **Sequence Boundary:** Evidence for subaerial exposure at the top of the ooid shoal. Dolomitization front at Magruder Mountain, arrow at edge of front. Location MM911102; bar is approximately 0.5 m.

Figure 23c. **Sequence Boundary:** Evidence for subaerial exposure at the top of the ooid shoal. Small scale dissolution and red cements (arrow). Location HR91114; 6 cm diameter lens cap.
CHAPTER 4

BIOSTRATIGRAPHY OF THE LOWER DEEP SPRING FORMATION AND IMPLICATIONS FOR THE PLACEMENT OF THE PRECAMBRIAN-CAMBRIAN BOUNDARY

Because of the important events that took place during the Precambrian-Cambrian transition, a great deal of emphasis is placed on defining the boundary and locating a type section for it. Recently, the type section for the Precambrian-Cambrian boundary was defined in the Chapel Island Formation of Newfoundland (Landing, 1992). The boundary designation is based on the first occurrence of the trace fossil *Phycodes pedum* (Landing, 1992). Despite the adoption of a formal boundary stratotype section (Landing, 1994), several problems still exist in global correlation of the Precambrian-Cambrian boundary. Because the Chapel Island Formation is a predominantly siliciclastic deposit that has undergone low-grade metamorphism, chemostratigraphic and magnetostratigraphic signatures are difficult to obtain (Brasier and others, 1992). Stratigraphic analyses of the Chapel Island Formation (Landing and Benus, 1988; Myrow and Hiscott, 1991; Myrow, 1992; Myrow and Landing, 1992), have concentrated on depositional environments but place little emphasis on sequence stratigraphy because of the dominance of shales and the subsequent
presence of correlative conformities instead of sequence boundaries.

Paleontologic data, especially small shelly fossil data, is sparse and poorly preserved in the Chapel Island Formation (Brasier and Cowie, 1989). Although correlations between successions in eastern North America and Europe have been fairly successful, correlations are much more difficult with successions of similar age in Antarctica, Australia, and western North America because of poor outcropping and preservation of the rocks. All of these facts indicate that well understood reference sections will play a critical role in global correlation of the Precambrian-Cambrian boundary (Knoll and Walter, 1992). It is especially important that some of these reference sections were deposited during or following the proposed breakup of western Laurentia and the Australia-Antarctic shield (Hoffman, 1991; Moores, 1991) because these areas have typically been excluded from studies correlating with the global stratotype in Newfoundland.

This study suggests the lower Deep Spring Formation should be considered as a possible reference section for the Precambrian-Cambrian boundary in western North America because it meets many of the criteria necessary for boundary recognition. As discussed in previous chapters, the lower member of the Deep Spring Formation is bounded by sequence boundaries and contains a third. These boundaries represent surfaces that, perhaps after further study, may be correlated with other such surfaces on a regional and perhaps global scale. A chemostratigraphic study of these rocks, utilizing stable isotopes, has revealed potentially large gaps in the rock record.
within the Vendian and Tommotian zones when compared to stratigraphic sections in Siberia (Corsetti, 1993). Sequence boundaries in the Deep Spring Formation could provide a viable stratigraphic explanation for this missing rock record.

The Deep Spring Formation also contains a limited, although important, record of body and trace fossils, the stratigraphic locations of which are shown in Figure 24. The Deep Spring is one of the few locations in North America that contains a small shelly fossil assemblage. Small shelly fossils represent the first recorded attempt by metazoans to produce skeletal hard parts (Bengtson, 1988). The reasons proposed for this production vary widely, partly because the small shelly fossils are so poorly understood (Bengtson, 1988; Conway Morris, 1988; Jiang, 1988; Grant, 1990). However, for a brief time, they appeared world wide, which makes them useful in intercontinental correlation (Grant, 1990). Therefore, the assemblage recognized in this study may be valuable in global correlation and continental reconstruction.

**Trace Fossils**

Trace fossils are present only in the shale-and-siltstone lithofacies (Lithofacies IX, Table 1) in the lower Deep Spring Formation. Recognized genera include *Planolites* or *Palaeophycus*, *Scolicia*, *Protopaleaeodictyon*, and a *Bergaueria*-like trace (*sensu* Hantzschel, 1975) (Figure 25, 26). *Planolites* and *Palaeophycus* are both cylindrical, smooth-walled, sinuous burrows, that are
Figure 24. Simplified stratigraphic column from Mt. Dunfee showing the location of small shelly fossils and trace fossils within Lithofacies IV and IX.
Figure 25a. **Trace Fossils:** Thin form of the trace fossil *Planolites* or *Palaeophycus* (arrows). Location MD91II30; scales shown.

Figure 25b. **Trace Fossils:** Thick form of the trace fossil *Planolites* or *Palaeophycus* (arrows). Location MD91II30; scales shown.

Figure 25c. **Trace Fossils:** Large form of the trace fossil *Scolicia* (arrows). Location MD91II30; scale shown.

Figure 25d. **Trace Fossils:** Small form of the trace fossil *Scolicia* (arrows). Location MD91II30; scale shown.
Figure 26a. **Trace Fossils:** Trace fossil *Protopalaedictyon* (arrow). Location MD91II30; scale shown.

Figure 26b. **Trace Fossils:** Mold of the *Bergaueria*-like resting mark. Location MD91II30; scale shown.

Figure 26c. **Trace Fossils:** Cast of the *Bergaueria*-like resting mark. Location MD91II30; scale shown.
typically unbranched but can exhibit occasional branching (Hantzschel, 1975) (Figure 25a, b). In the Deep Spring Formation, these traces are typically 2 mm to 4 mm in diameter with occasional branching. The two genera are often confused because of their similarities. However, *Planolites* burrows, unlike *Palaeophycus*, are lithologically different from the host rock. This difference in lithology results from sediment passing through the worm creating *Planolites*, whereas *Palaeophycus* is the trace of a passive feeder and does not result in any lithologic differences. However, poor preservation can make the distinction between these two genera difficult.

*Scolicia* is believed to be the creeping or feeding trail of gastropods (Hantzschel, 1975). *Scolicia* traces identified in the lower Deep Spring Formation are ridgelike to ribbonlike, ribbed trails with a median axis (Figure 25c, d). Their size varies from 1 mm to 4 mm in width with meandering trails over 10 cm in length. *Protopalaeodictyon*, also identified in the lower Deep Spring Formation, displays a horizontal, highly branching, irregular polygonal pattern that represents a meandering trail (Hantzschel, 1975) (Figure 26a). The poorly preserved trail is 1 mm in diameter and about 5 cm long with common branching of 5 mm in length. The *Bergaueria*-like fossil in the Deep Spring Formation is a 3.5 cm wide, slightly oblong 1.5 cm deep depression with a second 2 cm wide depression in the center of the outer depression (Figure 26b, c). *Bergaueria* is believed to represent the resting trace of suspension feeding anemones (Alpert, 1973).
Small Shelly Fossils

The presence of small shelly fossils in the lower Deep Spring Formation is important for a variety of reasons. First, this formation is one of the few places in North America, and particularly in the Basin and Range, where these fossils have been recognized, which makes the Deep Spring Formation an excellent candidate for global correlation. Secondly, the assemblage found in the Deep Spring is quite unusual, in that it contains an assemblage of small shelly fossils that, prior to this study, had never been found together.

Although several of the small shelly fossils reported in this study have yet to be identified at genus level, at least three have been classified, and confirmed by P. W. Signor (personal communication, 1993). These include *Cloudina*, *Nevadatubulus*, and *Sinotubulites* (Figure 27).

*Cloudina* is a tube-shaped organism with a cone-in-cone structure that produces a thin-walled, asymmetric circle-in-circle cross section (Grant, 1990) (Figure 27a). *Nevadatubulus* is also tube shaped with a fairly symmetric although irregular thick-walled cross-section (Signor and others, 1987) (Figure 27c). *Sinotubulites* exhibits longitudinal sculpturing on the tube, thus producing a regularly ornamented cross-section (Signor and others, 1987) (Figure 27b).

Several other fossil forms were observed in the lower Deep Spring Formation (Figure 28), but they were not identified to genus level. One fossil, a large, thick walled form (Figure 28a) has also been seen in rocks of correlative
Figure 27a. Small Shelly Fossils: Photomicrograph of the small shelly fossil *Cloudina* in plane polarized light using a whitecard. Location MD911107; bar is 1 mm.

Figure 27b. Small Shelly Fossils: Photograph of the small shelly fossil *Sinotubulites*. Photograph from P. Signor; bar is 1 mm.

Figure 27c. Small Shelly Fossils: Photomicrograph of the small shelly fossil *Nevadatubulus* in plane polarized light using a whitecard. Location MD911101; bar is 0.1 mm.
Figure 28a. **Small Shelly Fossils:** Photomicrograph of an unidentified thick walled fossil. Location is MD91II05; bar is 1 mm.

Figure 28b. **Small Shelly Fossils:** Photomicrograph of an unidentified multichambered fossil or clast containing several small fossils. Location is MD91II07; bar is 0.5 mm.

Figure 28c. **Small Shelly Fossils:** Photomicrograph of an unidentified long, thick walled fossil. Location is MD91II03; bar is 0.5 mm.

Figure 28d. **Small Shelly Fossils:** Photomicrograph of an unidentified long, highly sculptured fossil. Location is MD91II03; bar is 0.5 mm.
age in Namibia (B. Saylor, personal communication, 1994). Another form represents either a multichambered organism, or a clast containing several small organisms (Figure 28b). Finally, a long thick-walled organism (Figure 28c), and a long, highly sculptured form (Figure 28d) were also observed.

Neoproterozoic-Cambrian Boundary Placement

*Cloudina* has been suggested as a Proterozoic index fossil (Grant, 1990). It typically is found alone. Grant (1990) suggested that *Cloudina* was either co-generic with, or closely related to, *Sinotubulites* and *Nevadatubulus* reported in the Deep Spring Formation. However, he never actually recognized *Cloudina* in the lower Deep Spring. *Cloudina* was reported in terminal Proterozoic rocks from Namibia, Brazil, Spain, China, Oman, Argentina and Antarctica, thus encouraging its use as an index fossil (Conway Morris and others, 1990; Grant, 1990). Landing (1994), however, believed that if *Cloudina* was a Proterozoic index fossil, it occurred alone.

*Nevadatubulus* and *Sinotubulites* were identified previously in the lower Deep Spring Formation by Signor and others (1987), and they were considered Cambrian in age (Signor and others, 1987). The co-occurrence, documented in this study, of *Cloudina* with *Nevadatubulus* and *Sinotubulites*, as well as unidentified forms suggests two possible options: (1) the range of *Cloudina* is longer, extending into the range of *Nevadatubulus* and *Sinotubulites* (i.e. lower Cambrian); or (2) the Neoproterozoic assemblage was more diverse than
previously recognized, such that the range of *Nevadatubulus* and *Sinotubulites* extends down into the range of *Cloudina*. Depending on which of these possibilities is correct and how the criteria are defined for the placement of the Cambrian boundary, the Neoproterozoic-Cambrian boundary may be identified in the lower Deep Spring Formation.

The world-wide association of *Cloudina* with Ediacaran fauna (Grant, 1990) is strong evidence for a terminal Proterozoic age for *Cloudina* and the Deep Spring Formation. Although no Ediacaran fossils have been recognized in the Deep Spring Formation, the Ediacaran fossil *Ernietta* has been recognized in the lower Wood Canyon Formation (Horodyski and others, 1994). Thus, it is possible that these rocks may be equivalent to the lower Deep Spring Formation, suggesting at least some vague temporal equivalence of *Cloudina* with *Ernietta*. Additionally, *Phycodes pedum*, the trace fossil selected to mark the Precambrian-Cambrian boundary, first occurs well within the upper member of the Deep Spring Formation (Crimes, 1989). Assuming *Phycodes pedum* is not facies controlled, then the lower member is Neoproterozoic in age. This argument is further strengthened because traces are present in the lower Deep Spring but *Phycodes pedum* is not among them. Figure 29 shows the world-wide last appearance of *Cloudina* and its relation to the first occurrence of trilobites, *Phycodes*, and Ediacaran-aged fauna. In all cases, the last appearance of *Cloudina* is prior to the first occurrence of either trilobites or *Phycodes*.
Figure 29. Simplified stratigraphic columns showing the stratigraphic relationship between *Cloudina*, *Phycodes*, trilobites, Cambrian-aged traces and Ediacaran fauna. Note that *Cloudina* is never found stratigraphically higher than *Phycodes*, trilobites, or Cambrian-aged traces. Heavy lines represent formation boundaries, thin lines represent member boundaries within formations. Modified from Grant, 1990.

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This biostratigraphic information, combined with the sequence-stratigraphic framework provided by this study and the chemostratigraphic data of Corsetti (1993) and Corsetti and Kaufman (1994), suggest that the Neoproterozoic-Cambrian boundary lies within the Deep Spring Formation. In addition, a new sinuous, agglutinated tube has been reported in the middle member of the Deep Spring Formation, and it is believed to be the oldest Cambrian fossil in western North America (Signor and others, 1994). This finding suggests that the Neoproterozoic-Cambrian boundary should lie between the last occurrence of *Cloudina* and the first occurrence of the agglutinated tube in the Deep Spring Formation. This placement disagrees with the placement suggested by some previous studies (Alpert, 1977; Signor and Mount, 1986; Crimes, 1989). A summary of previous placements of the Neoproterozoic-Cambrian boundary in the Deep Spring Formation, as well as the proposed placement from this study is presented in Figure 30. Alpert (1977) based his placement of the boundary on the first occurrence of trilobite trace fossils in the upper member of the Deep Spring Formation as opposed to the simple traces that are found in underlying rocks. Signor and Mount (1986), suggested the boundary be placed at the contact between the Wyman Formation and the Reed Dolomite. Their placement was based on the occurrence of possible recrystallized fossils in the lower member of the Reed Dolomite and the definitive occurrence of *Wyattia* in the upper member of the Reed Dolomite, both occurring above a regional unconformity at the Wyman-Reed contact. Crimes
Figure 30. Simplified stratigraphic sections showing the location of the Neoproterozoic-Cambrian boundary in the Inyo Mountains and the evidence used to determine that location based on previous studies. Also shown are how the previously suggested locations relate to the proposed location based on this study and the work of Signor and others (1994).

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(1989) placed the boundary at the Reed-Deep Spring contact. He recognized the occurrence of *Phycodes* and other complex traces (his Zone II-III traces) in the upper member of the Deep Spring Formation. In addition, he argued that no suitable subfacies were present for these complex traces in the lower members, otherwise they would have been found there as well.

Brasier and Cowie (1989) placed the boundary at the contact between the lower and middle members of the Deep Spring Formation. To reach this conclusion, they compared the traces found within the Deep Spring Formation with the trace fossil zonations of China. This study provides body fossil evidence to support Brasier and Cowie's (1989) placement based on trace fossil zonations. Also, the placement of the boundary as shown by Brasier and Cowie (1989) falls within the range suggested by this study.
CHAPTER 5

CONCLUSIONS

The Precambrian-Cambrian boundary interval represents a time of profound change on Earth. Although a type section for the boundary has been identified in Newfoundland (Landing, 1992), recognition of reference sections is necessary in order to facilitate global correlation. The sequence stratigraphy and biostratigraphy provided by this study, combined with the chemostratigraphic studies of Corsetti (1993), indicate that the Deep Spring Formation should be considered as one of these reference sections.

Sequence-stratigraphic analysis of the Deep Spring Formation reveals three sequence boundaries (Figure 8): at the contact between the lower member and the Reed Dolomite; within the lower member; and at the contact between the lower and middle members of the Deep Spring Formation. This analysis also strongly suggests a period of faulting during the deposition of the Deep Spring Formation that may be the cause of Sequence Boundary B. In addition, because the formation represents deposition in a mixed carbonate-siliciclastic environment, a feature unique to mixed systems was recognized through sequence-stratigraphic analysis. This study proposed the term early highstand slumps for slumped bedding features that occurred in the Early
Highstand Systems Tract. This feature was a result of the interplay between the migration patterns of siliciclastic and carbonate facies. Thus, this feature is unique to mixed carbonate-siliciclastic systems, and as such, may be useful to others working in these distinctive systems.

Trace fossils in the lower member include *Planolites* or *Palaeophycus*, *Scolicia*, *Protopalaeodictyon*, and a *Bergaueria*-like trace. The small shelly fossil *Cloudina*, a typically Proterozoic form, co-occurs with the traditionally Cambrian forms *Nevadatubulus* and *Sinotubulites*, and other unidentified forms. This co-occurrence suggest the Neoproterozoic was more diverse than previously thought, and the biostratigraphy of the Deep Spring Formation needs revising. Also, because *Phycodes pedum*, the trace fossil chosen to designate the Precambrian-Cambrian boundary, does not occur until the upper member of the formation (Crimes, 1989), the lower Deep Spring Formation appears to be latest Neoproterozoic in age. This information, combined with the recognition of a new Cambrian body fossil in the middle member (Signor and others, 1994) and the chemostratigraphy of previous workers (Corsetti, 1993; Corsetti and Kaufman, 1994), suggests the Neoproterozoic-Cambrian boundary may lie within the lower Deep Spring Formation after the final occurrence of *Cloudina* in the lower member and prior to the first occurrence of Cambrian body fossil in the middle member.
APPENDIX I

LOCATION OF MEASURED SECTIONS OF LOWER DEEP SPRING FORMATION

Measured sections are represented on each photocopied map with a section line. Latitude and longitude are referenced on each map so that individuals who wish to locate the measured section can compare these maps with the topographic quadrangle referenced in each set of directions. The following quadrangles were used in this study:

Hines Ridge: Waucoba Mtn. SE, Quadrangle, California 7.5 minute series orthophotoquad, 1976.

Loretta Road: Waucoba Mtn. NE, Quadrangle, California 7.5 minute series orthophotoquad, 1976.

Mt. Dunfee: Gold Point Quadrangle, Nevada-Esmeralda Co., 7.5 minute series topographic map, 1986.

Directions to Hines Ridge Section

From Las Vegas, take Interstate 95 north to Highway 266 (Lida Junction).
Take Highway 266 to Highway 168 (Junction in Oasis, CA).
Take Highway 168 through Westgard Pass towards Big Pine, CA.
Just before the junction with Highway 395 in Big Pine, CA, turn off onto Death Valley Road.
From the junction with Death Valley Road, drive 11.3 miles to Papoose Flat Road sign (past the
#36 mark on the road, just before the #37 mark on the road); turn right.
Take road 0.2 miles to the Papoose Flat sign, veer left.
Take this road 1.1 miles, then take road that veers right (if you continue straight, a good campsite is
available 1.2 miles ahead on the right).
Take this road 1.0 miles and take the road off to the right; at the fork, veer right.
Continue on this road 0.9 miles to the top of the knoll (this will require 4-wheel-drive vehicle).
Facing north, the section begins at the base of the saddle between the light colored Reed Dolomite
and darker, brown Deep Spring Formation. (Shown on Waucoba Mtn. SE, Quadrangle, California
7.5 minute series orthophotoquad, 1976)

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Directions to Loretta Road Section

From Las Vegas, take Interstate 95 north to Highway 266 (Lida Junction).
Take Highway 266 to Highway 168 (Junction in Oasis, CA).
Take Highway 168 through Westgard Pass towards Big Pine, CA.
Just before the junction with Highway 395 in Big Pine, CA, turn off onto Death Valley Road.
From the junction with Death Valley Road, drive 15.9 miles (4.5 miles beyond the Hines Ridge turnoff). The section is on the right (east) side of the road. (Shown on Waucoba Mtn. NE, Quadrangle, California 7.5 minute series orthophotoquad, 1976)
Directions to Mt. Dunfee Section

From Las Vegas, take Interstate 95 north to Highway 266 (Lida Junction).
Take Highway 266 7.2 miles to the junction with Highway 774 (a corral will be to your left).
Take Highway 774 7.6 miles into Goldpoint, NV.
Turn left on 2nd Ave. in Goldpoint, NV.
You will encounter a series of forks in the road. Veer left at the first fork, right at the second fork, and right at the T in the road.
Continue down this road approximately 0.6 miles and take road to the left.
Drive 2.5 miles and turn off onto small road at small volcanic outcropping (called Buffalo Rock, marked on map with an arrow).
Turn right behind rock and continue on trail 0.5 miles to campsite.
It is necessary to hike the rest of the way into section. (Shown on Gold Point Quadrangle, Nevada-Esmeralda Co., 7.5 minute series topographic map, 1986)
Directions to Magruder Mountain Section

From Las Vegas, take Interstate 95 north to Highway 266 (Lida Junction).
Take Highway 266 7.2 miles to the junction with Highway 774 (a corral will be to your left).
Take Highway 774 7.4 miles into Goldpoint, NV; turn right onto Lida Road.
At the fork, veer left toward the Tule Canyon Mine (sign).
Continue straight on this road, veering away from the Tule Canyon Mine for 6.9 miles.
At 6.9 miles, you will cross a road, continue on straight. The road now becomes the State Line Spring Road (as shown on map).
Continue 2.9 miles to the top of the State Line Spring Road.
It is necessary to hike the rest of the way into section. (Shown on Magruder Mt. Quadrangle, Nevada-Esmeralda Co., 7.5 minute series topographic map, 1987)

Alternate Directions:
From Las Vegas, take Interstate 95 north to Highway 266 (Lida Junction).
Take Highway 266 19.2 miles into Lida, NV and turn left.
Take road to left 5.9 miles to the State Line Spring Road and turn right.
Take State Line Spring Road 2.9 miles to the top of road.
It is necessary to hike the rest of the way into section. (Shown on Magruder Mt. Quadrangle, Nevada-Esmeralda Co., 7.5 minute series topographic map, 1987)
REFERENCES CITED


Bridge, J.S., and Best, J.L., 1988, Flow, sediment transport and bedform dynamics over the transition from dunes to upper-stage plane beds: Implications for the formation of planar laminae: Sedimentology, v. 35, p. 735-763.


Cheel, R.J., 1990a, Horizontal lamination and the sequence of bed phases and stratification under upper-flow-regime conditions: Sedimentology, v. 37, p. 517-529.

Cheel, R.J., 1990b, Discussion: Flow, sediment transport and bedform dynamics over the transition from dunes to upper-stage plane beds: Implication for the formation of planar laminae: Sedimentology, v. 37, p. 549-551.


Reproduced with permission of the copyright owner. Further reproduction prohibited without permission.


Greene, L.H., 1982, Cyclic sedimentation within the Upper Member of the Deep Spring Formation (Lower Cambrian), eastern California and western Nevada: The anatomy of a grand cycle [unpublished master's thesis]: University of California at Davis, 198 p.


Reproduced with permission of the copyright owner. Further reproduction prohibited without permission.


Knebel, H.J., and Carson, B., 1979, Small-scale slump deposits, Middle Atlantic continental slope, off eastern United States: Marine Geology. v. 29, p. 221-236.


Naylor, M.A., 1981, Debris flow (olistostromes) and slumping on a distal passive continental margin; the Palombini limestone-shale sequence of the Northern Apennines: Sedimentology, v. 28, p. 837-852.


Vail, P.R., Mitchum, R.M., Jr., and Thompson, S., 1977, Seismic stratigraphy and
global changes of sea level, part 4: Global cycles of relative
changes of sea level. In Payton, C.E., (ed.), Seismic Stratigraphy-
Applications to Hydrocarbon Exploration: American Association of

Vail, P.R., Audemard, F., Bowman, S.A., Eisner, P.N., and Perez-Cruz, C., 1991,
The stratigraphic signatures of tectonics, eustasy, and
sedimentology- an overview, in Einsele and others, (eds.), Cycles

Van Wagoner, J.C., 1985, Reservoir facies distribution as controlled by sea-
level change (abs.). Society of Economic Paleontology and

Van Wagoner, J.C., Posamentier, H.W., Mitchum, R.M., Vail, P.R., Sarg, J.F.,
Loutit, T.S., and Hardenbol, J., 1988, An overview of the
fundamentals of sequence stratigraphy and key definitions, in,
Wilgus, C.K., Hastings, B.S., Ross, C.A., Posmentier, H.W., Van
Wagoner, J., Kendall, and Christopher, G.St.C., (eds.), Sea-Level
Changes- an Integrated Approach: SEPM Special Publication
No. 42, p. 39-45.

The Corrubedo beach-lagoon complex, Galicia, Spain: Dynamics,
sedimentation, and recent evolution of a mesotidal coastal

Walker, J.D., Klepacki, D.W., and Burchfiel, B.C., 1986, Late Precambrian
tectonism in the Kingston Range, southern California: Geology,

Walker, R.G., 1990, Facies modeling and sequence stratigraphy: Journal of
Sedimentary Petrology, v. 60, p. 777-786.

Ward, W.C., Weidi, A.E., and Back, W., 1985, Geology and Hydrology of the

Watts, W.A., 1982, Response of biotic populations to rapid environmental and
climatic changes, in Markgraf, V., Brubaker, L.B., and Chernicoff,
S.E., (eds.), Character and Timing of Rapid Environmental and
Climatic Changes: American Quaternary Association Conference,


