# Sedimentologic Analysis of Cores from the Upper Triassic Chinle Formation and the Lower Permian Cutler Formation, Lisbon Valley, Utah

U.S. GEOLOGICAL SURVEY BULLETIN 2000–E

**On the Front Cover:** View south toward La Sal Mountains along Colorado River between Cisco and Moab, Utah. Fisher Towers in center are composed of Permian Cutler Formation and capped by Triassic Moenkopi Formation. Prominent mesa at left center is capped by Jurassic Kayenta Formation and Wingate Sandstone and underlain by slope-forming Triassic Chinle and Moenkopi Formations. The Chinle-Moenkopi contact is marked by a thin white ledge-forming gritstone. Valley between Fisher Towers and Fisher Mesa in background is part of Richardson Amphitheater part of Professor Valley. Photograph by Omer B. Raup, U.S. Geological Survey.

# Sedimentologic Analysis of Cores from the Upper Triassic Chinle Formation and the Lower Permian Cutler Formation, Lisbon Valley, Utah

*By* Russell F. Dubiel *and* Janet L. Brown

EVOLUTION OF SEDIMENTARY BASINS—PARADOX BASIN
A.C. Huffman, Jr., Project Coordinator

U.S. GEOLOGICAL SURVEY BULLETIN 2000–E

*A multidisciplinary approach to research studies of sedimentary rocks and their constituents and the evolution of sedimentary basins, both ancient and modern*

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# METRIC CONVERSION FACTORS

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# Sedimentologic Analysis of Cores from the Upper Triassic Chinle Formation and the Lower Permian Cutler Formation, Lisbon Valley, Utah

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## ABSTRACT

Five uranium exploration cores from Lisbon Valley in the Paradox Basin of southeastern Utah provide examples of sedimentary structures and lithofacies from the Lower Permian Cutler Formation and the overlying Moss Back Member of the Upper Triassic Chinle Formation. The Cutler Formation consists of reddish-brown to purple, arkosic sandstone, siltstone, and mudstone of fluvial and floodplain origin interbedded with reddish-orange sandstone and mudstone deposited in sabkha, eolian dune, and sand-sheet settings. The colors are indicative of the respective depositional settings. An erosional unconformity separates the top of the Cutler from the overlying Moss Back Member of the Chinle Formation. The Moss Back consists of greenish- to bluish-gray limestone-nodule conglomerate, siliciclastic sandstone, and siltstone deposited in high-energy fluvial channels and crevasse splays and on adjacent levees and floodplains. The drab colors of the Moss Back reflect the high organic-carbon content of strata deposited and preserved below the water table. These five cores record eolian and fluvial sedimentary structures and lithofacies sequences not well preserved in outcrops, they provide the basis for interpretation and comparison of depositional environments from lithofacies analysis in core and outcrop studies, and they establish the sedimentologic background for future petrographic and geochemical research on cores and outcrops from the Lisbon Valley area.

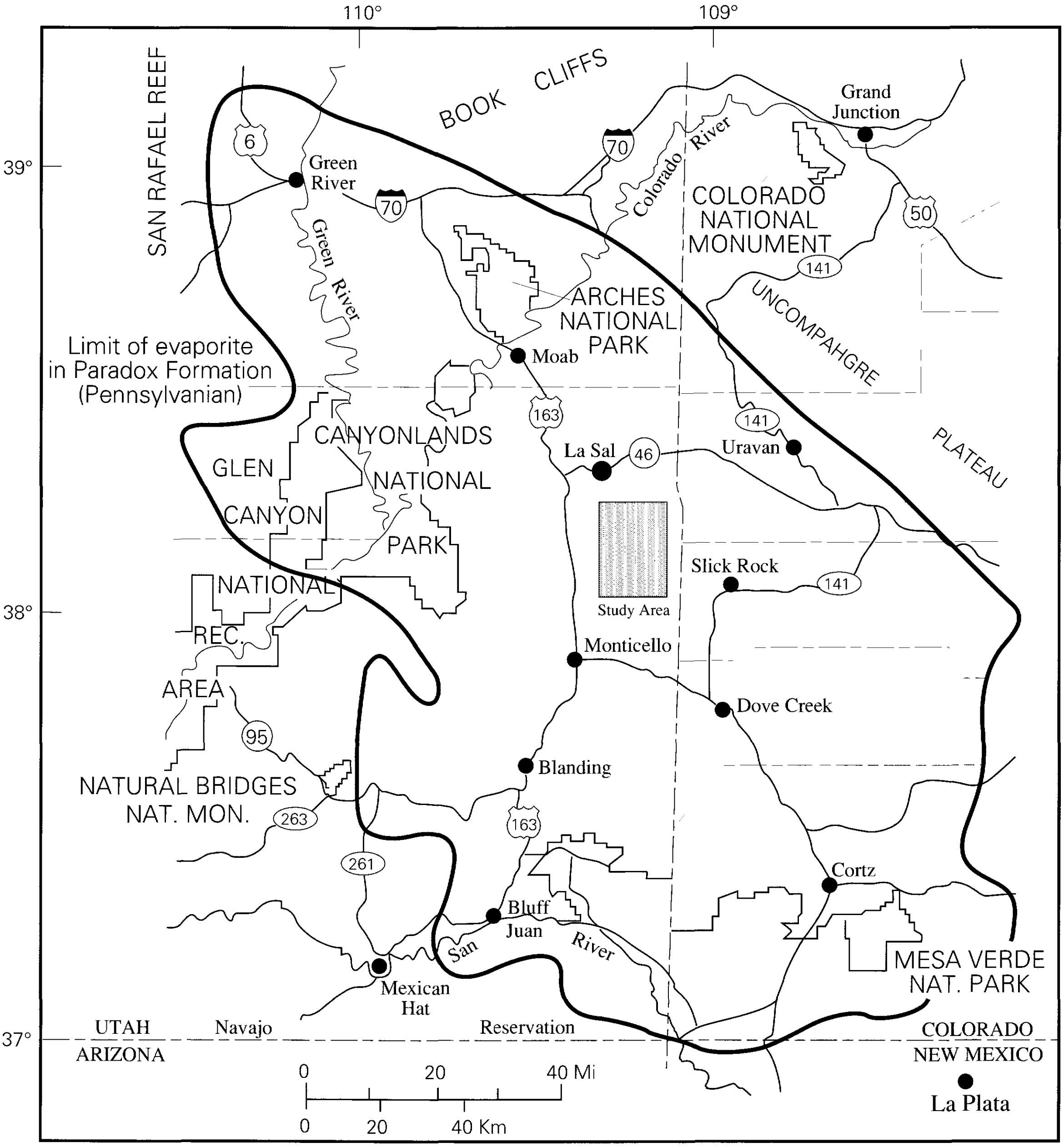
## INTRODUCTION

The Paradox Basin (fig. 1) is a tectonic depression of late Paleozoic age, the boundaries of which are generally defined by the geographic extent of halite deposited within the Paradox Formation during Middle Pennsylvanian time (Hite, 1968; Hite and others, 1972; Baars and Stevenson, 1981; Stevenson and Baars, 1987). The Paradox Basin was formed in Middle Pennsylvanian time and continued as a major site of deposition through and after Permian time. Prior to formation of the ancestral Rocky Mountains, the region was on the trailing edge of the North American craton and was the site of marine shelf deposition. During uplift of the ancestral Rockies, the basin subsided rapidly, accumulating as much as 9,000 ft of Middle and Upper Pennsylvanian evaporite, shale, and limestone, and about 6,000 ft of Permian marine and continental strata. Triassic and Jurassic deposition in the Paradox Basin was dominated by continental lacustrine, fluvial, and eolian systems.

Lisbon Valley is in the Paradox fold and fault belt, a tectonic region on the northeast side of the Paradox Basin dominated by northwest-trending folds and faults (fig. 1) (Kelley, 1955). Lisbon Valley encompasses the Lisbon Valley anticline and the Disappointment Valley syncline. The Lisbon Valley fault strikes northwest along the crest of the Lisbon Valley anticline and dips about 60° NE. The fault is a single plane in the central part of Lisbon Valley and is a fault zone near the northwest- and southeast-plunging noses of the anticline (Lekas and Dahl, 1956).

Continental deposits in the Paradox Basin are host to abundant energy and mineral resources. Uranium and vanadium, important energy and industrial resources abundant in sedimentary strata of the Paradox Basin, are present locally in Permian, Triassic, and Jurassic continental sandstones; major production is from the Lisbon Valley, Paradox Valley, and Sinbad Valley (fig. 1) structural area (the Uravan mineral belt) (Chenoweth, 1975, 1989). The Lisbon Valley uranium district is about 30 mi southeast of Moab, Utah.

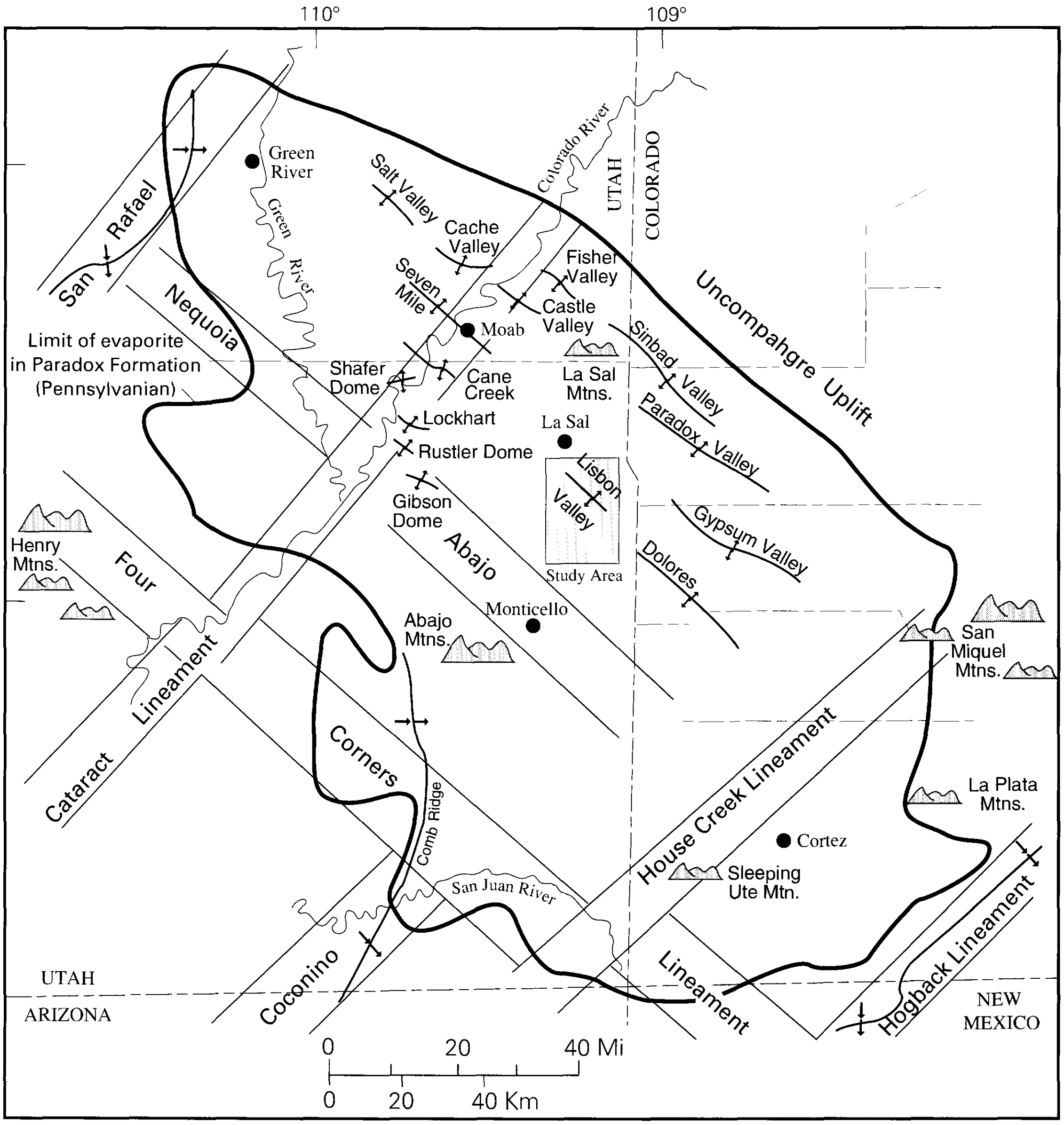
Many previous reports discuss the occurrence and origin of the uranium-vanadium deposits of the Paradox Basin, especially the large deposits in Lisbon Valley (for example, Gross, 1956; Lekas and Dahl, 1956; Williams, 1964; Wood, 1968; Chenoweth, 1975, 1989; Campbell and Steele, 1976; Campbell and Steele-Mallory, 1979a; Huber, 1979, 1980, 1981; Campbell, 1980; Weir and Puffett, 1981; Reynolds and others, 1985; and references therein).




*A*

**Figure 1 (above and facing page).** Major structural and cultural features of the Paradox Basin, Utah. *A*, Study area (shaded), cultural features, and maximum limit of evaporite (halite, anhydrite, and gypsum) deposition in the Middle Pennsylvanian Paradox Formation. Modified from Baars and Stevenson (1981); evaporite limit from Hite and others (1972). *B*, Major structural lineaments (thin parallel lines), salt anticlines (axis and anticline symbols), and maximum limit of evaporite (heavy line) in the Paradox Basin, Utah.

Several reports describe outcrop studies of the depositional setting of continental strata of both the ore-bearing Lower Permian Cutler Formation (Campbell, 1979, 1980; Campbell and Steele-Mallory, 1979a, b; Reynolds and others, 1985) and the Upper Triassic Chinle Formation (Huber, 1979, 1980, 1981), and a few reports describe studies of drill-hole geophysical logs, cutting samples, and petrography of core samples (Bohn, 1977; Huber, 1979, 1980) from Lisbon Valley. Other research has focused on petrography and diagenesis of uranium-vanadium ores in the Cutler and Chinle Formations from both surface and subsurface samples (Campbell, 1979; Campbell and Steele-Mallory, 1979a, b; Weir and Puffett, 1981); however, to our knowledge, few, if any, published reports describe continuous sequences of sedimentary structures and lithofacies in core from either the Cutler or the Chinle




***B***

within Lisbon Valley or from any other area of the Paradox Basin.

Uranium exploration drill cores from the Cutler and Chinle Formations in Lisbon Valley, which are reposited at the U.S. Geological Survey (USGS) Core Research Library at the Denver Federal Center in Lakewood, Colorado, were originally drilled as part of a uranium exploration program conducted by Kerr-McGee Corporation in the late 1960's and early 1970's (William L. Chenoweth, written commun., 1990). The cores contain excellent examples of sedimentary structures and lithofacies sequences formed in fluvial and eolian depositional environments. This report describes sedimentary structures and lithofacies sequences in five cores from Lisbon Valley (fig. 2) to provide the basis for recognizing small-scale features in core that are critical to interpreting depositional environments in the Cutler and Chinle Formations. These features and interpretations can be compared to outcrops and measured stratigraphic sections that, due to weathering, may not preserve details of fine-grained units. Measured stratigraphic sections of outcrops of the Chinle and Cutler Formations from Lisbon Valley are presented for comparison with the cores. These core descriptions provide initial interpretations of depositional environments in the subsurface and, combined with outcrop exposures, detail important lateral facies changes in both the Cutler and the Chinle Formation. This study is part of ongoing core and outcrop studies of the Cutler and Chinle Formations related to stratigraphy, sedimentology, uranium-ore geochemistry and paragenesis, basinal fluid movement, and salt anticline history. Each of these research

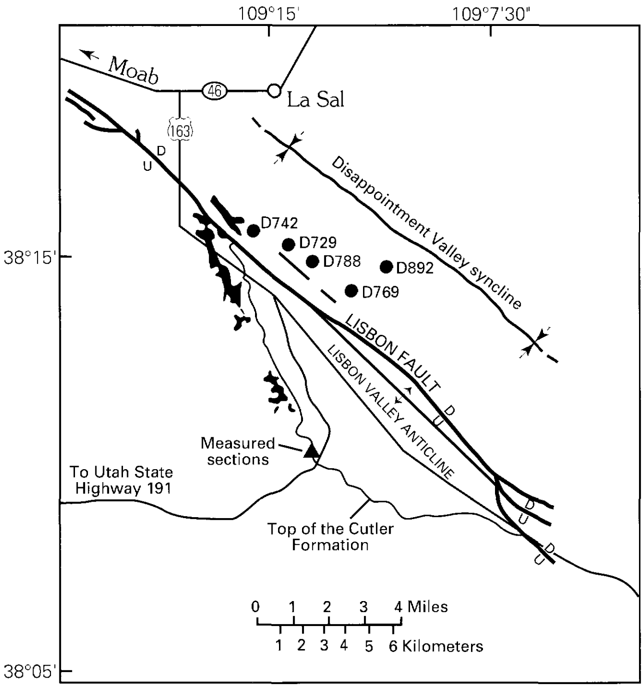



**Figure 2.** Locations of core holes (solid circles) and measured sections (triangles) of this report, general geologic structure, and approximate locations of uranium ore bodies (solid) in the Chinle Formation (U.S. Atomic Energy Commission, 1959, and Chenoweth, 1989), Lisbon Valley, Utah. Detailed location information for core holes is given in table 1.

efforts is part of a multidisciplinary project examining the Paradox Basin as part of the Evolution of Sedimentary Basins Program of the U.S. Geological Survey.

*Acknowledgments.*—We thank Steve Condon, Paula Hansley, and Jack Stanesco of the U.S. Geological Survey for reviews and comments to improve this manuscript. Tom Michalski and Gene Gay of the U.S. Geological Survey Core Research Library provided laboratory research space and superb core preparation that facilitated study of the cores described in this report. William L. Chenoweth, Grand Junction, Colorado, was instrumental in suggesting leads to information on and the localities of the cores from Lisbon Valley, and he provided additional comments and references for the manuscript. Lee Fairchild, Exxon Production Research, Houston, Texas, provided insightful lectures on extensional tectonics and salt anticline mechanics in the Paradox Basin.

## REGIONAL SETTING

The Paradox Basin is a major northwest-trending structural depression that formed during Middle Pennsylvanian time in association with uplift of the adjacent ancestral Uncompahgre highlands of the ancestral Rocky Mountains in southwestern Colorado (fig. 1). Vertical tectonism created a major structural and topographic high adjacent to a deep, asymmetrical subsiding basin. The major locus of subsidence and associated clastic deposition in the Pennsylvanian was on the northeast flank of the basin adjacent to the Uncompahgre uplift. Evaporite and limestone deposited in the central part of the basin interfinger with coarse clastic material shed from the highland source on the northeast. The clastic rocks are generally restricted to a narrow belt adjacent to the basin-bounding fault on the northeast edge of the basin, although turbidite beds may extend farther into the basin. In the Late Pennsylvanian, coarse clastic systems prograded into the basin and buried the evaporites under a wedge of interbedded carbonate and clastic strata that thickened toward the Uncompahgre uplift. As clastic sediments accumulated, a density inversion was established, and salt within the evaporite beds rose toward the surface as diapiric domes, anticlines, and walls (Lee Fairchild, oral commun., 1990). The location and orientation of many of the diapiric structures were controlled by preexisting basement faults, lineaments, and structural features (fig. 1) (Szabo and Wengerd, 1975; Campbell, 1979; Baars and Stevenson, 1981).

Clastic sedimentation to the southwest into the Paradox Basin from the ancestral Uncompahgre highlands continued into the Permian, maintaining growth of the salt anticlines. During the Triassic, marginal-marine to continental red beds of the Lower and Middle(?) Triassic Moenkopi Formation and variegated continental strata of the Upper Triassic Chinle Formation filled the basin. Angular unconformities within Permian and Triassic rocks attest to continued salt diapirism and movement on the salt anticlines through the Triassic and into the Jurassic (Weir and Puffett, 1981; Goydas, 1989).

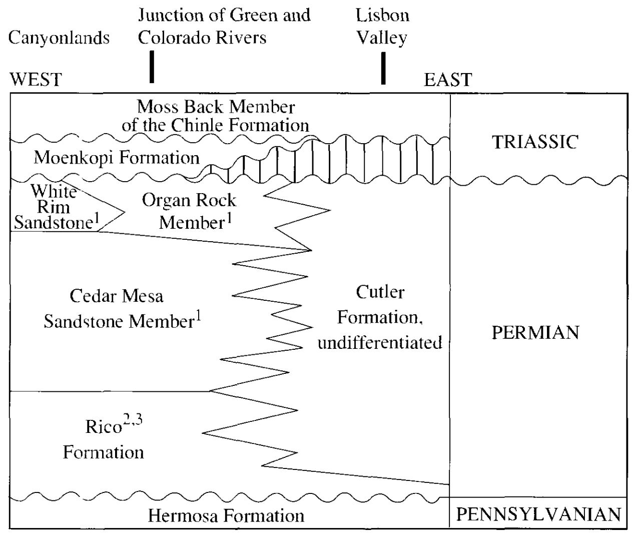
The Lisbon Valley anticline is one of the prominent salt anticlines of the Paradox Basin, and it differs from several others in that Pennsylvanian salts did not breach the surface (Cater, 1970). The northeast side of the anticline has been dropped along the Lisbon fault approximately 4,000 ft at the crest, juxtaposing Cretaceous rocks northeast of the fault against Pennsylvanian strata on the southwest. The absence of the upper part of the Cutler Formation and the Moenkopi Formation (Campbell and Steele-Mallory, 1979a; Weir and Puffett, 1981) from the central part of the Lisbon Valley anticline, the presence of the Moenkopi between Cutler and Chinle strata in adjacent synclinal areas (Budd, 1960; Wood, 1968), and the slight disparity in structural strike and dip between the Chinle and the Cutler (Campbell and Steele-Mallory, 1979a; Weir and Puffett, 1981) all suggest that salt diapirism within the Lisbon Valley structure was active during the Triassic. Sedimentologic studies of fluvial systems in the Chinle Formation in and around Canyonlands National Park near Moab, Utah (Blakey and Gubitosa, 1983), and in Lisbon Valley (Huber, 1979) propose that Late Triassic fluvial systems were affected by active movement on salt anticlines.

## STRATIGRAPHY

Limestone and sandstone of the Middle and Upper Pennsylvanian Hermosa Formation are the oldest rocks exposed in Lisbon Valley, cropping out along the axis of the Lisbon Valley anticline (Weir and Puffett, 1981). The Hermosa is overlain by Permian rocks that in the Lisbon Valley area have previously been mapped as Cutler Formation undifferentiated (Williams, 1964). The beds at the base of the Cutler section in Lisbon Valley and other areas contain marine sandstone and limestone and have been referred to by various authors as Elephant Canyon Formation (Baars, 1962, 1987), part of the marine facies of the Cedar Mesa Sandstone (Campbell, 1979; Campbell and Steele-Mallory, 1979b), Rico Formation (Stanesco and Campbell, 1989), and lower Cutler beds (Loope and others, 1990). Other subdivisions of Permian rocks recognized elsewhere in southeastern Utah have not been used or mapped in Lisbon Valley, although rock types representing facies of those units are present in southeastern Utah (Campbell and Steele-Mallory, 1979a; Stanesco and Campbell, 1989). The present report follows the Permian terminology proposed by Baars (1962) and subsequently adopted in Campbell and Steele-Mallory (1979a), Weir and Puffett (1981), and Stanesco and Campbell (1989) (fig. 3). The age of the Cutler Formation in Lisbon Valley is thought to be Wolfcampian (Campbell and Steele-Mallory, 1979a, b), although Baars (1962) and McKee and others (1967) suggested that the upper part of the Cutler, which may not be preserved in Lisbon Valley, may be Leonardian.

The upper contact of the Cutler Formation in Lisbon Valley is a regional unconformity. Above the unconformity, the Lower and Middle(?) Triassic Moenkopi Formation, present in adjacent areas of southeastern Utah, is missing in Lisbon Valley due to either nondeposition or erosion, and the Upper Triassic Chinle Formation rests directly on the Cutler Formation (Campbell and Steel-Mallory, 1979a, b; Weir and Puffett, 1981). Weir and Puffett (1981) and Huber (1980) reported less than a 5° angularity between the Cutler and the Chinle, and Campbell and Steele-Mallory (1979a, b) described the Cutler as having a steeper and more southerly dip than the overlying Chinle; both observations suggest that at least slight tectonic movement occurred on the Lisbon Valley anticline prior to Chinle deposition.

The Upper Triassic Chinle Formation is present throughout southeastern Utah, where seven formal members and several stratigraphically equivalent informal members are recognized. The seven formal members, in ascending order, are the Temple Mountain, Shinarump, Monitor Butte, Moss Back, Petrified Forest, Owl Rock, and Church Rock (Stewart and others, 1972). The Shinarump and Monitor




[1] of the Cutler Formation
[2] Also called Elephant Canyon Formation (Baars, 1962)
[3] Also called "lower Cutler Beds" (Loope and others, 1990)

**Figure 3.** Schematic stratigraphic section showing Permian and Triassic nomenclature in southeastern Utah between Canyonlands and Lisbon Valley. Modified from Stanesco and Campbell (1989).

Butte Members are thought to be absent in Lisbon Valley (Stewart, 1969). The basal sandstone of the Chinle in Lisbon Valley is generally assigned to the Moss Back Member, and the remaining part of the Chinle is referred to the Church Rock Member (Stewart and others, 1972; Weir and Puffett, 1981); however, Stewart and others (1972) also suggested that the lower sandstone unit of the Chinle in Lisbon Valley may be younger than the type Moss Back. In addition, O'Sullivan (1970) contended that the Church Rock in southeastern Utah, as used by Stewart (1957), Stewart and others (1959), and subsequently by both Stewart and others (1972) and Weir and Puffett (1981), is older than the type Church Rock Member farther south along Comb Ridge in Arizona. O'Sullivan and MacLachlan (1975) did not use formal nomenclature for the Chinle because of the marked facies changes recognized in southeastern Utah. They used instead an informal lithologic terminology that included the claystone, limy, and siltstone members, in ascending order. Huber (1979, 1980) referred to the sandstone at the base of the Chinle in Lisbon Valley as the Moss Back Member and termed the remaining overlying units the upper part of the Chinle. The sandstone units of the Chinle in the cores described in the present study are all at the base of the formation and are considered to be part of the Moss Back Member as used by Stewart and others (1972), Weir and Puffett (1981), and Huber (1979, 1980). Reconciliation of the nomenclature of the upper part of the Chinle, not present in these cores, is deferred pending further field investigations.

## SEDIMENTOLOGY

Five cores from the northwest end of Lisbon Valley (fig. 2, table 1) were slabbed at the USGS Core Research Library prior to examination. The cores provide superb examples of sedimentary structures and lithofacies sequences that support interpretations of continental depositional systems in the Cutler and Chinle Formations. The depositional environments in these cores include fluvial and eolian facies. Both of these general facies have been recognized from outcrop studies in Lisbon Valley (Campbell, 1979, 1980, 1981; Campbell and Steele-Mallory, 1979a, b; Huber, 1979, 1980), and additional lithofacies not recognized from outcrop studies are well preserved in the cores. Depositional environments in the cores were interpreted on the basis of lithology, sedimentary structures, lithofacies sequences, and comparison with previously published descriptions of fluvial and eolian facies in the Cutler and in other units, from both outcrop, laboratory, and core examples (for example, Campbell and Steele-Mallory, 1979a, b; Fryberger and others, 1979; Fryberger and Schenk, 1981, 1988; Ahlbrandt and Fryberger, 1982; Cant, 1982; Kocurek and Nielson, 1986; Fryberger and others, 1990; Schenk, 1990; Fryberger, 1991).

The five cores are referred to herein by their USGS Core Research Library number: D729, D742, D892, D769, and D788 (table 1). Each of the cored intervals begins in the lower part of the Chinle Formation and extends down through the Cutler-Chinle contact into the uppermost part of the Cutler Formation. The cores are archived at the U.S. Geological Survey Core Research Library, Building 810, Denver Federal Center, Denver, Colorado, 80225. A sequence of photographs (fig. 4) depicts the entire cored interval in D729, which contains examples of each of the depositional environments in the other cores. A stratigraphic section of both the Chinle and Cutler Formations was measured just north of Big Indian Rock on the west side of Lisbon Valley and about 5 mi south of the core locations (fig. 2). The outcrop section and the sedimentary structures and lithofacies within it provide an insightful comparison with features preserved in the cores.

The following sections describe sedimentary features and lithofacies in the cores and in the measured stratigraphic sections. The complete core descriptions (appendix), measured sections (figs. 5, 6), and associated data were recorded on standardized forms and as field notes that include descriptions of lithology, grain size, color (Goddard and others, 1948), sedimentary structures, and other parameters.

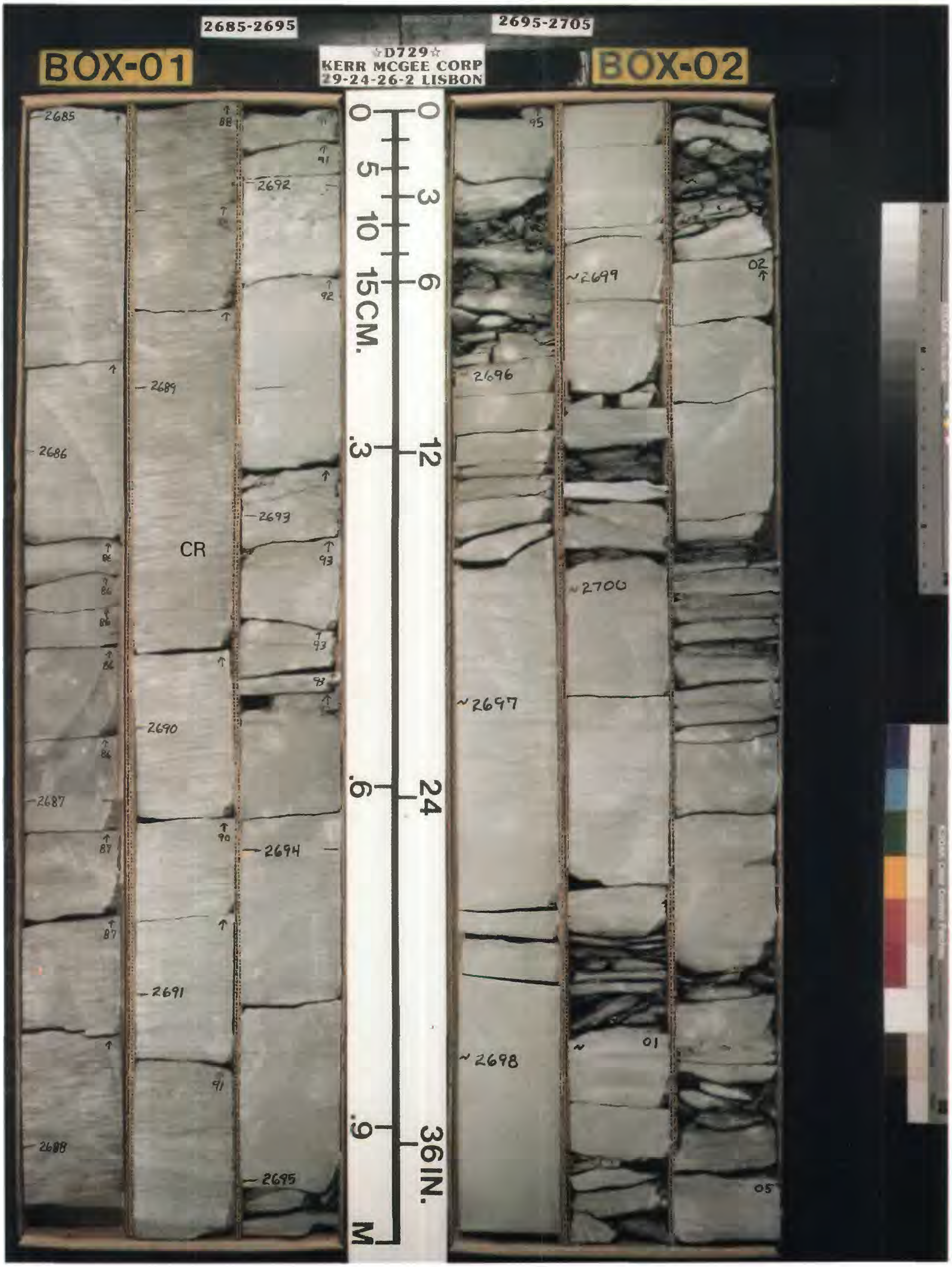
**Table 1.** Location and length of cores used in study, Lisbon Valley, San Juan County, Utah.

| USGS core number | Location | Length of core (in feet) |
|---|---|---|
| D729 | Sec. 26, T. 29 S., R. 24 E. | 174 |
| D742 | Sec. 16, T. 29 S., R. 24 E. | 130 |
| D769 | Sec. 6, T. 30 S., R. 25 E. | 61 |
| D788 | Sec. 25, T. 29 S., R. 24 E. | 231 |
| D892 | Sec. 32, T. 29 S., R. 25 E. | 290 |

## LITHOFACIES AND DEPOSITIONAL ENVIRONMENTS OF THE CUTLER FORMATION

In the cores, the Cutler Formation consists primarily of arkosic sandstone, siltstone, shale, and mudstone; the cores apparently did not extend deep enough to intercept marine limestone and sandstone identified in the lower part of the Cutler from outcrop studies in Lisbon Valley (Campbell and Steele-Mallory, 1979a, b; Campbell, 1980). Similar to the measured outcrop sections in this and previous studies, the reddish-brown and orange strata of the Cutler in the cores distinguish it from the overlying greenish-gray rocks of the Moss Back Member at the base of the Chinle Formation (fig. 4). The upper part of the Chinle Formation, as seen at the measured section in outcrop in Lisbon Valley, is

**Figure 4 (following page).** Whole-core photographs of sequence of slabbed core D729 (Kerr-McGee Corp., sec. 26, T. 29 S., R. 24 E., Lisbon Valley, Utah) of the Lower Permian Cutler Formation (undifferentiated) and the overlying Moss Back Member of the Upper Triassic Chinle Formation. The photographs start at the top of the cored sequence within the Moss Back Member in box 01 and proceed to the base of the cored interval in the Cutler Formation in box 18. Scale and color bars are on each photograph. Specific features and explanations for the entire cored interval can be compared to the core descriptions in subsequent figures. Selected examples of sedimentary structures and environments labeled on the photographs. AV, avalanche grainfall laminae in dune facies; BB, animal burrow or zone of intense bioturbation; CC, clay-chip conglomerate; CL, coarse fluvial-channel lag deposit; CR, climbing ripple lamination; M, massive or structureless sediment; OS, oversteepened cross sets; PF, plant fragments and organic material; PL, planar cross laminae; RN, rooted zone with or without secondary carbonate nodules; SF, slump feature from failure and nontransport of sediment blocks; SK, sabkha deposit; SS, soft-sediment deformation; SY, stylolite. The interval from 2,685 to 2,767.5 ft shows drab-colored fluvial channel fill and crevasse splay deposits of the Chinle Formation. The Cutler Formation below 2,767.5 ft is characterized by its red to orange color and represents interbedded fluvial and eolian deposition. See appendix for details of depositional environments. Location of core hole is shown in figure 2. Color negatives of the photographs are available for use at the USGS Core Research Library, Denver Federal Center, Lakewood, Colorado 80225.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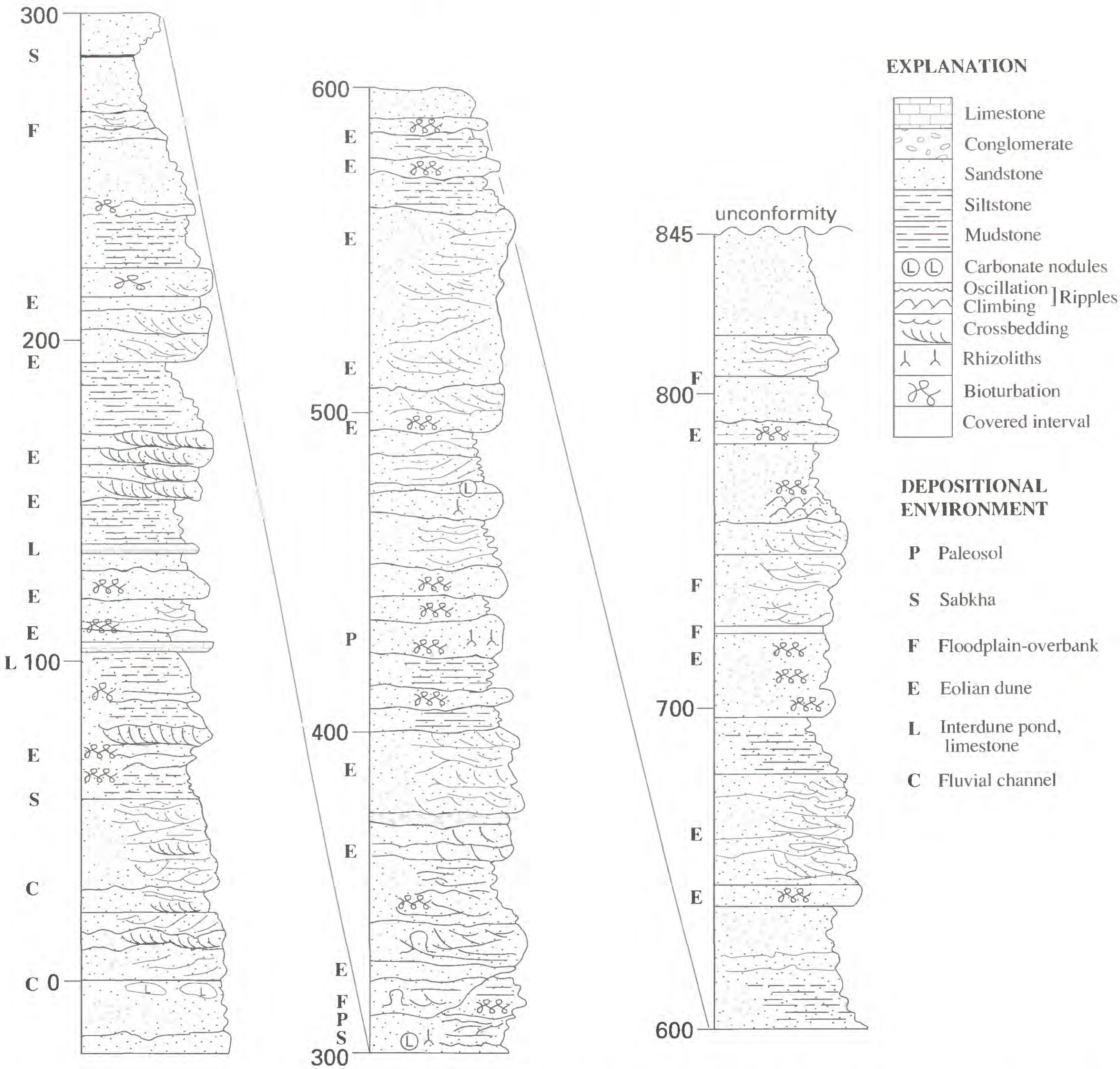


**Figure 5.** Measured stratigraphic section of the upper part of the Cutler Formation in Lisbon Valley, Utah, just north of Big Indian Rock. About 20 ft of outcrop is depicted below the base of the measured section, which begins at 0 ft. An undetermined thickness of the Cutler exists below that shown on the section. The Chinle Formation unconformably overlies the Cutler at 845 ft. Location of measured section is shown in figure 2.

composed of the Church Rock Member, which is dark reddish brown, but everywhere in Lisbon Valley the greenish Moss Back Member separates the Church Rock Member from the underlying Cutler Formation. None of the cores described to date extends high enough in the Chinle to include the reddish-brown Church Rock Member.

Within the Cutler Formation, five distinct lithofacies can be recognized on the basis of color, grain size, bedding, and sedimentary structures. The suite of sedimentary structures within each lithofacies compares well with those described in the literature for distinct fluvial and eolian depositional environments, including fluvial channel, eolian dune, interdune, eolian sand sheet, and sabkha.

## FLUVIAL

Fluvial channel deposits in the Cutler consist of coarse- to fine-grained, poorly sorted, subrounded to subangular,

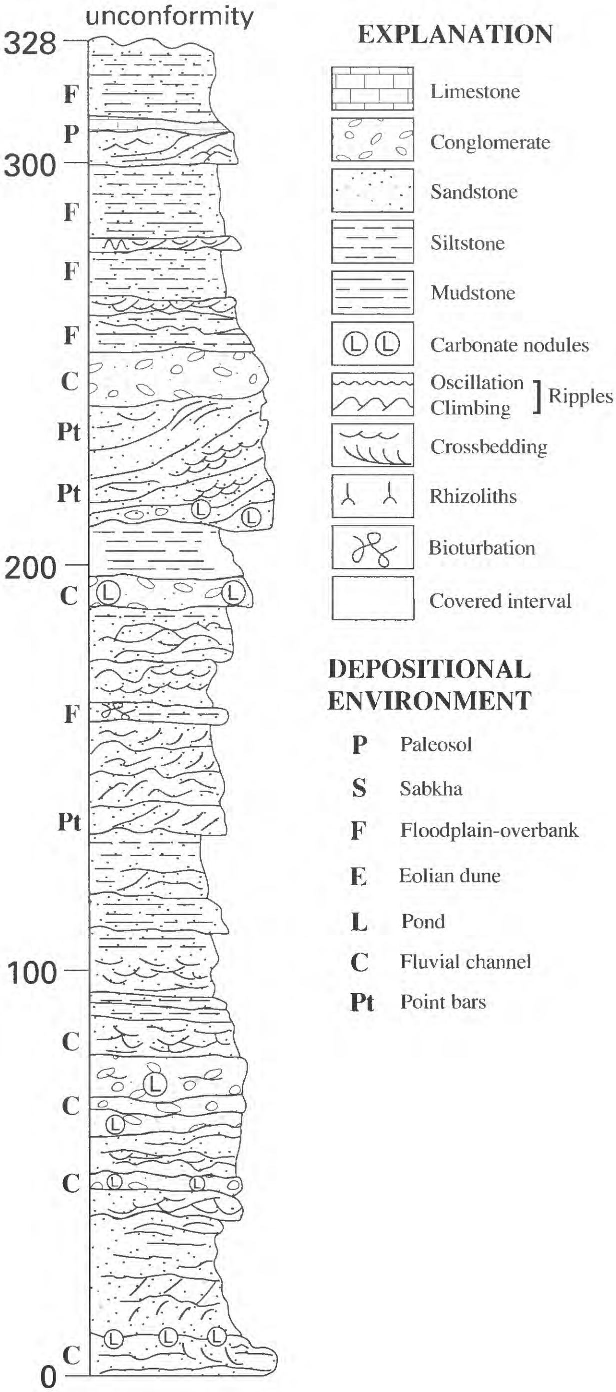


**Figure 6.** Measured stratigraphic section of the entire Chinle Formation in Lisbon Valley, Utah, just north of Big Indian Rock. The Chinle Formation unconformably overlies the Cutler Formation at 0 ft at the base of the measured section. The Wingate Sandstone unconformably overlies the Chinle at 328 ft. The Moss Back (0–80 ft) and Church Rock Members (80–328 ft) of the Chinle Formation are present at this outcrop locality. Location of measured section is shown in figure 2.

thin- to thick-bedded, arkosic sandstone that generally fines upward within individual beds. High-angle crossbeds indicative of either large-scale trough or planar cross stratification are abundant, and low-angle to horizontal laminations are common. The scale of the sedimentary structures decreases in size upward in the beds. The fluvial channel deposits have sharp scoured bases and transitional upper contacts. Clay and mudstone clasts are common at the bases of the units. The grain size of the sandstone, the scoured lower contacts, and the trough and planar crossbeds indicate high-energy, channelized flow, probably in fluvial environments. Campbell and Steele-Mallory (1979a, b) attributed this lithofacies on outcrops in Lisbon Valley to meandering stream deposits on the basis of lenticular outcrop pattern and lateral facies relations.

Associated with the fluvial sandstone are dark- to moderate-reddish-brown siltstone, shale, and mudstone. Sedimentary structures within these units are well preserved in the cores, in contrast to the measured section of this report and other outcrop studies. Campbell and Steele-Mallory (1979a) noted poor preservation and limited interpretation of fine-grained red-bed units in the Cutler of Lisbon Valley due to outcrop weathering. The fine-grained strata are structureless to laminated, and they commonly are contorted or bioturbated. Siltstone and mudstone locally exhibit carbonate nodules arranged in downward-bifurcating patterns, and they locally contain climbing ripples. Interpretation of these units as fluvial floodplain deposits is based on their sedimentary structures and comparison with similar features described in other studies (for example, Cant, 1982) and on the vertical association in the cores with adjacent lithofacies. However, the units contain some characteristics also common to the eolian sand-sheet and sabkha deposits, and an unequivocal interpretation based on core exposures is commonly impossible.

The downward-bifurcating nodules are interpreted as rhizocretions formed around the traces of former plant roots (Klappa, 1980) The gradation upward from coarse fluvial channel deposits into these fine-grained strata suggests a spatial association of fluvial and floodplain settings; however, unlike outcrop exposures in which lateral facies associations can commonly be observed, only the vertical lithofacies associations of these units can be seen in the cores.

Color is the most notable feature distinguishing fluvial strata from eolian rocks within the Cutler Formation in the cores. Fluvial deposits are grayish red and purple to dark reddish brown, whereas eolian strata are generally moderate reddish orange to light brown or white. This distinction between reddish-brown and purple fluvial rocks and orange eolian strata has also been recognized in the Chinle Formation (Dubiel and Skipp, 1989; Dubiel, 1992) and in the Middle Pennsylvania to Lower Permian Maroon Formation of the Eagle Basin in western Colorado (Johnson and

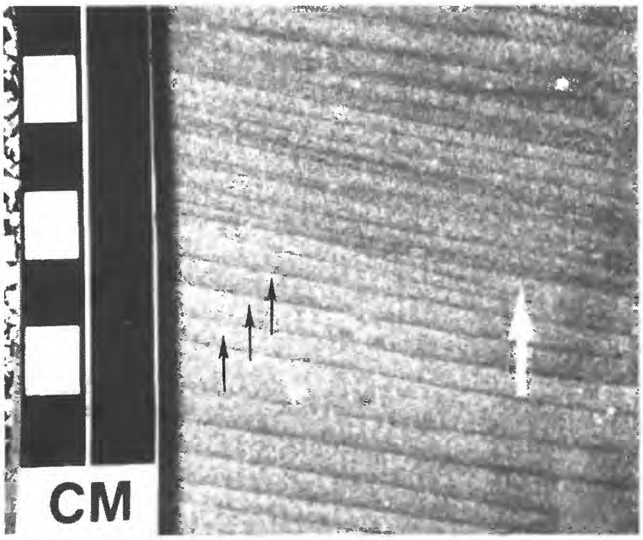


**Figure 7.** Photograph of core D729 at 2,833 ft showing oblique cut through high-angle cross stratification in an eolian dune in the Cutler Formation. The coarsening upward in each lamination (small arrows) and the preserved ripple cross lamination (large arrow) in the center of the photograph are characteristic of eolian wind-ripple deposition. Location of core hole is shown in figure 2.

others, 1988; Dubiel, unpublished data). On the outcrop, the color distinction between orange eolian and reddish-purple fluvial deposits is striking.

Both the measured section for this report and previous studies (Campbell and Steele-Mallory, 1979a, b) show that the fluvial channel beds on the outcrop are lenticular in cross section and grade laterally into finer grained clastic deposits.

## EOLIAN DUNE

Eolian deposits of the Cutler Formation in the cores are distinguished in part by their light-reddish-orange color (fig. 4, 2,825–2,835 ft). The eolian strata are well sorted, fine to very fine grained sandstone interbedded with minor siltstone. Abundant sedimentary structures indicate eolian deposition, and the included sedimentary structures distinguish eolian dune from eolian sand-sheet deposits. Steeply to moderately dipping, concave-upward crossbeds in medium to thick beds are indicative of large bedforms and are interpreted as deposits from migrating eolian dunes. Internally, the crossbedded units contain upward-coarsening laminations, pinstripe laminations, and high-index ripple laminations that are diagnostic of migrating wind ripple origin (fig. 7) (Hunter, 1970; Fryberger and others, 1979; Fryberger and Schenk, 1981, 1988; Ahbrandt and Fryberger, 1982). Locally, the ripple form can be recognized within the pinstripe lamination (fig. 7). Oversteepened and slumped crossbeds are rare features (fig. 4, 2,830 ft). The eolian dune deposits are commonly several feet thick (fig. 4, 2,825–2,835 ft) and are bounded above and below by eolian sand-sheet and interdune deposits. The thicker crossbed sets distinguish the eolian dune environments from smaller eolian bedforms on the eolian sand sheets.

On the outcrop this facies is well exposed as thick, light-orange to white beds that show large-scale cross-bedding. Units can be traced for long distances on the west rim of Lisbon Valley. Eolian dune deposits are one of the most distinctive (Campbell and Steel-Mallory, 1979a, b) and easily recognizable facies within outcrops of the Cutler Formation in Lisbon Valley (fig. 5).

## INTERDUNE

Interbedded with eolian dune strata are rare, dark-reddish-brown, very thin bedded to thin-bedded, finely laminated to locally bioturbated mudstone, claystone, and siltstone (fig. 8). These units have sharp upper and lower contacts with bounding eolian deposits. Internally they contain small-scale soft-sediment deformation, probably due to water saturation and loading by overlying sediment. These fine-grained laminated mudstone and claystone units are interpreted as wet interdune deposits and probably represent small interdune ponds (Ahlbrandt and Fryberger, 1982).

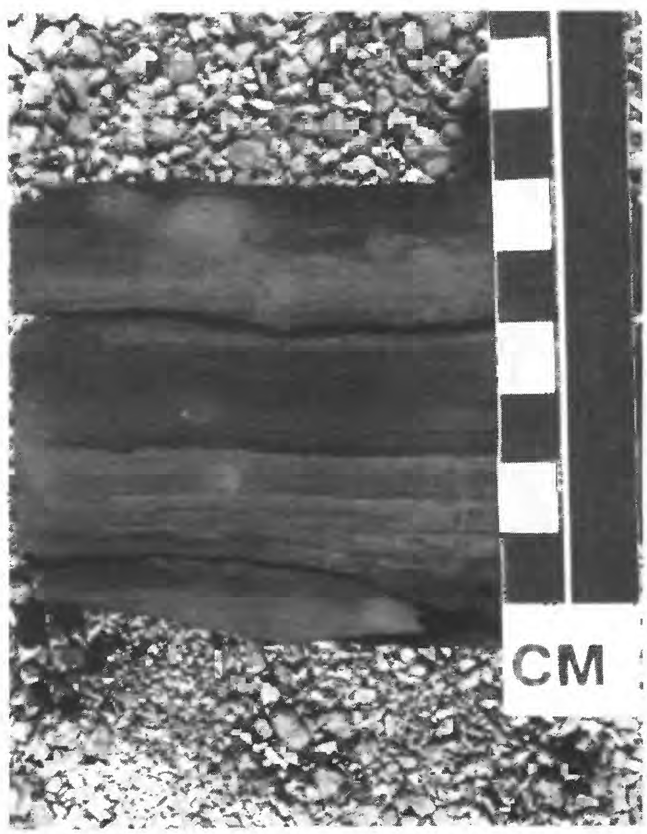


**Figure 8.** Photograph of core D788 at 2,806 ft showing finely laminated claystone and mudstone of an interdune pond deposit in the Cutler Formation. Location of core hole is shown in figure 2.

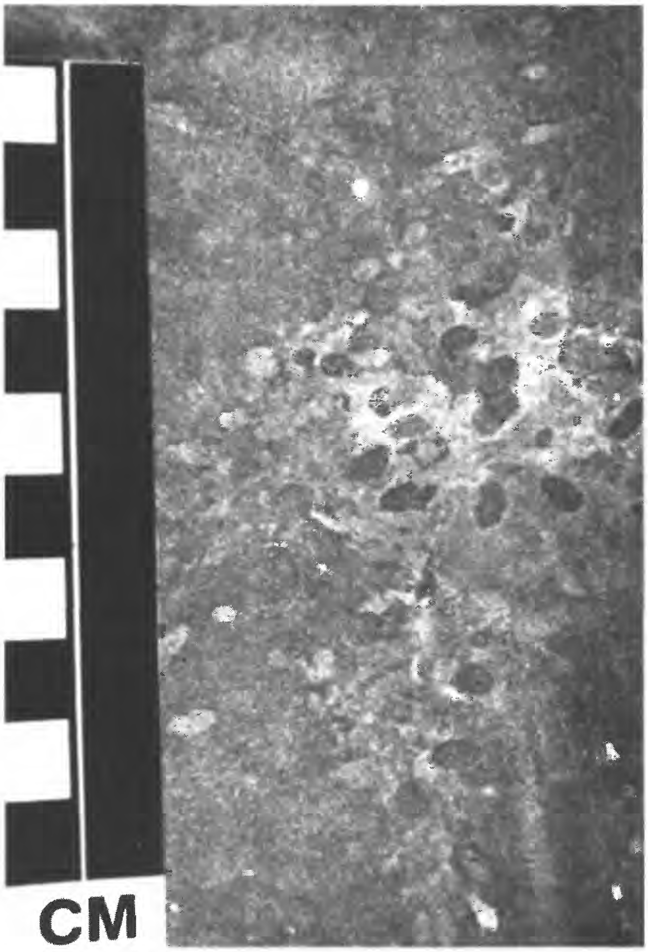


**Figure 9.** Photograph of core D729 at 2,820 ft showing extensively bioturbated sandstone of an eolian sand-sheet deposit in the Cutler Formation. Location of core hole is shown in figure 2.

Eolian dune, sand-sheet, and sabkha strata are commonly bleached white for several inches adjacent to interdune pond mudstone, possibly as a result of the removal of iron due to reducing fluids generated in the pond mudstone.

On the outcrop, interdune pond deposits are rare. Because of destruction by weathering of primary depositional structures and fabrics inherent to specific facies, fine-grained deposits of interdune ponds are difficult to distinguish from floodplain, sand-sheet, and sabkha deposits. Previous studies grouped by association on the outcrop these fine-grained shale, siltstone, and mudstone deposits into overbank, floodplain, levee, and lacustrine environments. The preservation of depositional structures and fabrics in the cores affords the potential to define and recognize several specific facies within these fine-grained rocks.

## EOLIAN SAND SHEET

The eolian sand-sheet deposits in the Cutler Formation contain sedimentary features similar to those of the eolian dune deposits, but the sand-sheet deposits commonly contain thinner bedding and smaller scale crossbedding. The eolian sand sheets comprise moderate-reddish-orange to pale-reddish-brown, very fine grained to fine-grained sandstone. Locally abundant, small lenses of coarse-grained sand are probably deflationary lag grains. Sand sheets contain small-scale, low-angle crossbeds and wavy parallel laminations that are commonly disrupted by extensive bioturbation (fig. 9). Meniscate backfilled burrows are common and were formed by arthropods. Two kinds of bioturbation were formed by roots. Vertically stacked carbonate nodules probably formed as carbonate precipitates along root traces (fig. 10), and downward-bifurcating, purple and white mottled root alteration haloes formed from alteration of iron oxide minerals along decomposing roots. The low-angle crossbedding, the abundant root traces, and extensive bioturbation indicate deposition on low-relief eolian sand sheets (Fryberger and others, 1979; Ahlbrandt and Fryberger, 1982; Kocurek and Nielson, 1986).

On the outcrop, sand-sheet deposits are recognized by their light-orange to white color, small-scale crossbedding, and laterally persistent thin to medium beds. The beds pinch out laterally over large distances. Sand-sheet deposits

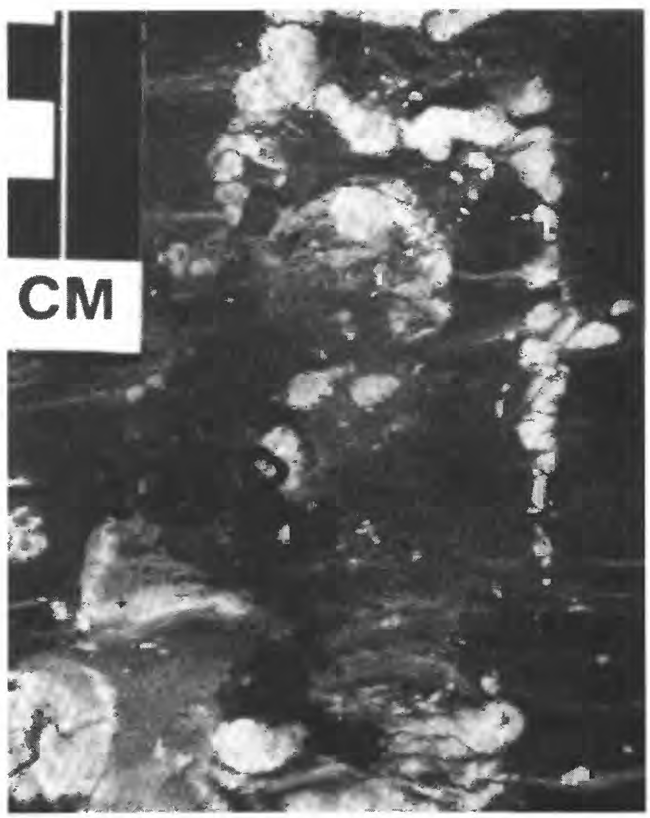


**Figure 10.** Photograph of core D742 at 2,757 ft showing vertically stacked and downward-bifurcating carbonate nodules. This is a rhizocretion, a carbonate precipitation around a former plant root in a floodplain deposit of the Cutler Formation. Location of core hole is shown in figure 2.

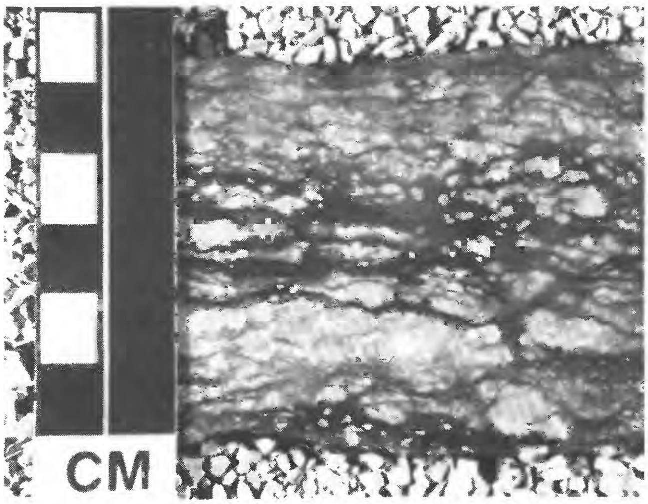


**Figure 11.** Photograph of core D729 at 2,856 ft showing sabkha deposit in the Cutler Formation. Note the evaporite nodules and the deformed bedding. Location of core hole is shown in figure 2.

contain more sandstone and less mudstone than floodplain deposits, and they are distinguished from sabkha deposits by the lack of evaporite minerals, by their fabric, and by disrupted bedding caused by mineral growth and dissolution.

### SABKHA

Sabkha deposits within the Cutler Formation are composed of deformed and planar laminated, light- to dark-reddish-brown, very fine grained to fine-grained sandstone and siltstone beds that have silty mudstone drapes (fig. 11). The beds are thin to thick. Internally the deposits are commonly disturbed by wavy parallel, wavy nonparallel, and wavy discontinuous laminations. The disruption of laminae is commonly centered about small nodules and coalesced mosaics of small nodules. Some beds are only moderately disturbed and others are extremely disrupted, exhibiting small, randomly oriented remnants of the original laminae. Nodules in the beds are present as small displacive growths and as nodular-mosaic thin beds and wavy mudstone interbeds. The nodules are primarily gypsum and minor anhydrite. In addition, the cores contain deformed beds that do not contain visible nodules.

These beds were deposited on siliciclastic-dominated sabkhas and are distinguished by extensive haloturbation that has destroyed primary depositional fabric (fig. 11) (Ahlbrandt and Fryberger, 1982; Schreiber and others, 1982; Mazzullo and others, 1991). Both the displacive growth of evaporites in the original depositional environment and the subsequent replacement or removal of evaporites by dissolution probably account for the varying degrees of disruption in the beds. The displaced laminae that contain no apparent nodules argue for the complete dissolution of some former evaporite mineral.

The sabkha beds, although well represented in the cores, were difficult to discern on the outcrop measured section. The fine-grained lithology of the beds probably results in extensive destruction of small-scale features and nodules on the outcrop, making identification of this facies very difficult in the field. Reports of previous outcrop studies in Lisbon Valley (Campbell and Steele-Mallory, 1979a, b) do not mention sabkha deposits. Weathering may have made the sabkha deposits indistinguishable from other fine-grained units, or the sabkha facies may only be present north of Lisbon Valley, where it is present in the subsurface.

## LITHOFACIES AND DEPOSITIONAL ENVIRONMENTS OF THE CHINLE FORMATION

The Moss Back Member is the only unit of the Chinle Formation present in the cores. The Moss Back Member unconformably overlies the reddish-brown and orange Cutler Formation. The contact is easily distinguished because the Moss Back in the cores, as on the outcrop, is generally very pale green to light greenish gray and light bluish gray, in contrast to the reddish-brown upper part of the Chinle and the underlying Cutler. The drab colors of the Moss Back Member are thought to reflect a lack of oxidation of iron to red-colored hematite due to the abundant detrital organic matter preserved in the unit. The preservation of this abundant organic matter is thought to reflect rapid sedimentation, high water tables, and subaqueous deposition within the Chinle (Dubiel, 1989). In several of the cores, the base of the Moss Back Member contains abundant sulfide mineralization and minor uranium mineralization.

The Moss Back Member in the cores and on the outcrop can be divided into three dominant facies: fluvial channel, crevasse splay, and overbank floodplain.

### FLUVIAL CHANNEL

Fluvial deposits in the cores of the Moss Back Member are dominated by conglomerate and sandstone and minor siltstone and mudstone. The conglomerate in the cores contains few siliciclastic pebbles and is composed primarily of rounded to subrounded clay intraclasts and intrabasinal carbonate nodules (fig. 12), presumably reworked from older Chinle and possibly Moenkopi and Cutler deposits. Clay clasts were reworked from mudstone within the Moss Back Member. Carbonate clasts are common in nodule-bearing paleosols in Chinle floodplain deposits that were proximal to the ancestral Uncompahgre highland source area (Dubiel and Skipp, 1989; Dubiel, 1992) and in the underlying Cutler Formation floodplain and sabkha deposits. Huber (1980) reported quartz-pebble conglomerate at several outcrop localities in Lisbon Valley, however, only carbonate conglomerate and claystone conglomerate are present in the cores.

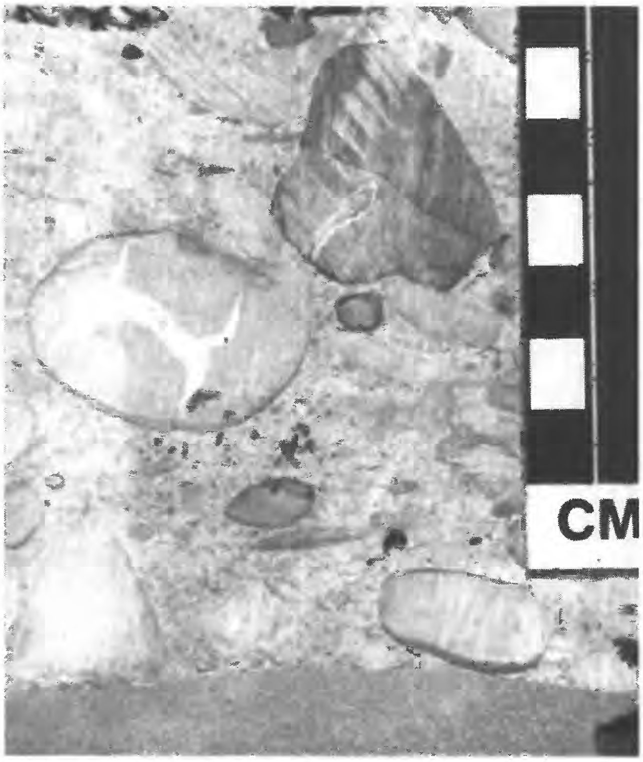


**Figure 12.** Photograph of core D742 at 2,726 ft showing carbonate-clast conglomerate in the fluvial channel facies of the Chinle Formation. Location of core hole is shown in figure 2.


**Figure 13.** Photograph of core D729 at 2,716 ft showing ripple cross-laminated sandstone in the Chinle Formation. Location of core hole is shown in figure 2.

The carbonate-clast conglomerate is medium to thick beddded and exhibits crude crossbedding and fining-upward sequences into low- to moderate-angle, crossbedded siliciclastic and carbonate sandstone. The sandstone fines upward into thick beds of ripple-cross-laminated sandstone and siltstone (fig. 13). The abundant ripple sets, some having high angles of climb, indicate rapid sedimentation. Many of the ripple laminae are defined by finely comminuted organic matter and plant fragments.

The scoured, lag-filled channel bases, thick beds, and upward succession of grain size and structures indicate deposition in fluvial channels. Despite the abundance of conglomerate in the basal channel fills, these strata are interpreted as high-sinuosity fluvial deposits on the basis of sedimentary structures such as climbing ripples, high suspended-sediment load, and comparison with modern and ancient meandering stream systems (Cant, 1982). In addition, the outcrop pattern of sandstone in the measured section and interpretations in previous studies (Huber, 1979, 1980) suggest a meandering fluvial system.

## CREVASSE SPLAY

Crevasse splay deposits in the Moss Back are similar in lithology and sedimentary structures to the fluvial channel deposits but are distinguished by their thinner bedding, smaller grain size in conglomeratic beds, and rapid transition between beds of differing sedimentary structures. The similarity in color, lithology, and grain size of crevasse splay deposits to the fluvial channel deposits and the lack of observation of lateral persistence of beds and facies changes in the cores makes some interpretations of crevasse splay environments equivocal. In general, the crevasse splay deposits have a greater percentage of fine-grained, suspended load deposition and commonly contain abundant plant fragments (fig. 14), whole plant fossils, and laminations of finely comminuted plant material on bedding planes in the siltstone and mudstone. Rare carbonized logs and sticks are present in the fine-grained units (fig. 4, 2,740.5 ft), as are rare reworked unionid bivalves.

The crevasse splays represent overbank or through-the-bank deposition during flood events. Both coarse-grained bed-load and fine-grained suspended-load sediment were deposited from the main channel or channel system. Crevasse splay deposits may coarsen or fine upward depending on whether the splay was prograding or was being abandoned. Crevasse splay deposits are complexly interbedded with the fluvial channel deposits.

On the outcrop, crevasse splay deposits are more easily distinguished from the channel deposits than in the cores.


**Figure 14.** Photograph of core D729 at 2,724 ft showing terrestrial plant fragments in the Chinle Formation. Location of core hole is shown in figure 2.

Lateral relations commonly reveal the thin but persistent splay units and their association with a channel deposit. Small-scale features such as ripple lamination are more visible in core than in outcrop, presumably due to weathering of fine-grained clastic material and clay on outcrop. In addition, oxidation due to weathering may have removed sulfide mineralization and uranium mineralization, which were not noted in the measured section of the Chinle.

### FLOODPLAIN

Fine-grained sandstone, siltstone, and mudstone are a minor part of the Moss Back Member in the cores. The units are drab colored and thin bedded to very thin bedded. Planar horizontal laminations and rare ripple cross lamination are marked by clay drapes and organic-matter fragments. Locally these units contain root traces and small, isolated carbonate nodules.

These units are interpreted as floodplain deposits formed from suspended-load deposition out of the main channels and crevasse splays during flood events. The carbonate nodules represent incipient paleosol development during periods of nondeposition.

On the outcrop, fine-grained units in the Moss Back generally weather to a debris-covered slope, and details of the facies are not generally visible. In the upper part of the Chinle, fine-grained mudstone lateral to lenticular channel sandstone is thicker and more common than in the Moss Back, and paleosol development is more pronounced, suggesting more time between flood events.

## CONCLUSIONS

Sedimentologic analysis of five cores from the northwest part of Lisbon Valley that penetrate the lower part of the Chinle Formation and the upper part of the Cutler Formation provides excellent examples of sedimentary structures in the two units. The Cutler contains reddish-brown to purple-gray fluvial deposits interbedded with reddish-brown floodplain strata. These fluvial units alternate with reddish-orange to white eolian dune and sand-sheet strata and dark-reddish-brown sabkha deposits. The sand-sheet and sabkha deposits contain thin interdune pond deposits. The contrast in colors between the fluvial and eolian rocks is characteristic of the depositional environment and can be applied both in the cores and on the outcrop.

The Moss Back Member of the Chinle Formation unconformably overlies the Cutler Formation. The Moss Back is composed of greenish- to bluish-gray limestone nodule conglomerate, sandstone, and siltstone. Deposition was in high-sinuosity fluvial channel systems and crevasse splays on adjacent floodplains. The drab colors of the Moss Back reflect the high organic-carbon content of strata deposited and preserved below the water table. The color contrast between the Cutler Formation and the Moss Back Member of the Chinle is also distinctive, both on the outcrop and in the cores.

These cores from Lisbon Valley provide excellent examples of sedimentary structures and lithofacies sequences from several fluvial and eolian components of continental systems. Several of the fine-grained facies are better represented in the cores because of weathering of the facies on the outcrop. Many details of the sedimentary fabric in the Cutler, especially within fine-grained sabkha and interdune pond deposits, are visible in the cores but are not well preserved at the outcrop section. The cores, although lacking the advantage of lateral facies analysis, allow a vertical sequence analysis that includes details of fine-grained sabkha and sand-sheet environments that are poorly represented on the outcrop. The descriptions and interpretation of these depositional environments also provide a stratigraphic and environmental basis for future petrographic and geochemical studies of these units, both in outcrop and in the subsurface.

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# APPENDIX—DESCRIPTION OF SLABBED CORES

Core descriptions of the Lower Permian Cutler Formation, undifferentiated, and the overlying Upper Triassic Chinle Formation were recorded onto standardized forms that are reproduced here. The forms are divided into vertical columns that contain different types of information.

*Thickness/Sample no.*—This column is used to indicate thickness of the measured units in feet.

*Box no.*—Box numbers are given in this section.

*Formation/Member*—Formation and member names are shown in this column.

*Rock type*—Schematic representation of weathering profile of the outcrop, a lithologic symbol for rock type (symbols explained below), and sketches of sedimentary structures within the units are shown in this column.

*Color*—Both of these columns indicate color of units. Colors were estimated by a comparison with the Geological Society of America rock-color chart (Goddard and others, 1948). Where possible, colors were estimated from fresh, dry outcrops.

*Dominant grain size*—This column shows a continuous line chart of the dominant grain size of the measured unit. Grain size was estimated by a comparison to a standard grain size chart. Class divisions line indicate variations from the norm. V, very; Fn, fine; Sd, sand; Med, medium; Cse, coarse; Pbl, pebble.

*Bedding*—Bedding refers to set thickness of sedimentary units. VTK, very thick; TK, thick; MED, medium; TN, thin; VTN, very thin; MASS, massive.

*Sedimentary structures*—This column indicates the type of sedimentary structure that is shown graphically in the rock type column. CLL, curved, parallel laminations (trough or wedge-planar crossbeds); TAB. PLANAR, tabular-planar crossbeds; WLL, wavy lamination (flatbedding); ELL, even, parallel laminations (horizontal laminations); STRLESS, structureless.

*Biology/Organics*—This column indicates the presence of organic material, burrows, or bioturbation.

*Sorting/Roundness*—Sorting: VWS, very well sorted; WS, well sorted; MWS, moderately well sorted; FS, fairly well sorted. Roundness: A, angular; SA, subangular; SR, subrounded; R, rounded.

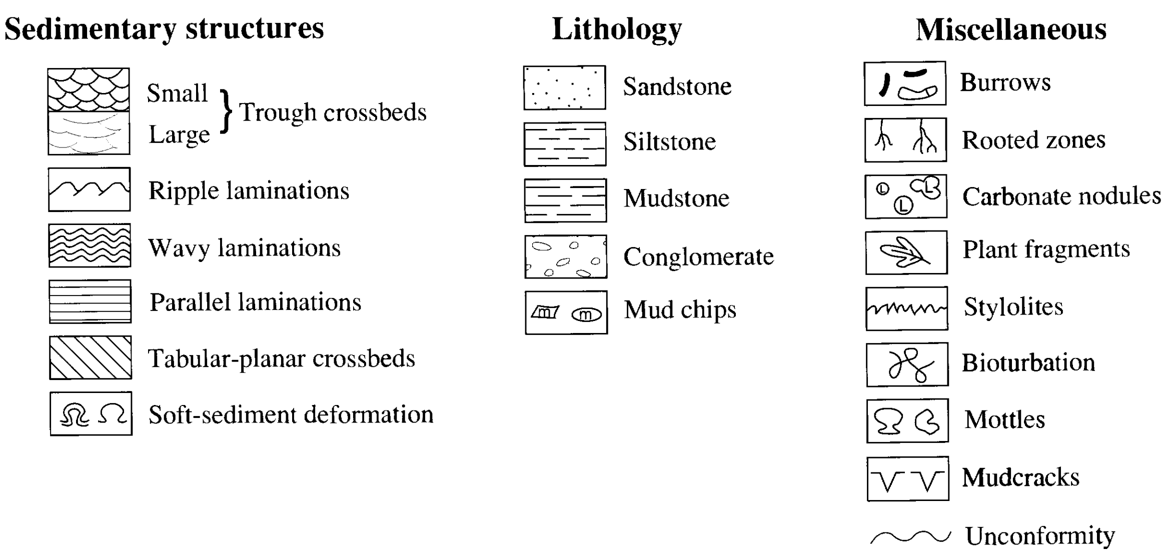
*Cement*—This column indicates the presence of calcite cement. VC, very calcareous; MC, moderately calcareous; SC, slightly calcareous; NC, noncalcareous.

*Accessory minerals or fragments*—Colors of unidentified accessory minerals or rock fragments are indicated in this column: BLK, black; GRN, green; GY, gray; WHT, white.

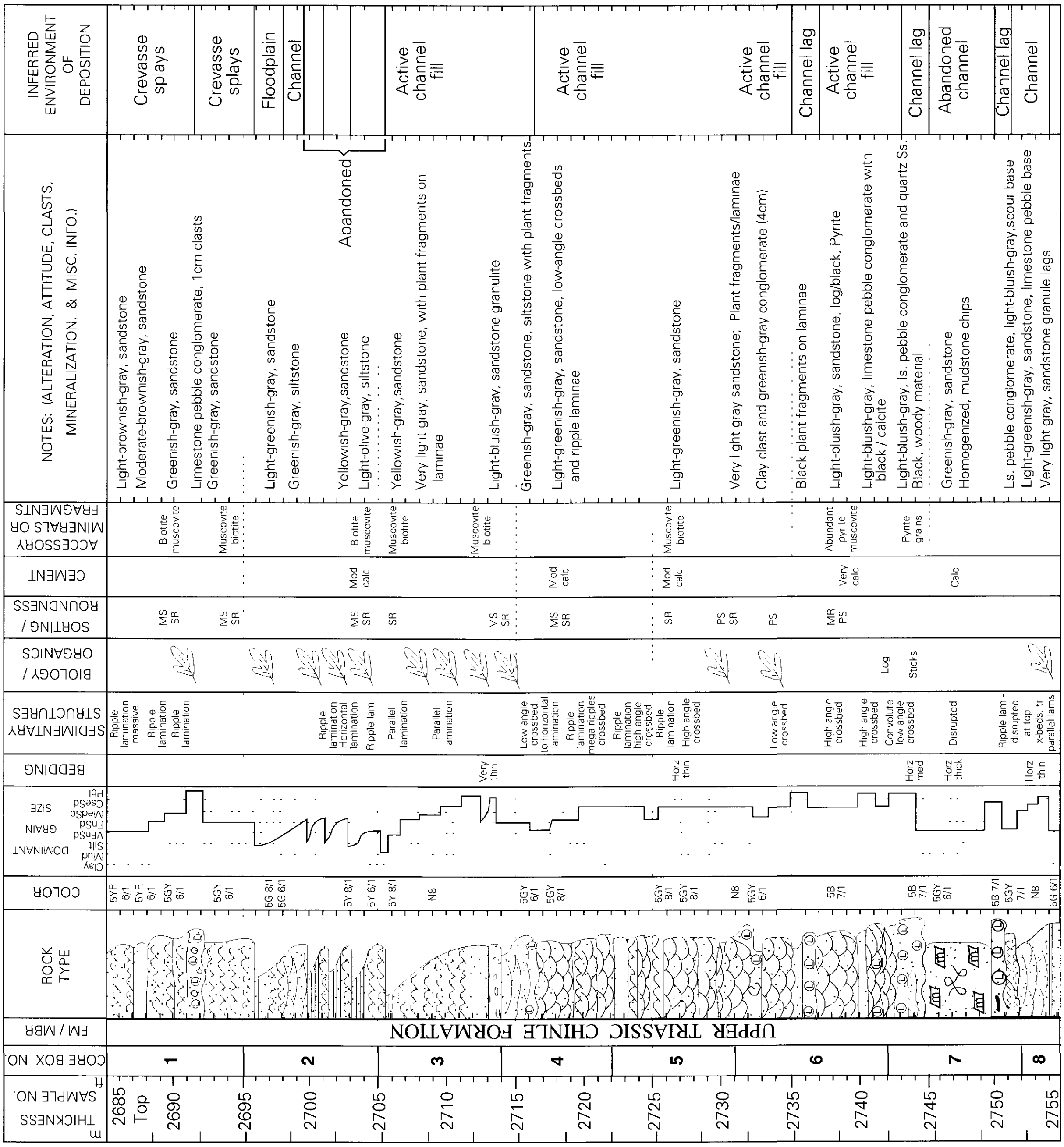
*Notes*—Additional comments and descriptions are given; circled abbreviations refer to sedimentary structures labeled on photographs in figure 4.

*Inferred environment of deposition*—Interpreted environments of deposition of the rock unit are shown in this column.

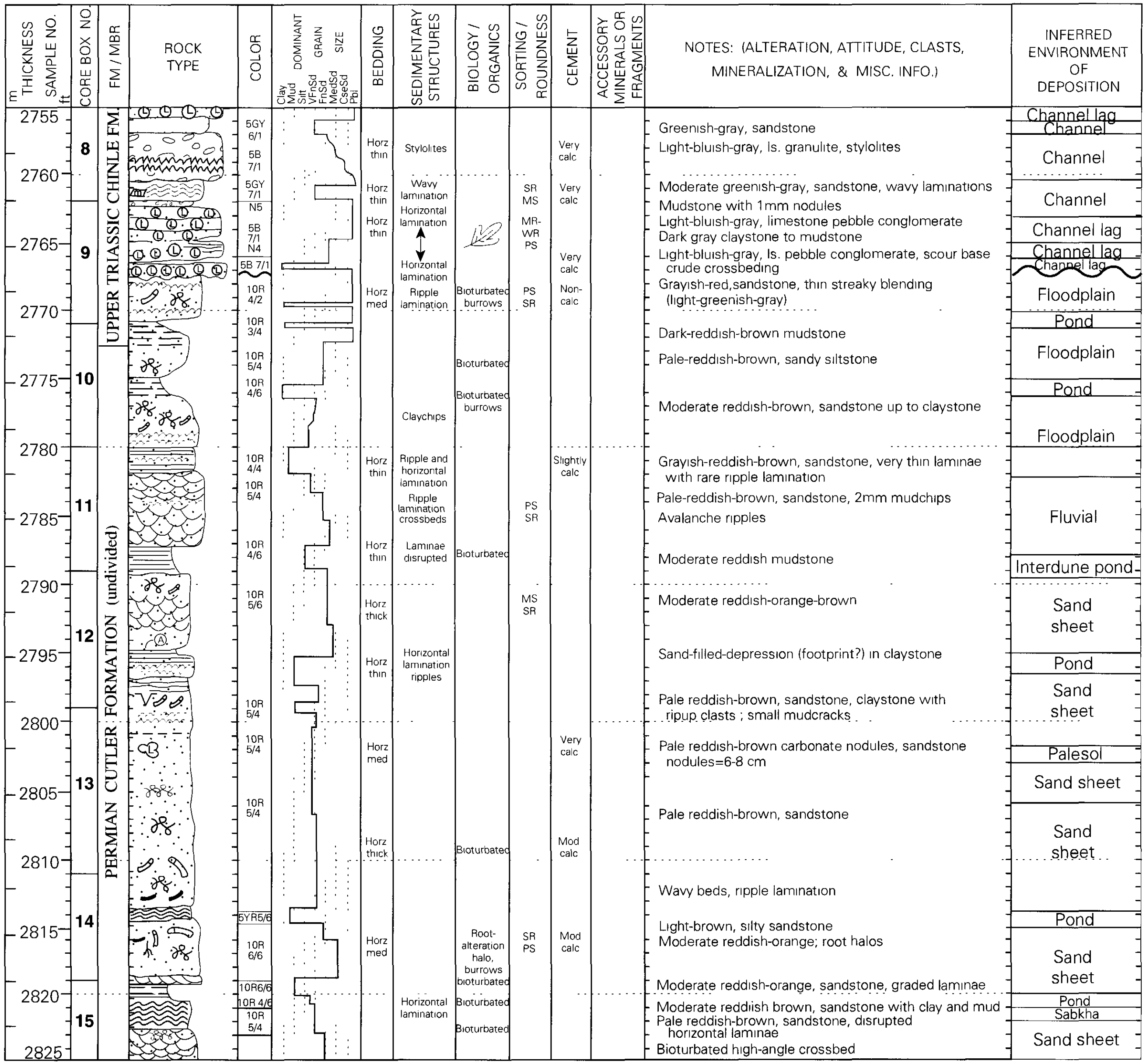
## EXPLANATION



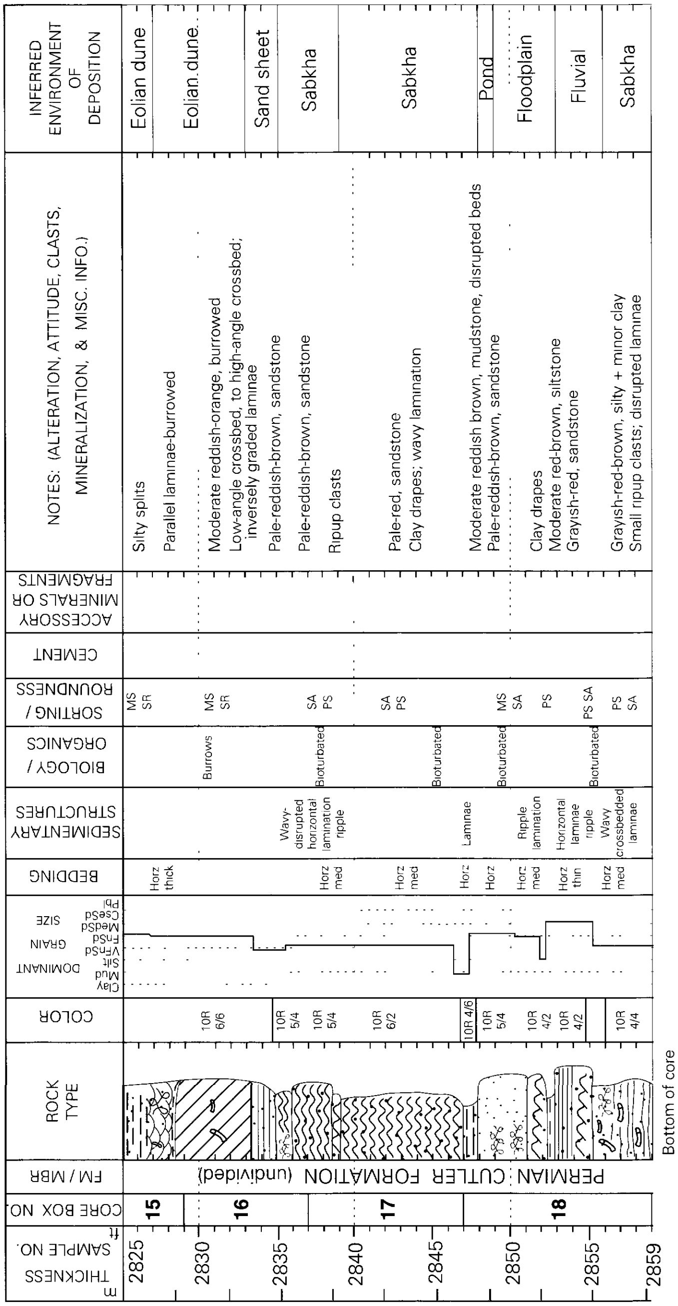
A. Core D729. Cored interval is 171 ft thick. See figure 4 for photographs of this entire cored interval.

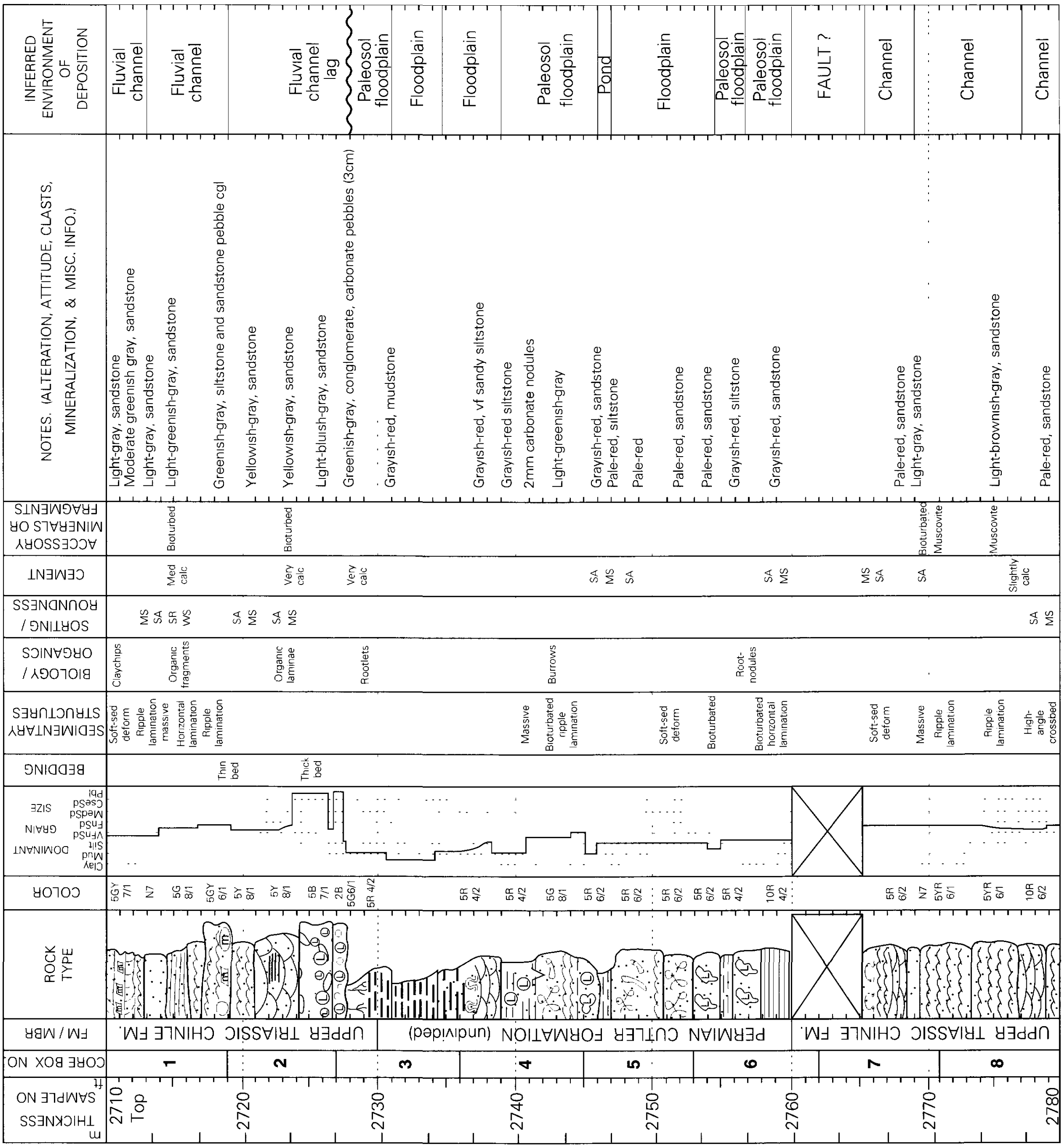


A. Core D729—Continued.



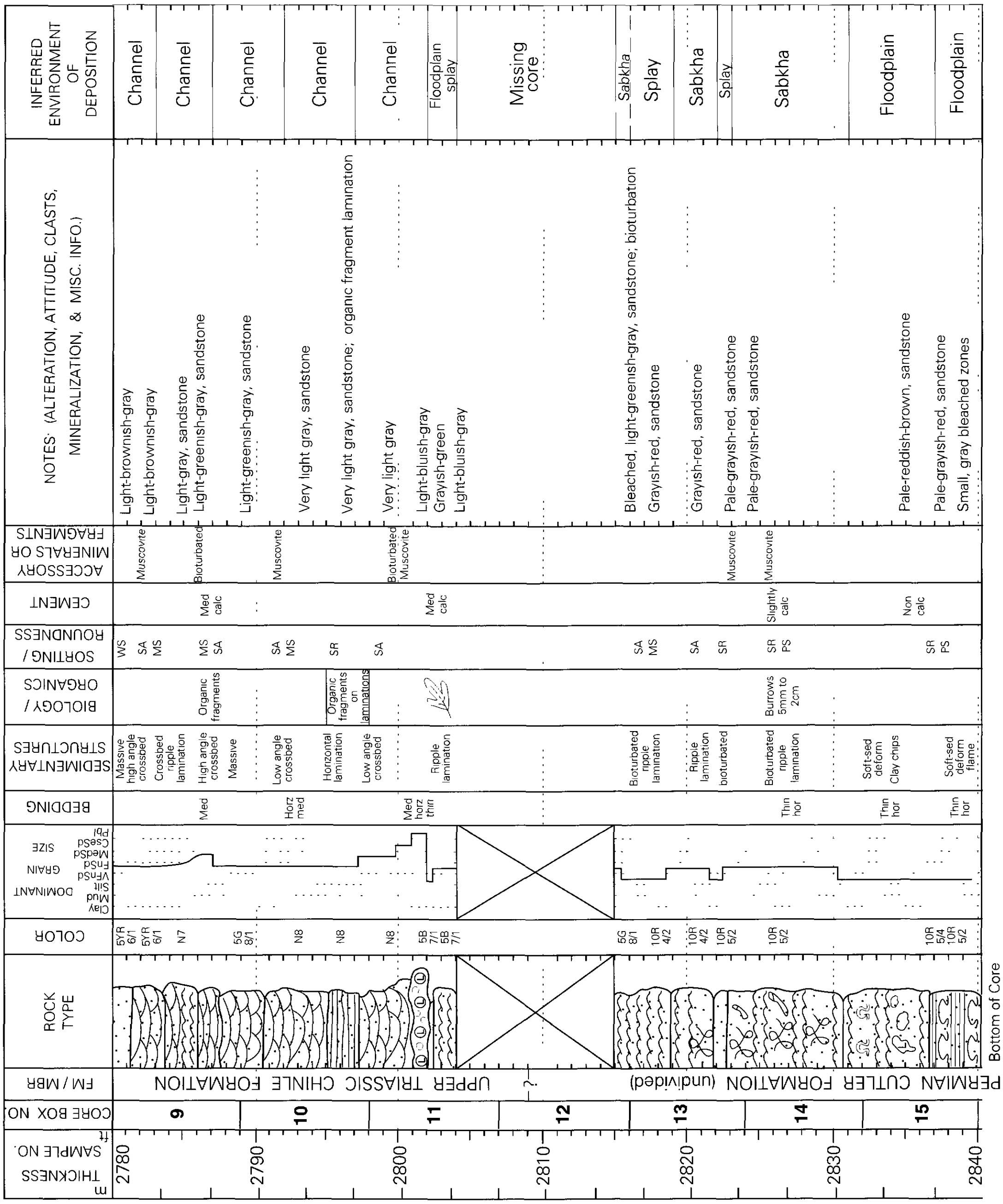
A. Core D729—Continued.

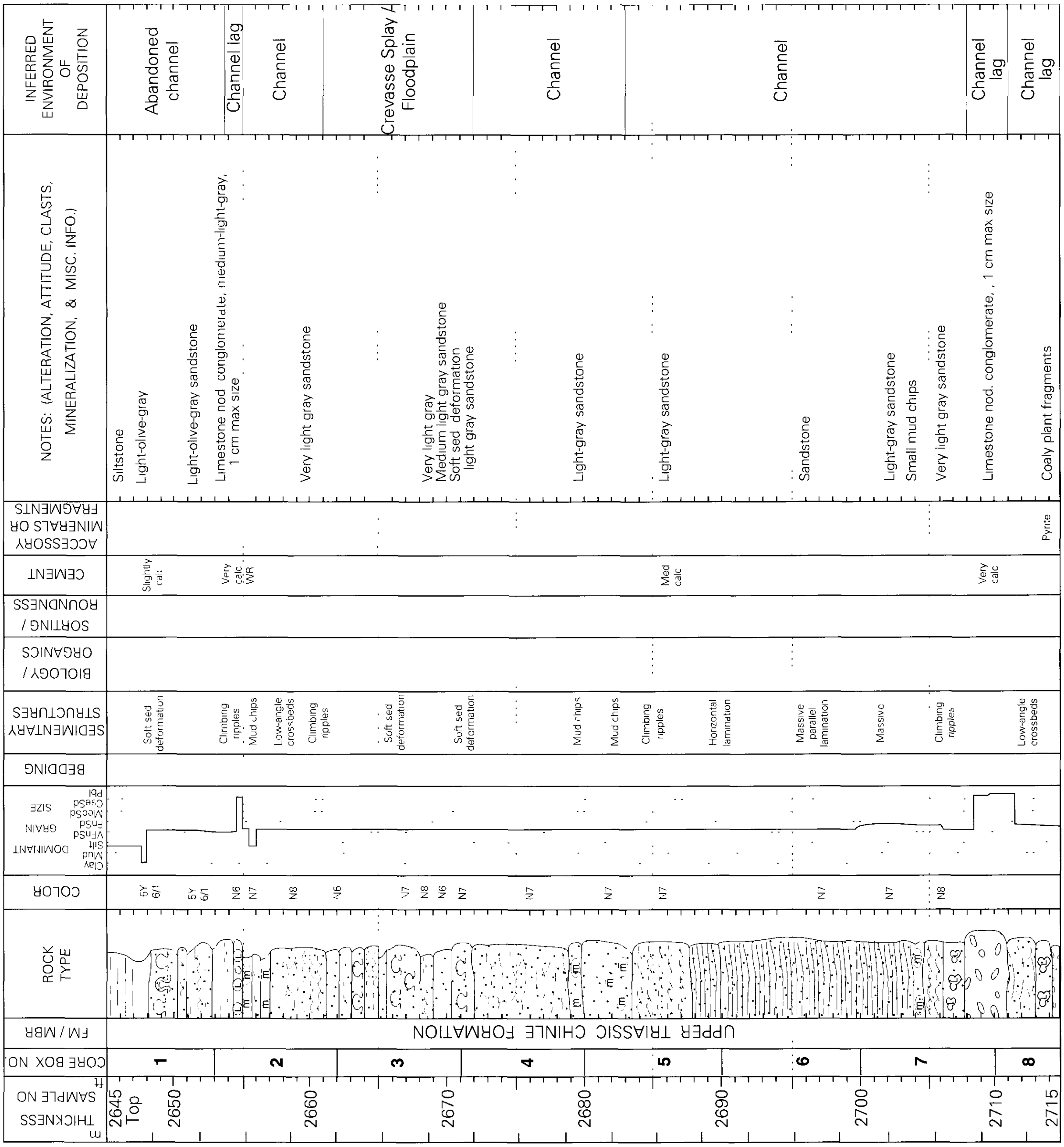





B. Core D742. Cored interval is 130 ft thick.

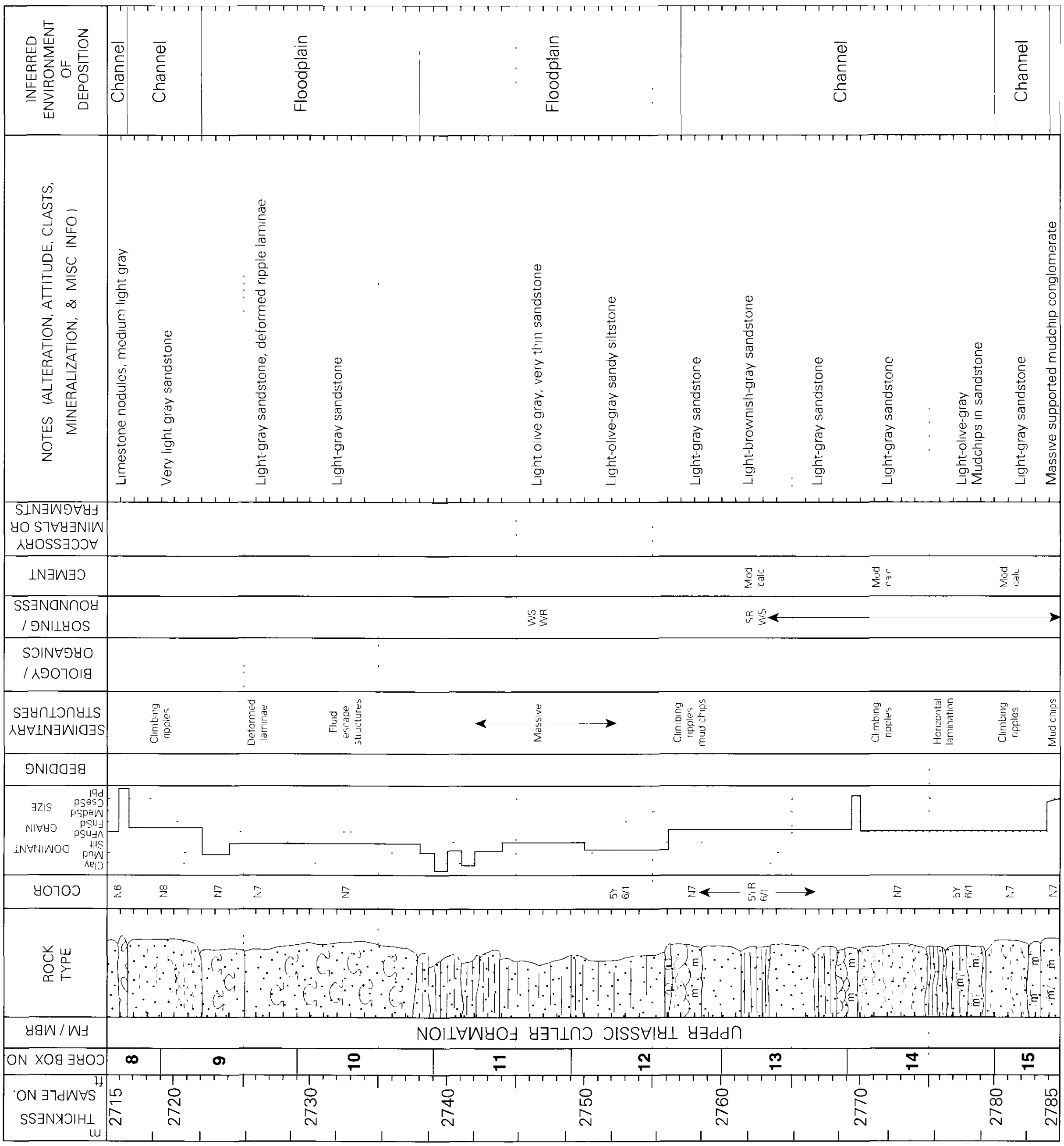
B. Core D742—Continued.

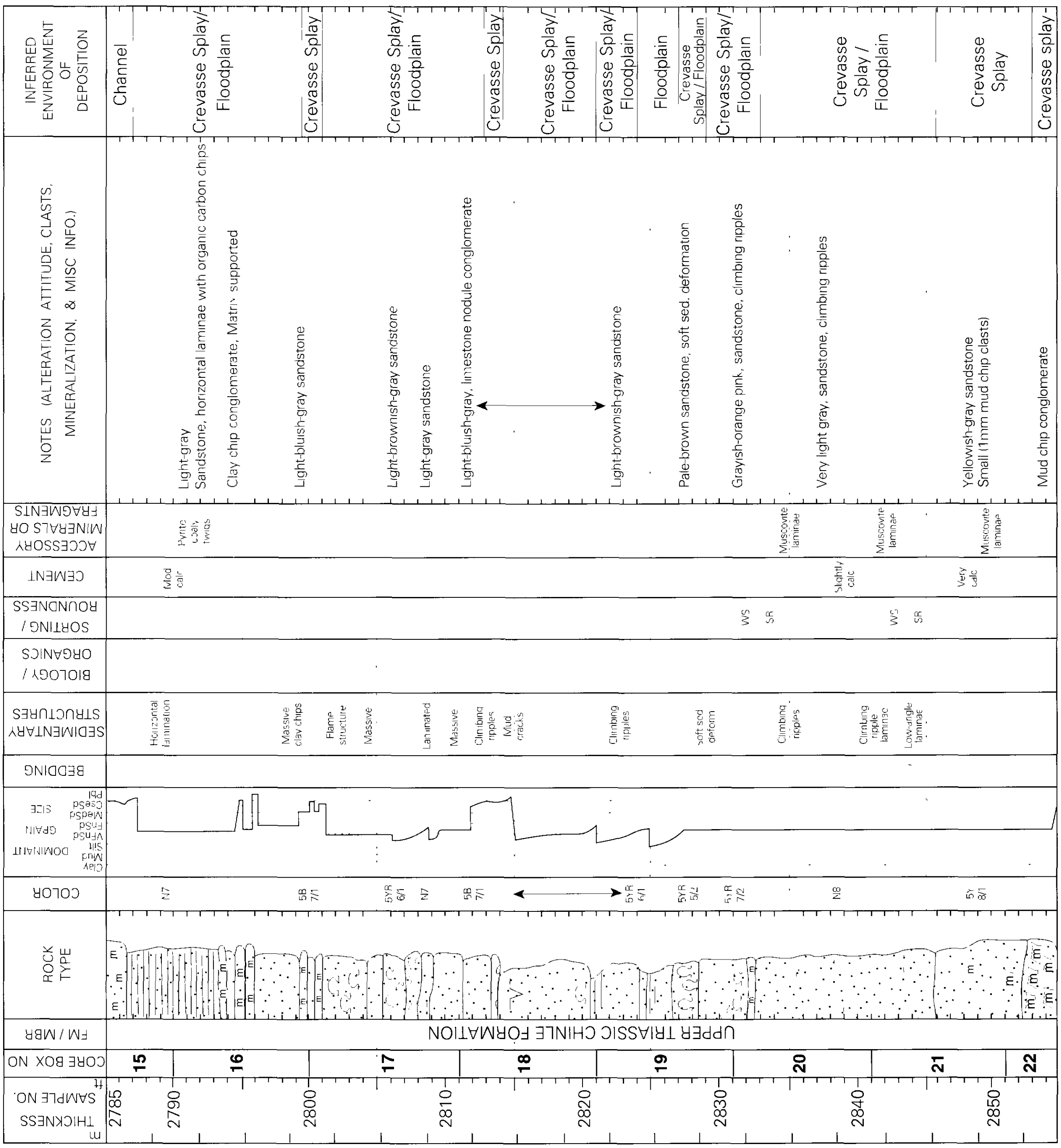




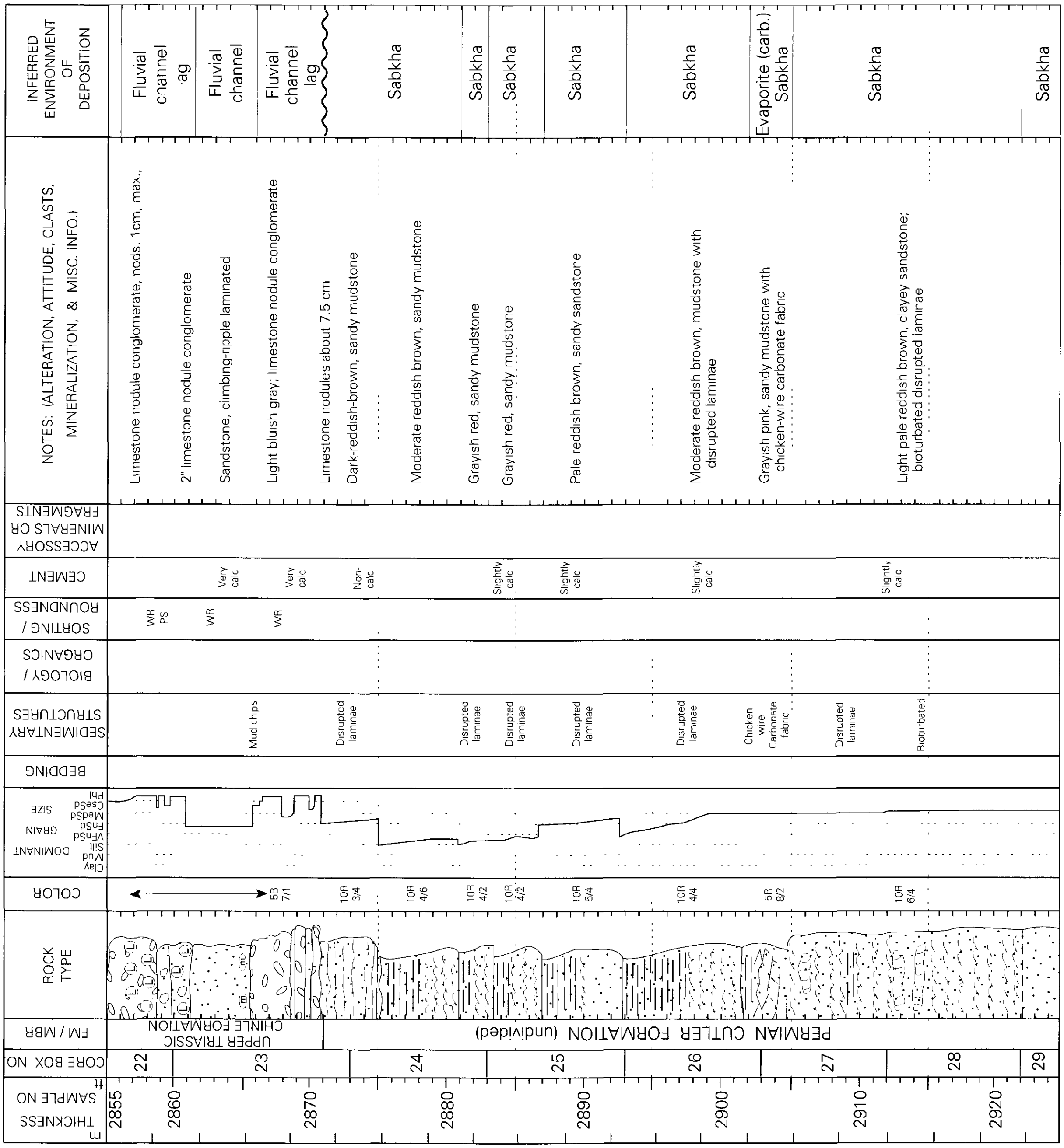

C. Core D892. Cored interval is 290 ft thick.

C. Core D892—Continued.



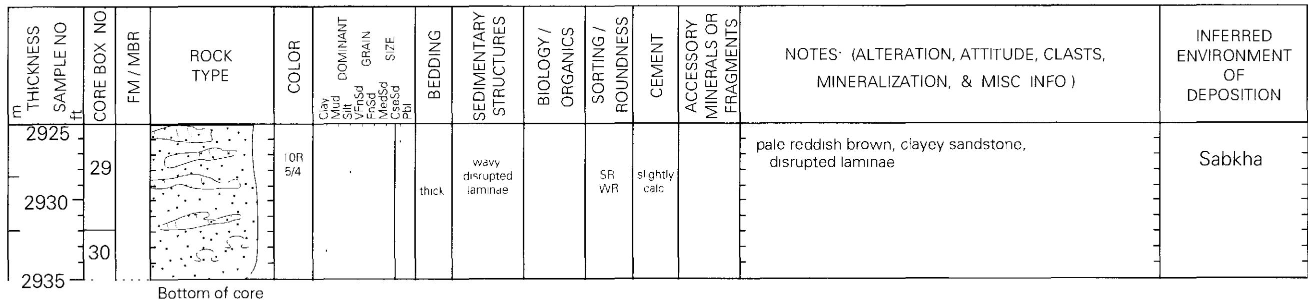


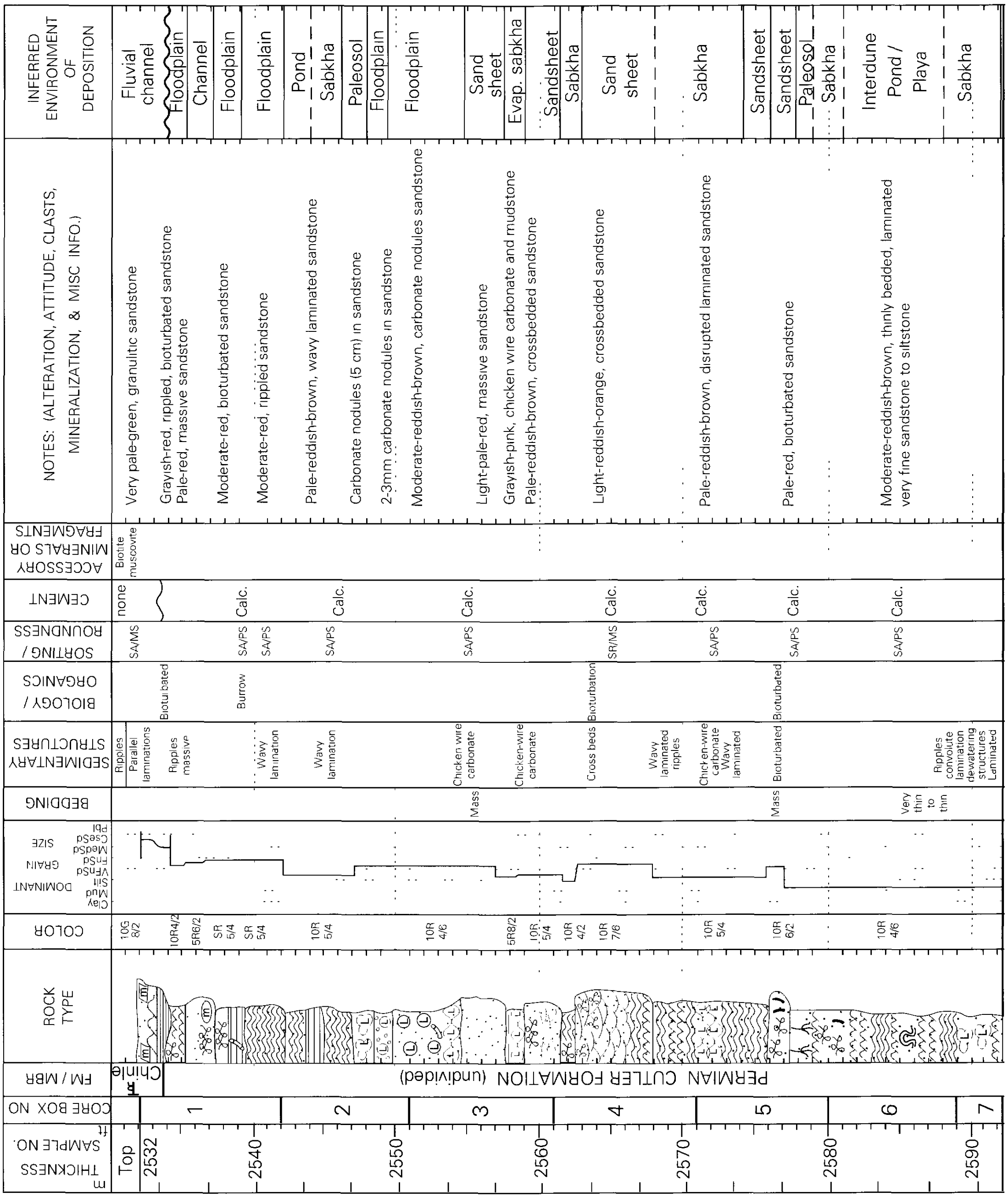

C. Core D892—Continued.



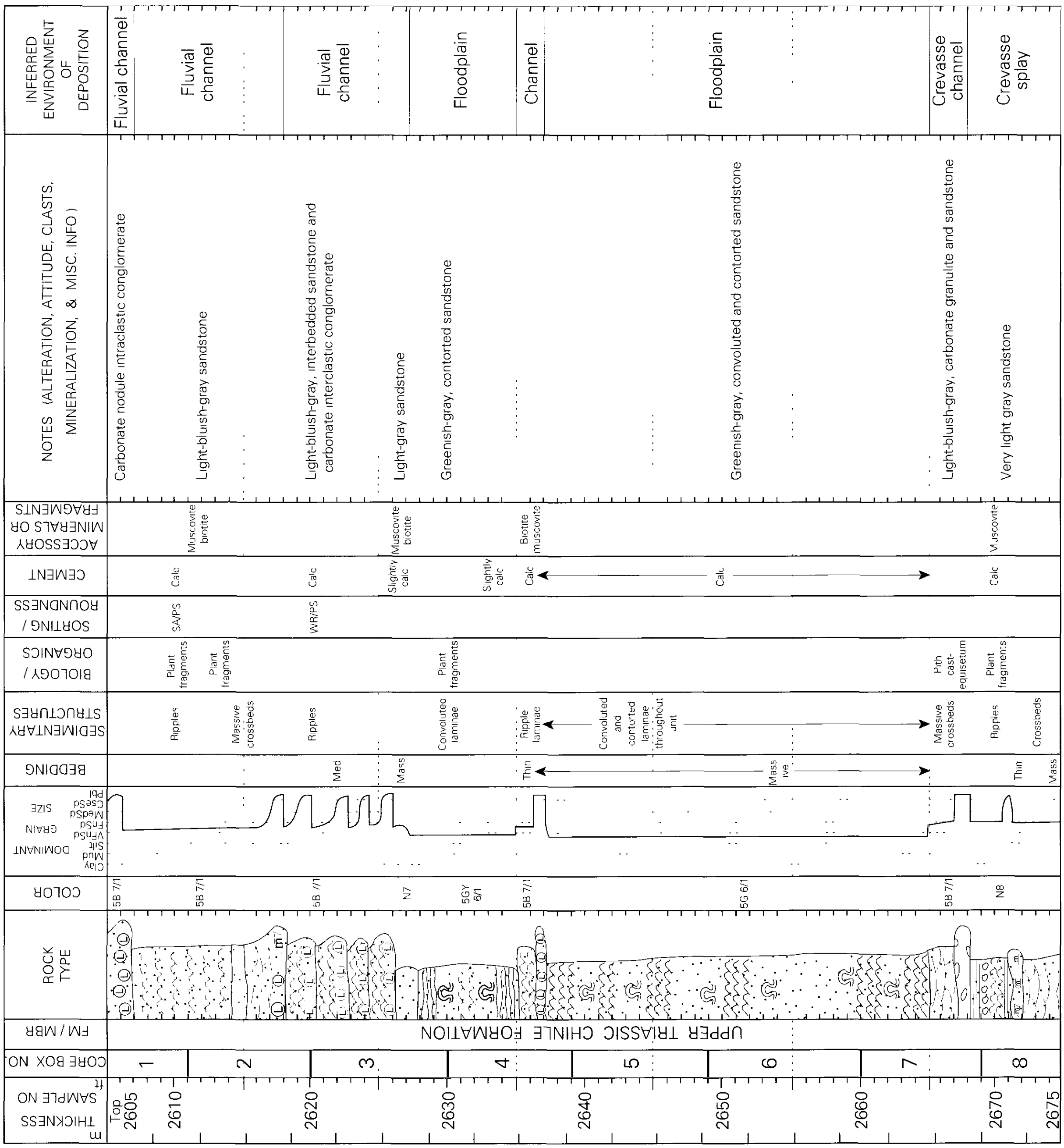

C. Core D892—Continued.

C. Core D892—Continued.

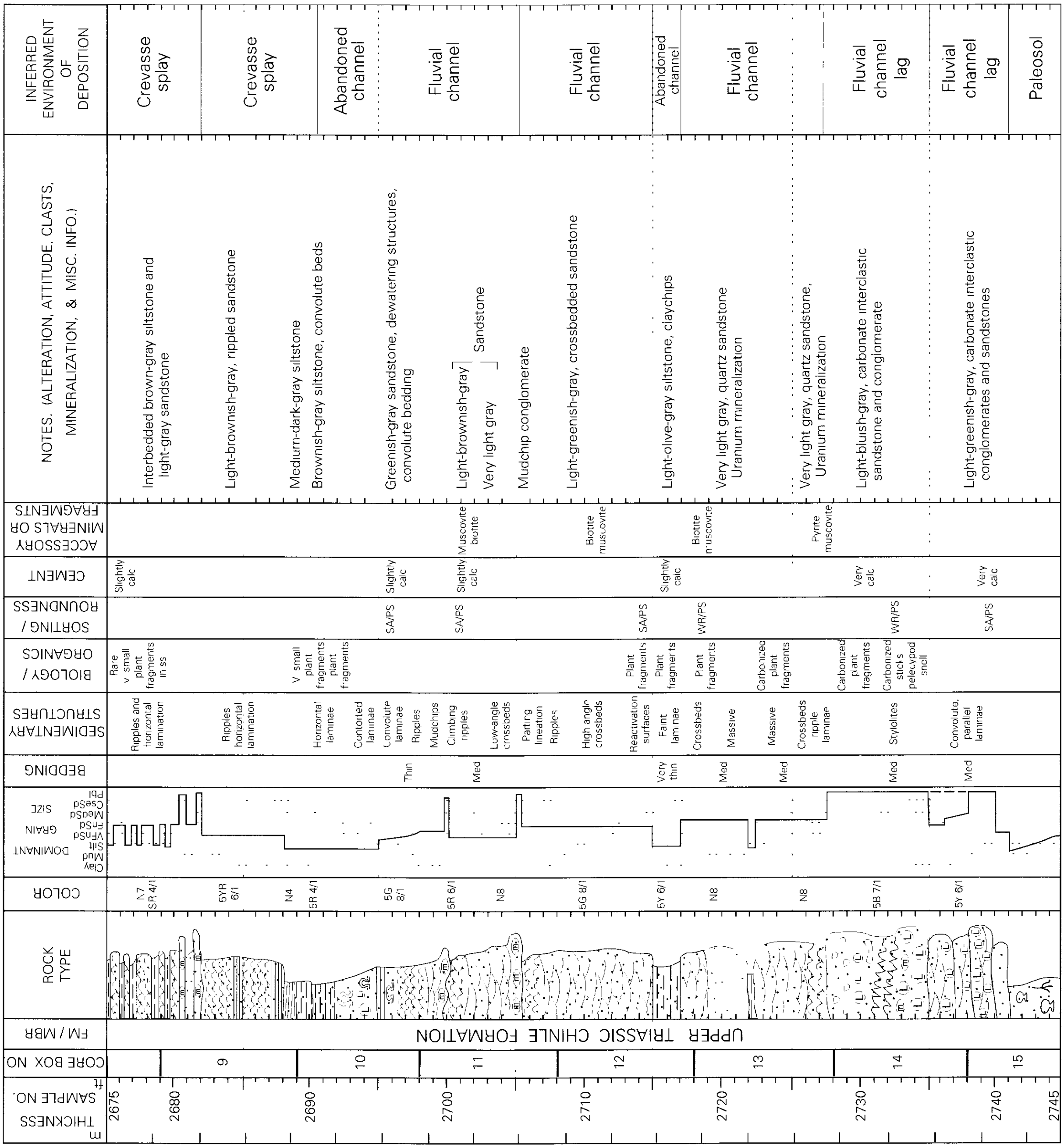





D. Core D769. Cored interval is 61 ft thick.

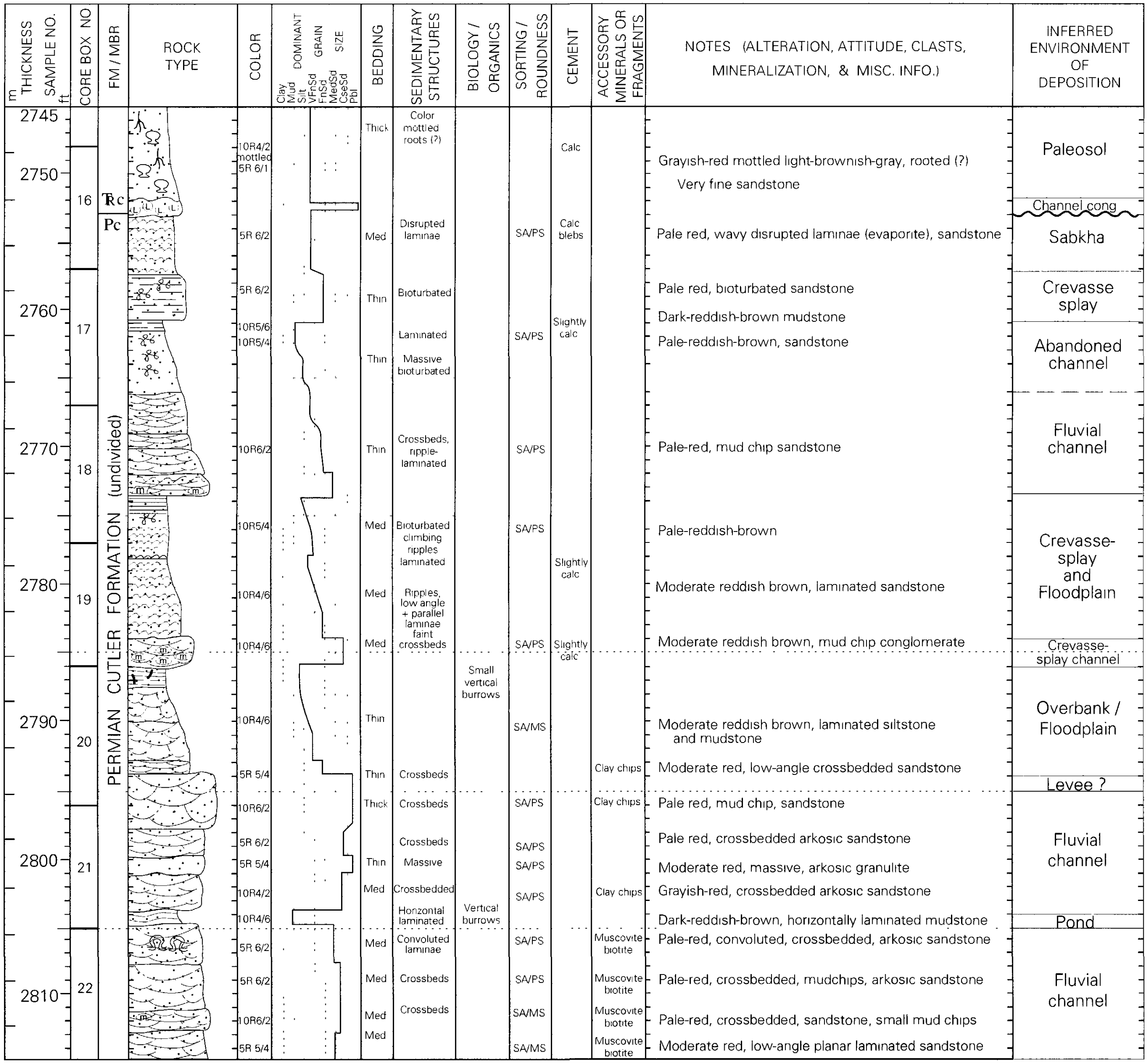



E. Core D788. Cored interval is 231 ft thick.




E. Core D788—Continued.

E. Core D788—Continued.



E. Core D788—Continued.