THE UNIVERSITY OF ALBERTA

Sedimentology, Stratigraphy and Ichnology of the Bluesky Formation in northeastern British Columbia

by

Harold Peter Oppelt

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled STRATIGRAPHY, SEDIMENTOLOGY, AND ICHNOLOGY OF THE BLUESKY FORMATION IN NORTHEASTERN BRITISH COLUMBIA submitted by HAROLD P. OPPELT in partial fulfillment of the requirements for the degree of MASTER OF SCIENCE.

S. George Pemberton

Supervisor

J. O. Scott

Date

Dedication

This thesis is dedicated, with love, to my wife, Joan.

ABSTRACT

The Bluesky Formation of early Albian age consists of a relatively thick succession of sandstones, mudstones and conglomerates that were deposited in shallow marine shelf bar and delta front settings. The Bluesky Formation, in the Peace River Plains, lays between the continental Gething Formation below and by the marine shales of the Moosebar Formation above. Shelf bar complexes are found primarily in the north half of the study area and attain thicknessess of over 30m. These shelf bar complexes were deposited at least 100 km from any recognizable paleoshoreline at middle to inner shelf depths. In the Foothills area of the study area, the Bluesky consists of a thick progradational succession of prodelta and delta front deposits which is overlain by the delta plain deposits of the Chamberlain Formation. The base of the Bluesky Formation probably records a sequence boundary between underlying non-marine Gething and marine Bluesky strata. Deposition of the Bluesky Formation began when the Gething delta plain was transgressed by the boreal Moosebar Sea, flooding most of northeastern British Columbia.

Eight lithofacies were defined in core and outcrop on the basis of physical and biogenic sedimentary structures and lithology. The offshore bar complexes are, in general, coarsening-upward successions which can be divided into: central bar deposits, composed of clean, trough cross-bedded sandstone; bar margin deposits (proximal and distal) deposits, composed of highly bioturbated, muddy sandstones; and interbar deposits, composed of bioturbated, silty mudstones. The finer-grain and bioturbated bar margin sandstones were probably deposited lateral to the higher energy portions of the ridge and probably represent periods of slow sedimentation and lower energy. Proximal bar margin deposits contain abundant infaunal suspension-feeding structures characteristic of the *Skolithos* ichnofacies; whereas, distal bar margin deposits contain predominately epifaunal deposit-feeding structures, characteristic of the *Cruziana* ichnofacies. Silty mudstone interbar sediments were deposited lateral to the bar proper in a very low energy environment.

The prodelta deposits consist of a rhythmically bedded series of cross-laminated sandstones and burrowed mudstones implying a seasonal cyclicity of sediment input. Delta Front sediments, composed of small-scale cross-laminated sandstones, were deposited as distributary mouth bar and wave-reworked sheet sands. The presence of ichnofossils common to both the *Cruziana* and *Skolithos* ichnofacies is characteristic of a Brackish Water ichnofacies. The regressive deltaic succession is capped by the coal-bearing delta plain sediments of the Chamberlain Formation. Continued transgression of the Moosebar Sea flooded the entire basin allowing deposition of quiet water shales over the Bluesky and Chamberlain intervals.

Acknowledgements

Sincere thanks are expressed to Dr. S.G. Pemberton from the University of Alberta for his support, guidance and, especially patience during the preparation of this thesis. The author would also like to thank the British Columbia Ministry of Energy, Mines and Petroleum Resources, notably W. Kilby, for their support. Appreciation is also extended to C. Gendron for her encouragement on numerous occasions when it was needed the most.

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1.0 INTRODUCTION

Lower Cretaceous Bluesky Formation sediments found in the subsurface of the Peace River Plains can be traced into the inner Rocky Mountain Foothills of northeastern British Columbia. In the plains region, Bluesky sandstones produce significant quantities of oil and gas from a series of structural and stratigraphic traps; equivalent Bluesky strata in northwestern Alberta also contain large hydrocarbon reserves. Interest in exploring for hydrocarbons in the Bluesky Formation began in Alberta as early as the late 1940's. Since that time a number of large oil and gas pools have been discovered in both northwestern Alberta and northeastern British Columbia. Estimated resource potential for the Bluesky Formation in northeastern British Columbia is placed at 18 x 10^6 m^3 of gas and 2 x 10^6 m^3 of oil (Dewar, 1977).

The Bluesky Formation is also used as a lithostratigraphic datum for the correlation of coal measures in the Peace River coalfields. Within the Foothills structural belt of the study area, the potential for gas entrapment is considered to be low since most of the lower Cretaceous section has been brought to the surface by extensive folding and faulting. However, despite the economic significance of the Bluesky Formation, very little is known of its distribution or sedimentological characteristics. Comprehensive regional lithofacies and biofacies analyses are not yet available for the Bluesky Formation in northeastern British Columbia. As a result, regional stratigraphic correlations are inconsistent.

It is important, then, to gain a better understanding of this poorly documented and complex unit so as to provide better sedimentologic models in exploring for hydrocarbons in the Bluesky Formation. In addition, it provides an excellent opportunity to study in detail the sedimentology and ichnology of a shallow marine sequence.

1.1 PURPOSE OF THE STUDY

The geologic record for the Cretaceous Period in the western interior of North America contains numerous examples of marine shelf sedimentation. Many of these deposits are lenticular sandstone bodies encased in mudstones, such as the Upper Cretaceous Shannon Sandstone in the Powder River basin (Spearing, 1976). Traditionally, these sand buildups have been interpreted as shallow marine bars and are frequently employed as models for the interpretation of other shallow marine sandstones (Johnson and Baldwin, 1986). Many of these sand bodies occur hundreds of kilometers offshore, far from any recognizable time-equivalent paleoshoreline and are suggested to have been molded by tidal and storm processes. However, the exact mode of how sand can be transported across a muddy shelf and emplaced into discrete sand bodies is still an enigma. It has recently been suggested, though, that similar units, such as the Cardium Formation of central Alberta, represent shoreline deposits which were stranded offshore as a result of a rise in sea level and concomitant retreat of the coastline (Bergman and Walker, 1986, 1988).

Recent studies of modern shelves - the U.S. Atlantic shelf (Stubblefield *et al.*, 1984), the Alaskan shelf (Howard *et al.*, 1981), and the North Sea (Houboult, 1968) - also describe sand bodies which have originated, in part, in a setting different than their present one. These sand bodies are considered to have formed in a nearshore environment and are now stranded many kilometers from the present coastline as a result of the recent Holocene rise in sea level (Walker, 1984). However, these sand bodies continue to be modified and shaped accordingly by modern tidal or oceanic processes indigenous to their areas, losing their former nearshore sedimentologic characteristics.

The Bluesky sandstones, in some regards, also share the same controversy in origin. The thick sandstone buildups of the Peace River region may have originated as shoreface deposits which were stranded after a sea level change or they may represent shallow marine bar sedimentation. Nevertheless, this debate in depositional origin provides an interesting problem which may be resolved by examining the sedimentologic and stratigraphic framework of the Bluesky Formation.

In order to establish the geometry, sedimentary structures and depositional history of these sediments, this study will examine the stratigraphy, sedimentology and ichnology of the Bluesky Formation in northeastern British Columbia. Specifically, the objectives of this study are:

1) to define the regional stratigraphic relationships between the Bluesky, Chamberlain and Moosebar Formations.

2) to describe and analyze the lithofacies present within the Bluesky Formation,

3) to determine the ichnogenera present, their associations and their significance within the clastic sequences,

4) to correlate and map the distribution of relatively thick, discrete sand bodies within the study area.

5) to determine the environments of deposition, including the method of sand transport onto the shelf,

6) and to reconstruct the paleogeography of the study area prior to-, during-, and after-Bluesky time.

1.2 LOCATION AND PHYSIOGRAPHY

The study area is located in northeastern British Columbia and covers the National Topographic System mapsheets of 93P, 94A, and the majority of 94H. Peripheral areas include the eastern halves of 930, 94B, and 94G mapsheets. It is bounded on the south and north by the 54^{o} and 57.5^{o} lines of latitude north, respectively, and by the British Columbia - Alberta border to the east. The total study area is on the order of 42,000 square kilometers (Figure 1.1).

The study area includes part of the Western Interior Plains and the Rocky Mountain Foothills physiographic provinces (Figure 1.1). For ease of reference throughout the study, the plains region will be referred to as the Peace River Plains. Much of the Peace River Plains is immediately underlain by early Cretaceous undeformed shales and is, therefore, relatively flat. The subdued topography is marked by only gentle undulations and is covered by muskeg or dense forest. The Foothills province ranges from 50 to 90 km in width, and is underlain by mildly folded and faulted Cretaceous rocks in the outer Foothills and more severely deformed older Mesozoic and Paleozoic strata in the inner Foothills. The topography of the Foothills generally reflects the underlying structure with broad synclinal valleys and sharp anticlinal hills.

The Lower Cretaceous strata of the Foothills belt have been brought to the surface by faulting along the leading edge of the Rocky Mountain Fold and Thrust Belt. In the subsurface of the Peace River Plains, the Bluesky unit has a regional dip towards the southwest (Figure 1.2). The deepest top of the Bluesky Formation is 1860 m below sea level (gas well 93-I-16 b-28-J).

1.3 DATA BASE

The data base for this thesis uses a combination of coal exploration diamond drill core, petroleum well core, geophysical logs from both types of well borings and outcrop exposures. Rarely is outcrop, coal, gas and oil well data integrated to broaden the geologic data base. Yet northeastern British Columbia has a wealth of geologic information stored in a collection of well cores, diamond drill cores, and data files.

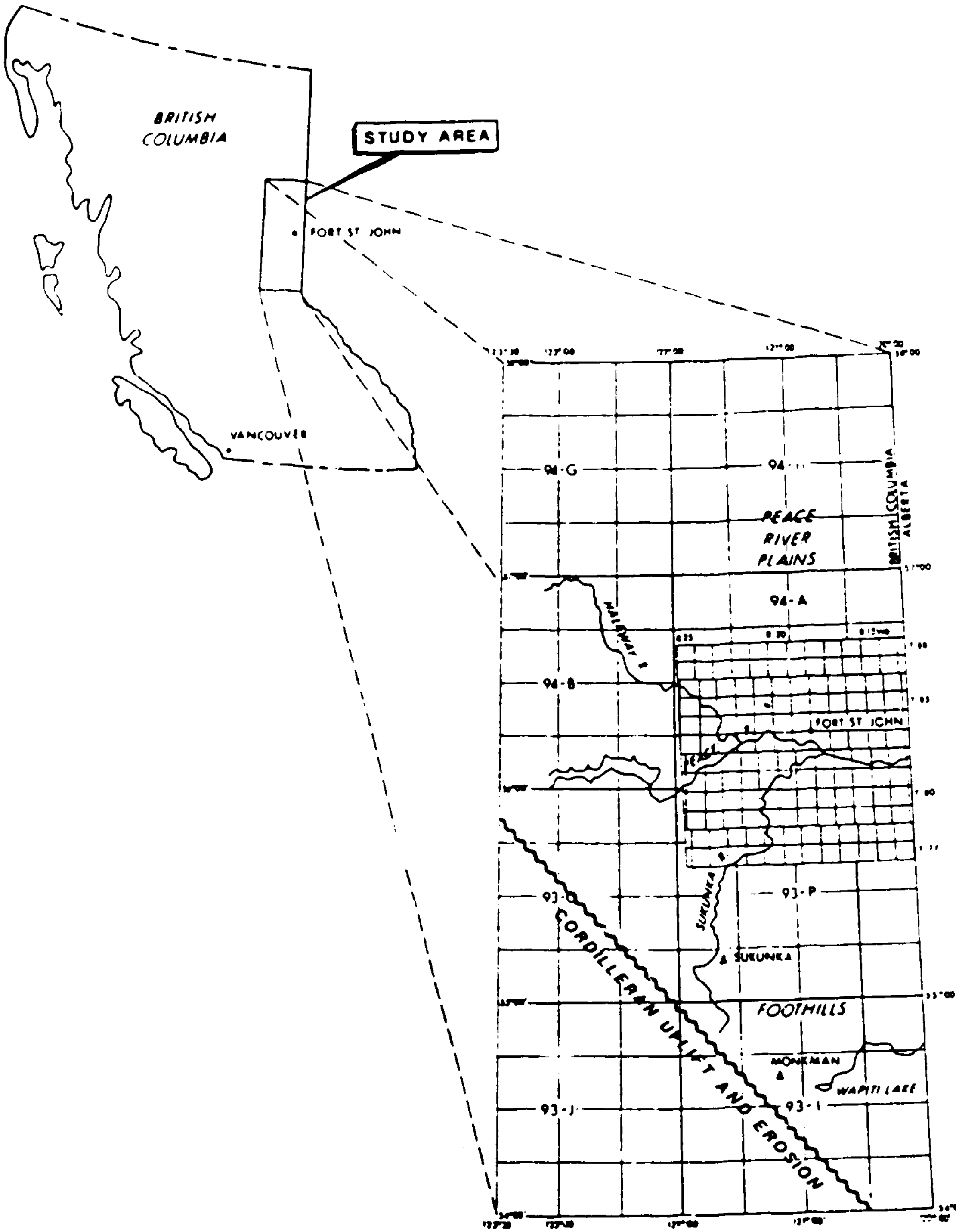

Figure 1.1: Study area map indicating geographical subdivisions of the Peace River Plains area and Foothills area referred to in text.

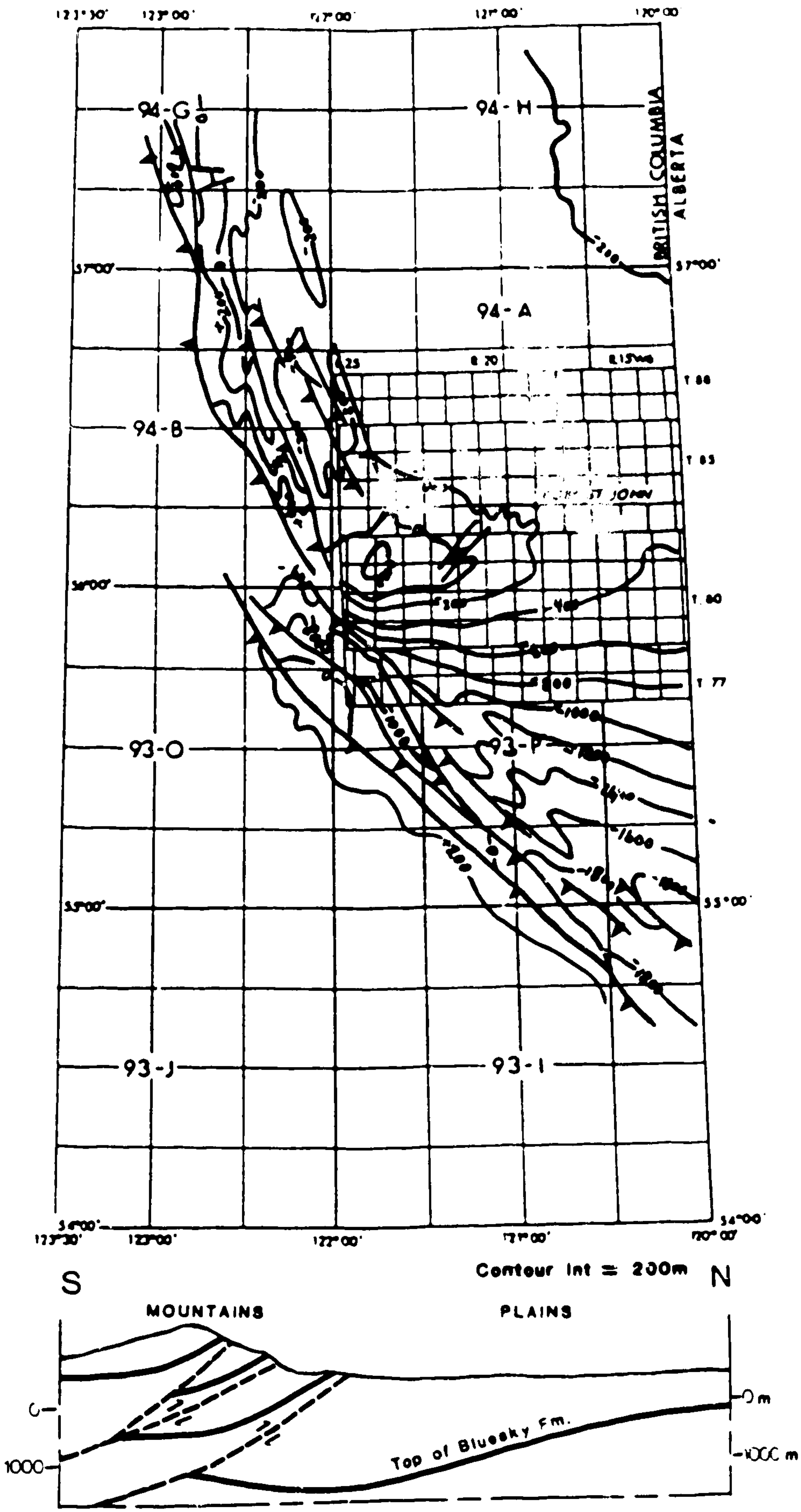

Figure 1.2: Structural map and cross-section of the Bluesky Formation from the Peace River Plains subsurface to the Rocky Mountain Foothills.

For this study, a total of 78 cores were examined, of which 57 were petroleum well cores and 21 were coal exploration diamond drill cores (Figure 1.3). In total more than [illegible] of core was described for sedimentological, lithological and ichnological features.

Outcrop and coal exploration diamond drill cores are strictly limited to the foothills belt. Outcrop exposure of the Bluesky Formation is fairly sparse because of its easily erodable lithology. Although the more resistant conglomerate beds found at the base of the sequence are better exposed, even those are exposed in two dimensions. Thus, data control in the foothills is gained best by coal exploration diamond drill intersections. Descriptions for the Chamberlain Formation strata were also gathered from diamond drill intersections.

Oil and gas wells are generally restricted to the undeformed belt of the Peace River Plains area. Whenever possible, well cores were supplemented with geophysical logs. Isopach plots rely greatly on geophysical logs for their data base. Generally, each township and range is represented by one well log. The thickness of Bluesky strata was interpreted using gamma-ray and sonic log signatures. Self-potential logs in this area generally do not provide the resolution required to analyse complex and interbedded lithologies.

1.4 PREVIOUS WORK

1.4.1 GENERAL

The Bluesky Formation was first recognized formally by the Alberta Study Group (1954) as "... a marine sandstone which occurs in wells between the Bullhead group below and the Spirit River Formation above". Its name is derived from the Bluesky Well No. 1 (Shell-British American 4-29-81-1) in the Peace River region of Alberta. The Alberta Study Group described the unit as

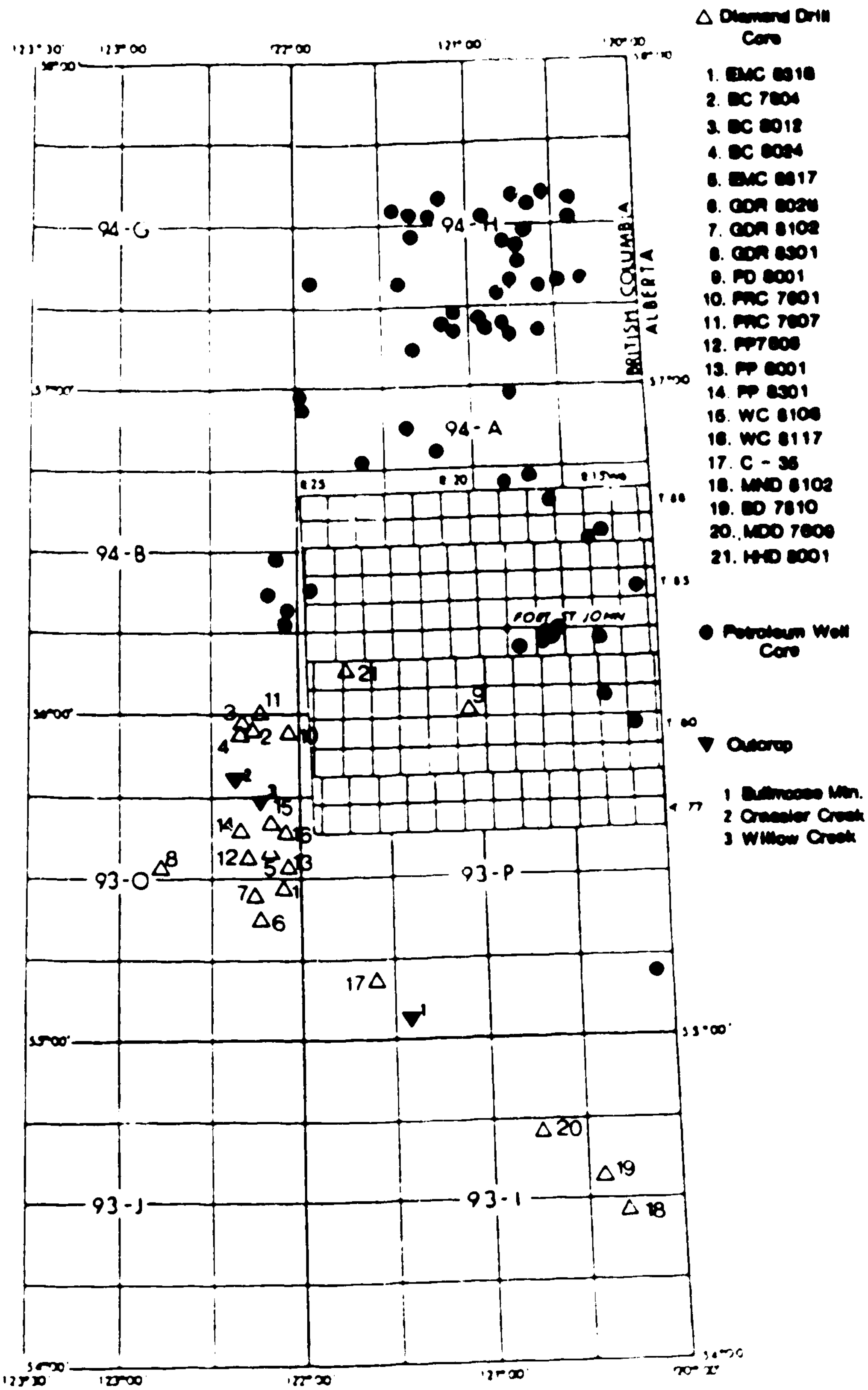

Figure 1.3: Location of cored intersections and outcrop examined in the study area

a fairly porous, glauconitic and pebbly sandstone with minor interbeds of shale. At the base, an ostracod-bearing black shale was tentatively correlated with the Ostracod Zone of central Alberta (Alberta Study Group, 1954).

The Bluesky Formation has never achieved formational status in northeastern British Columbia. The sandstones which comprise it have been traditionally considered as only the basal sandstone unit of the Moosebar Formation and its equivalent the Wilrich Member of the Spirit River Formation. Pugh (1960) proposed that the name Bluesky Formation be informally used for the basal sandstone of the Moosebar Formation in northeastern British Columbia arguing that it was lithologically and stratigraphicallyequivalent to the Bluesky succession found in Alberta.

The origin of the Bluesky Formation has been described, in previous literature, as a thin nearshore lag deposit formed during the early transgressive stage of the Moosebar Sea (Pugh, 1960; Rudkin, 1964). Although this may be accurate in some locales, there are occurrences of thick, coarsening-upwards marine sandstones which are difficult to explain depositionally strictly using the transgressive lag proposal.

Stott (1968) briefly discussed the stratigraphic relationship of the Bluesky Formation to the Gething Formation in the Peace River area by addressing what he termed the 'Gething-Bluesky Problem'. Stott suggested that the Bluesky Formation is either: (1) a facies equivalent to the Foothills Gething Formation; or (2) the earliest near-shore unit of the Fort St. John shales that overlys the Gething Formation. Stott (1968) indicated that the Bluesky Formation was probably a facies equivalent to the upper beds of the Gething Formation. Bluesky Formation strata, up to this point, had never been recorded within the Foothills belt.

Duff and Gilchrist (1981) were the first to document the presence of a marine unit interbedded within the beds of the mostly continental Gething Formation in the Sukunka-Wolverine areas of the Foothills. Recognition of this marine tongue was largely based on the identification of marine body fossils by C.R. Stelck. Consequently, Duff and Gilchrist (1981) divided the Gething Formation into three units: (1) the Lower Gething Formation, (2) the Gething Marine Tongue, and (3) the Chamberlain Member of the Gething Formation (Figure 1.4). According to Duff and Gilchrist, the Chamberlain Member merges with the Lower Gething Formation in the Sukunka region, indicating that the Marine Tongue is not present south of the Wolverine area. In a correlation section by Duff and Gilchrist (1981) the top of the 'Lower' Gething Formation in the Peace River region is correlated to the top of the Chamberlain Member in the Sukunka-Wolverine area (Figure 1.5). However, it is suggested, in this study, that the top of the Gething Formation in the Peace River area should be correlated to the top of the Lower Gething Formation (Duff and Gilchrist, 1981) in the Foothills. Also, it is suggested that the Bluesky Formation sediments of the Peace River Plains is equivalent to the 'Gething Marine Tongue' of Duff and Gilchrist (1981).

Kilby (1984) briefly described the distribution and stratigraphy of the Bluesky Formation within the Foothills of northeastern British Columbia. Kilby (1984) indicated that the marine unit had a coarsening-upward profile on gamma ray logs and could be subdivided into three distinct units. A recent study of Lower Cretaceous stratigraphy in Alberta and northeastern British Columbia by Smith *et al.*, (1984) provided a brief description of the stratigraphy and sedimentology of the Bluesky Formation. Smith *et al.*, (1984) stated that the Bluesky Formation consisted of a basal transgressive lag, followed by the growth of a progradational barrier-bar succession in the Elmworth area, and offshore bar complexes in the Peace River Plains.

Jackson (1984), in a study of the Mannville Group in the Deep Basin, recognized the development of coarsening-up deltaic sediments in the Elmworth area and offshore bar cycles, built

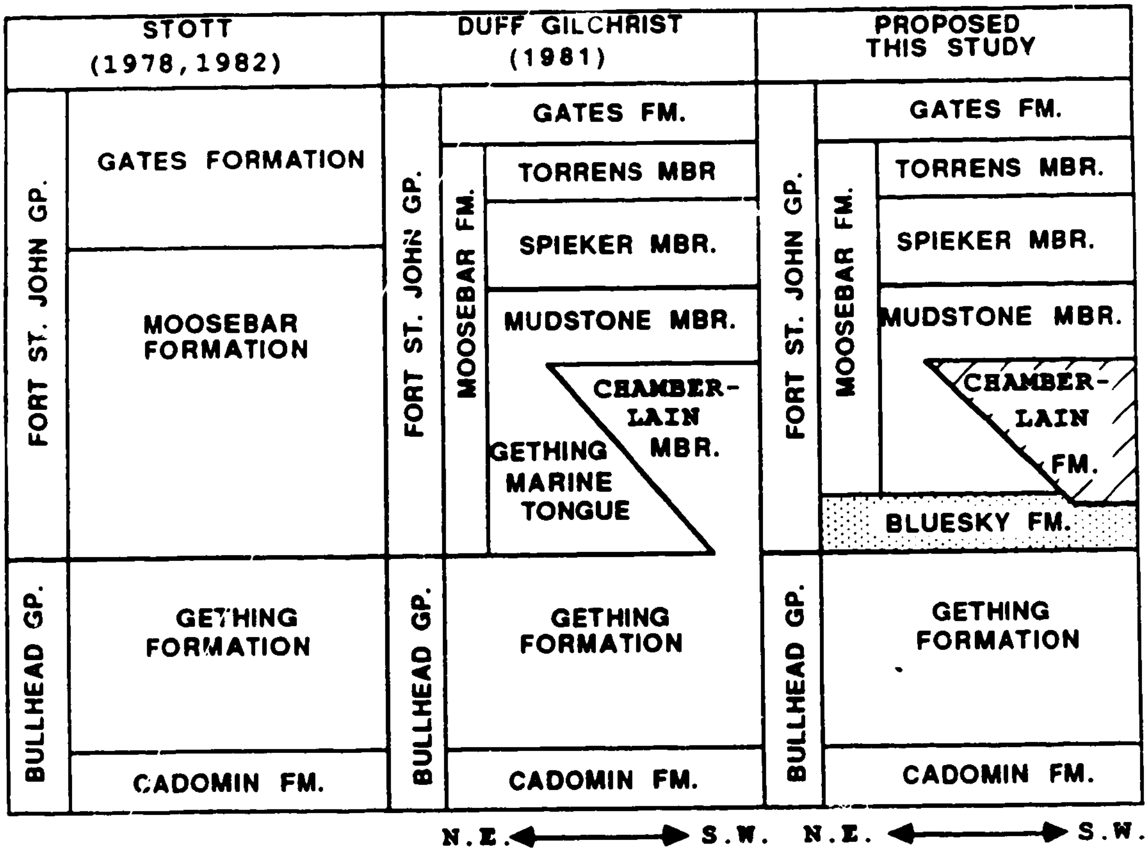

Figure 1.4: Correlation chart for the Bullhead and Fort St.John Groups comparing Stott (1978,1982), Duff and Gilchrist (1981) ar proposed nomenclature for this study.

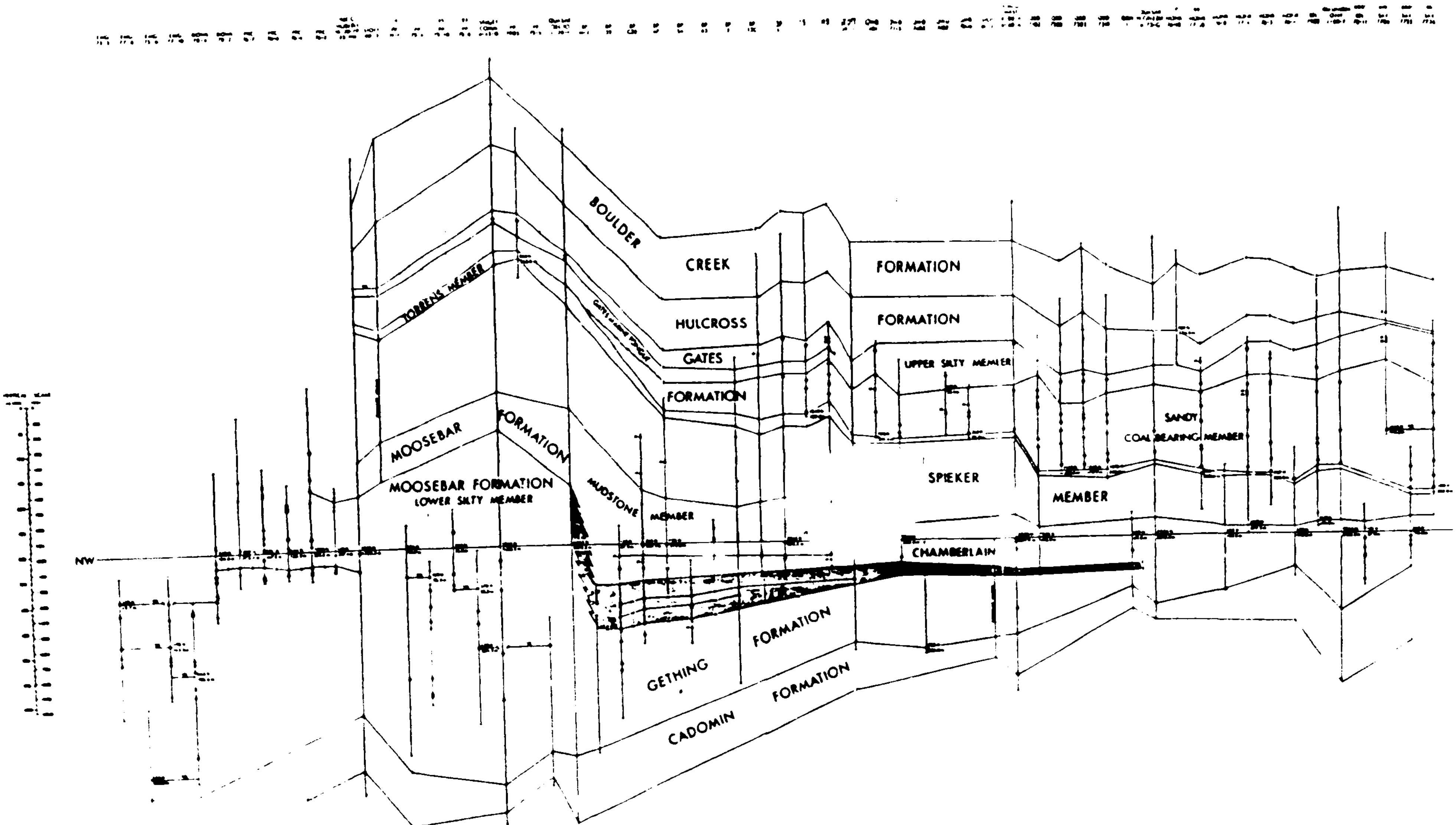

Figure 1.5 Correlation chart of Lower Cretaceous Formations by Duff and Gilchrist (1981). NOTE Correlation between Chamberlain unit with upper part of Gething Formation. Shaded area represents marine tongue.

by regressive pulses, in the more northern areas of the basin during Bluesky time. Jackson (1984) included the deltaic sequence, which is equivalent to the Chamberlain Member of Duff and Gilchrist (1981), in Bluesky Formation stratigraphy. A recent study by O'Connell (1988) suggests that the Bluesky sediments of northwestern Alberta were deposited as a multistoried series of wave-dominated shoreface deposits in a barrier island setting accompanied by tidally-dominated nearshore channels and interchannel brackish-bay sediments. Moslow and Pemberton (1988) provided a brief sedimentological and ichnological review of shoreface sandstones in the Bluesky Formation of northwestern Alberta

A number of general regional geologic studies of Lower Cretaceous strata have been produced by the Geological Survey of Canada. The Bullhead Group has been documented by Stott (1968, 1973) and the Fort St. John Group was discussed by Stott (1968, 1982). Comprehensive reviews of the development of Lower Cretaceous clastic sequences were given by Stott (1984) and Caldwell (1984). McLearn and Kindle (1950) provided one of the first comprehensive descriptions of the geology of the Foothills area in northeastern British Columbia. For a more comprehensive list of earlier studies the reader is referred to the bibliographies provided by Stott (1973, 1982).

1.4.2 PREVIOUS PALEOGEOGRAPHIC RECONSTRUCTIONS

A number of regional stratigraphic studies of the Bullhead and Fort St. John Groups have discussed the history of the initial transgressive phase of the Moosebar Boreal Sea during Aptian(?) - early Albian time. Paleogeographic maps for the Moosebar sea have been published by McLearn (1944), Badgely (1952), Rudkin (1964), Stott (1968, 1973, 1982, 1984) and Williams and Stelck (1975). Unfortunately, however, paleogeographic reconstructions specific to the Bluesky period, northeastern British Columbia, are scant. Recent brief summaries of Lower

Cretaceous strata by Jackson (1984) and Smith *et al.*, (1984) have provided basic reviews of Bluesky paleogeography. McLean and Wall (1981) and Taylor and Walker (1984) both provided some comments on the early depositional history of the Moosebar Sea in British Columbia. And finally, O'Connell (1988) has provided new insight into the sedimentology and paleogeography of the Bluesky Formation in northwestern Alberta. The following discussion is a brief summary of reconstructions provided by the above references.

As early as 1915, paleogeographical maps of the Western Interior Basin were drawn by Dowling. However, because little was known about the distribution of the vast marine successions, the maps were incomplete. McLearn (1944) provided the first series of paleogeographic maps for the Lower Cretaceous. He suggested that Gething Formation sediments were marginal to a marine embayment to the north, which ultimately flooded what is now the present site of the Foothills and Plains. This marine transgression was responsible for the deposition of the Moosebar (Wilrich) shales. Figure 1.6 illustrates McLearns' interpretation of the paleogeographic setting of the Gething Formation prior to and after the invasion of the Boreal Sea. This interpretation of a Gething alluvial plain bordering a marine embayment, in the Peace River Plains, was also supported by Stott (1968).

The paleogeography of the Western Interior was also discussed by Jeletzky (1971) and Williams and Stelck (1975). Jeletzky (1971) inferred that the Boreal sea in Aptian time was restricted to the Arctic region (Figure 1.7) and did not flood southwards until the late Early Albian. Stott (1968), though, considered that the sea had already flooded much of the northern areas during Aptian time and completely transgressed the basin by early Albian time. Williams and Stelck (1975) considered that the latest Early Albian *(Subarcthophites)* sea had by that time already extended southward into southern Alberta.

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Figure 1.6: Paleographic reconstruction by McLearn (1944) of the Western interior of Canada during A) Lower Blairmore time and B) Clearwater time. Dotted areas are continentally dominated deposits and dashed lines represent marine invasion.

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Figure 1.7: Paleogeographic reconstructions by Jeletzky (1971) inferring the maximum extent of the Clearwater Sea in A) Aptian time and B) late lower Albian.

Stott (1973) had refined earlier interpretations to suggest that the sediment catchment basin, in northern Alberta and northeastern British Columbia, was divided by a broad emergent island (Figure 1.8). Gething and Bluesky Formation sediments were deposited on the west of the island and McMurray Formation sediments were deposited on the opposite side. The boreal seaway bordered the Gething alluvial plain to the north. Subsequent subsidence had resulted in the complete flooding of the alluvial plain in Early Albian time. Bluesky strata were suggested to be a reworking of Gething Formation sediments by the advancing sea (Stott, 1973).

McLean and Wall (1981), in tracing the movements of the Moosebar Sea in Alberta, suggested that the initial transgression of the Boreal sea had only flooded the areas north of the Wolverine River and south of Grand Cache, bypassing the area in-between (Figure 1.9). They suggested that the bypassed area remained emergent producing a deltaic complex. Continued migration of the Moosebar sea along early Cretaceous drainage systems formed extensive estuaries in Alberta prior to the complete flooding of northern British Columbia and Alberta by the Moosebar Sea (Figure 1.9).

Jackson (1984) in a study of Mannville stratigraphy presented a series of paleogeographic maps which documented the early transgressive history of the Moosebar Sea. The events are portrayed in a series of diagrams starting in early Gething time to Bluesky time (Figure 1.10). According to Jackson, Bluesky strata were deposited subsequent to the flooding of the basin as a collage of offshore bars and barrier islands. Smith *et al.*, (1984) provided similar interpretations also for Bluesky deposition. Both Smiths' and Jacksons' interpretations are based largely on geophysical signatures for subsurface control rather than core descriptions.

O'Connell (1988) suggested that the Bluesky Formation in northwestern Alberta consisted of stacked barrier shoreface deposits and nearshore tidal channel and inter-channel deposits. An

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Figure 1.8: Paleogeographic map of the Bullhead Group in early Albi. Stott 1982. The axial highlands separated the basin into two troughs for Murray sedimentation.

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Figure 1.9: Paleogeographic sketches of the Alberta Basin for Lower Mannville time according to McLean and Wall (1981). A) Basal Mannville dominated by extensive drainage systems; B) Ostracod time is suggested to be continental.

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Figure 1.10: Paleogeography of the Lower Mannville for the A) Gething Formation and B) Bluesky Formation according to Jackson (1984). The Alberta Basin in Bluesky time consisted of a collage of marine sand bars and deltaic deposits.

underlying structural influence provided by the Peace River Arch graben has caused a local thickening of the Bluesky shoreface sandstones. According to O'Connell, a sea-level rise in Bluesky time caused the shoreline to retreat southwards, where a series of stacked barrier deposits accumulated at this site. Shoreline positions and sand body geometry as envisaged by O'Connell for Bluesky deposition are illustrated in Figure 1.11.

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Figure 1.11: Simplified paleogeographic sketch of the Bluesky sediments in northwestern Alberta according to O'Connell (1988).

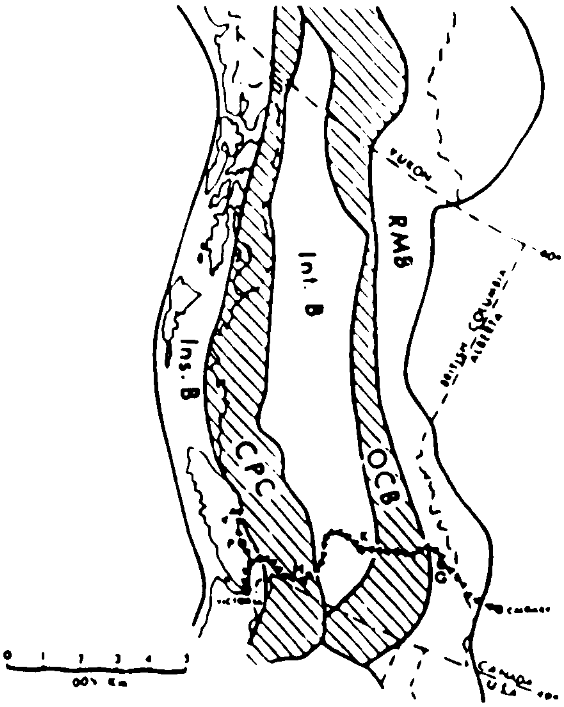
2.0 EVOLUTION OF THE WESTERN CANADIAN CORDILLERA

2.1 TECTONIC HISTORY - INTRODUCTON

The Canadian Cordillera is the result of an arc-continent collision orogen that evolved from the accretion of two allochthonous superterranes onto the North American craton edge in Mesozoic and Cenozoic time (Monger and Price, 1979; Templeman-Kluit, 1979). The result was the formation of five distinct geologic and physiographic belts that comprise the Cordillera (Figure 2.1). The amalgamation of exotic terranes onto the western margin of North America initiated the intense metamorphism, igneous activity and the deformation of para-authocthonous platform and foredeep deposits found in the present-day setting (Templeman-Kluit, 1979; Monger and Price, 1979; Price, 1981; Monger *et al.*, 1982). The following discussion is a synthesis of processes and products of the tectonic development of the Canadian Cordillera based largely on the cited references. It is intended to provide the reader with a basic understanding of the dynamic tectonism that occurred and its effect on shaping Cretaceous sedimentation patterns.

2.1.1 DEVELOPMENT OF THE CORDILLERA

The earliest stage in the development of the Cordillera was the deposition of a northeasterly tapering miogeoclinal sedimentary wedge on the leading edge of the North American craton. The miogeoclinal sedimentary package, of Late Proterozoic to Mid-Devonian age, consists of distinct linear belts of extensive shallow water carbonate banks passing progressively seaward to deeper water sand and shale deposits. These sediments are considered to have accumulated in a tectonically-active back-arc basin (Templeman-Kluit, 1979). Thin volcanic layers intercalated with the outer shale facies lends support to the theory of rifting accompanying miogeoclinal



RMB: Rocky Mountain Belt, a northeasterly tapering wedge of mid Proterozoic to Late Jurassic (1500-150 Ma) miogeoclinal and platformal carbonate and craton-derived clastic rocks, and overlying Late Jurassic to Paleogene exogeoclinal, cordillera-derived clastic rocks, horizontally compressed and displaced up to 200 km northeastwards on to the craton in Late Jurassic to Paleogene time. OCB: Omineca Crystalline Belt: mid Proterozoic to mid-Paleozoic miogeoclinal rock, Paleozoic and early Mesozoic volcanogenic and pelitic rock, local Precambrian crystalline basement highly deformed and variably metamorphosed up to high grades in mid Mesozoic to early Tertiary time, and intruded by Jurassic and Cretaceous plutons. Int. B: Intermontane Belt: late Paleozoic to mid-Mesozoic marine volcanic and sedimentary rock, mid Mesozoic to late Tertiary marine and non-marine sedimentary and volcanic rock; granitic intrusions comagmatic with the volcanics; deformed at various times from early Mesozoic to Neogene. CPC: Coast Plutonic Complex: Sedimentary and volcanic strata of known late Paleozoic to Tertiary ages and probable early Paleozoic and Precambrian ages, variably metamorphosed up to high grades, and dominant, mainly Cretaceous and Tertiary, granitic rocks. Ins. B: Insular Belt: Late Cambrian to Neogene volcanic and sedimentary strata, granitic rocks in part comagmatic with the volcanics; deformed at various times from Paleozoic to Neogene.

Figure 2.1: Physiographic belts of the Canadian Cordillera (modified from Price *et al.*, 1981)

sedimentation in a back-arc setting (Templeman-Kluit, 1979). A diagram atic sequence of events for the development of the Cordillera is provided in Figure 2.2.

Miogeoclinal sedimentation continued from Late Devonian until Middle Triassic time, accumulating thick carbonate sequences in the eastern half of the basin and shales and cherts in the western half. These are the sedimentary rocks which now comprise mo of the Rocky Mountain Fold and Thrust belt.

In Early Mississippian time, a second episode of rifting and seafloor spreading created the ancient Anvil Ocean which was bordered on the western margin by an the kinian exotic terrane (Templeman-Kluit, 1979) (Figure 2.2). By late Triassic to early Jurassic time spreading had reversed, and the Anvil Ocean lithosphere began subducting beneath Stikinia. Eventually, this subduction resulted in the first island-arc/continent collision and the amalgamation of several exotic of terranes, including Stikinia, onto the North American continent (Templeman-Kluit, 1979). Recognition of Tethyan Permian fusulinids has provided, in part, evidence for the allochthonous nature of these terranes (Monger and Price, 1979). The collision of this 'superterrane' and the North America continent resulted in substantial uplift and compression of miogeoclinal strata within the Omineca Crystalline Belt (Monger *et al.*, 1982). This collision orogen, referred to as the Columbian Orogeny, lasted well into Mid-Cretaceous time, and was responsible for the formation of an Andean-type foreland basin east of the mountains.

This newly created foreland basin was the new site for the deposition of clastic wedge sediments derived from the erosion of the recently uplifted Hinterland (Stott, 1984). The Bluesky Formation lies within part of these stacked clastic wedge packages.

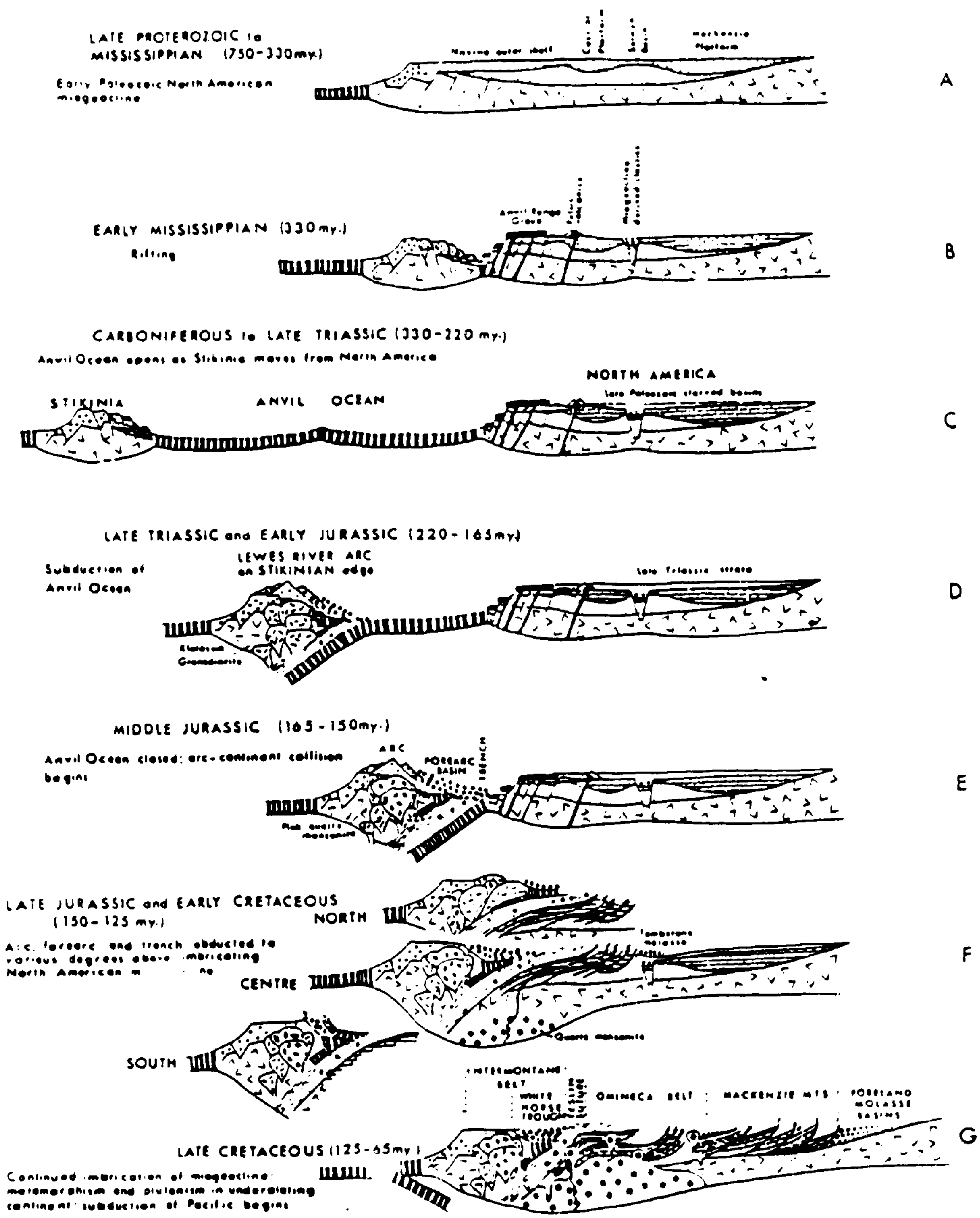


Figure 2.2: Stages of development for the Canadian Cordillera from Late Proterozoic to Late Cretaceous. (modified from Templeman-Kluit, 1979)

Beaumont (1981) indicated that the trough of the foreland basin was formed by loading of successive thrust plates onto an already thinned continental crust. This basin, which was inundated by epeiric seas in the Cretaceous, is rimmed by nearshore sequences along its western margin indicating that sedimentation had kept pace with loading and subsidence of the crust (Kauffman, 1984). The progressive eastward displacement of the thrust plates in the structural belt caused a likewise eastward migration of the depo-axis of the foreland basin.

Renewed subduction at the new continental edge led to the accretion of another exotic superterrane onto the continental margin during late Cretaceous and early Tertiary time (Templeton-Kluit, 1979). The collision of this superterrane, composed of the Alexander and Wrangellia terranes, led to the formation of the Coast Plutonic Complex and the eastward thrusting and folding of miogeoclinal and molasse deposits (Price, 1981).

2.1.2 TECTONIC ELEMENTS

The sedimentologic patterns of the foredeep in early Cretaceous time were not only influenced by the advancing structural front but, also by two subtle structural elements found within the study area. Bluesky deposition was, thereby in part, influenced also by these structures. These structural features are the known as the 'foredeep hingezone' and the 'Peace River Arch' (Figure 2.3).

The east-central portion of the basin, in northeastern British Columbia, contains a hinge zone which acted as an axial highland or margin to the basin. The axial high is marked by thin and rapid facies changes along its flank and sites of non-depositon (disconformities) on its emergent low-lying land mass. The eastern limit of the Barremian Cadomin Formation (Stott, 1984) lies juxtaposed to the axial high at the site of the ancient Fox Creek Escarpment (Smith *et al.*, 1984).

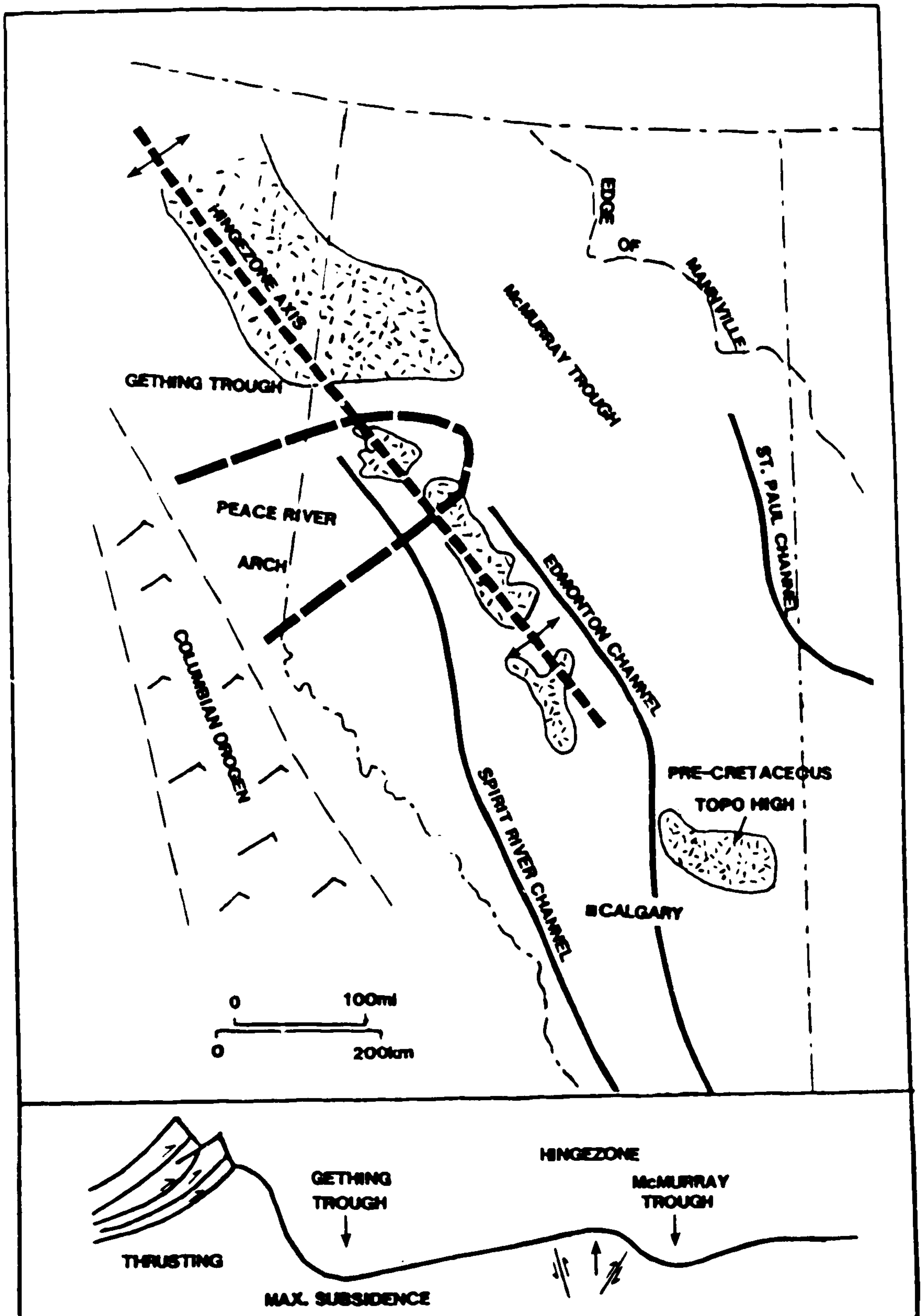


Figure 2.3: Tectonic elements controlling Cretaceous deposition, including the Peace River Arch and the hingeline separating the McMurray and Gething troughs (modified from Stott, 1982)

During the Lower Cretaceous, the northwest-trending Fox Creek Escarpment and the northwest flowing Spirit River Channel paralleled the hinge zone reflecting the structural grain of the hinge (Jackson, 1984). Any incursions of the boreal sea were, thus, likely separated into two arms by the hinge zone. The hinge zone consists of bevelled Mississippian and Triassic strata which probably contributed a large volume of detritus to the Bluesky Formation and its bounding strata.

Variable amounts of subsidence for different areas of the basin has been suggested to be related to the hinge zone. The role of the hinge zone is suggested to have acted as a rigid pivot point for the subsiding crust (Kauffman, 1984). Subsidence west of the axial highland, towards the deepest part of the basin, would have been the highest; moderate along the flanks of the hinge; and low along the stable platform east of the hinge (Kauffman, 1984). Although this may in part be true, it would seem more plausible that the higher rate of subsidence experienced in the western portion of the basin is due to the proximity of the loading thrust plates and the differential stresses accompanying the load. The hinge zone itself is probably a product of structural readjustment in response to the downwarp of tectonically-thinned crust.

The Peace River Arch is a second tectonic element which influenced sedimentation patterns in northeastern British Columbia. The Peace River Arch is a Precambrian basement structural feature that was a major positive element during the Paleozoic and a negative or subsiding element during the majority of the Mesozoic (Stott, 1982). The collapse of the Peace River Arch is considered to have begun in Mississippian time (Stott, 1982). Depositional trends in Aptian and Albian deposits seemed to have been influenced by the subsiding nature of the the Peace River Arch whereby the most frequent and thickest occurrences of paralic sediments seem to center on and parallel the axis of the ancient Peace River Arch (Stott, 1982; 1984). By Late Albian, the Peace River Arch apparently exhibited little control on sedimentation patterns in the foredeep (Stott, 1982).

2.1.3 SUMMARY

The origin and structural history of the foredeep and the contemporaneous rise and fall of sea level in the basin was primarily controlled by: (1) the activity, timing and intensity along the Proto-Pacific coast with the accretion of exotic terranes; (2) the timing and amount of compressional tectonics (thrust plates) in the eastern portion of the Cordillera; and (3) the formation and structural integrity of minor structural elements in the accentuated crust. The combination of these tectonic elements ultimately contributed to shaping the foreland basin. The history of epi-cratonic marine transgressive and regressive cycles in relation to these tectonic parameters are reviewed in the following discussion.

2.2 REGIONAL GEOLOGIC SETTING

The objective of this discussion is to provide a synthesis of Cretaceous strata in relation to the transgressive-regressive cycles which dominate its history. The importance of this becomes apparent when considering what role tectonics had, if any, in causing these cyclic events in the foredeep. More precisely, both the Bluesky/Moosebar series, a major regional flooding event, and the Chamberlain, a regressive-progradational event, may owe their origins to periods of structural activity rather than just simply basin subsidence or changes in eustatic sea level within the seaway.

The physiography of the seaway varied between a low-relief coastal plain and a shallow epicratonic shelf (Stott,1982). Sediments deposited into this seaway, including those which comprise the Bluesky Formation, were greatly modified by frequent changes in sea level of the boreal sea. As a result of these sea-level fluctuations, the Cretaceous section varies from alluvial to deep-water shale deposits.

The history of tectonic events in the Cordillera and the development of the foredeep are well recorded in the Cretaceous succession of the Rocky Mountain Foothills. Three clastic wedge sequences (Stott, 1975) contain alternating sediments of marine, floodplain, deltaic and alluvial environments. Each clastic wedge is comprised of several large cyclothems in which a number of major transgressive - regressive cycles and minor subcycles are found (Stott, 1984). The Gething, Bluesky, and Moosebar Formations form part of the second clastic wedge.

The oldest and lowermost clastic wedge contains two large-scale cycles. The first cycle contains Jurassic Upper Fernie shales followed by the early Cretaceous massive Monteith sandstones, reflecting basin-filling after marine transgression and flooding (Stott, 1975; 1984). The second event called the Beattie Peaks cycle is of Cretaceous Valanginian age and is composed of the remaining Minnes Group strata (i.e. Beattie Peaks, Monach and Bickford Formations).

The second clastic wedge, of late Neocomian to early Cenomanian age, includes six transgressive - regressive cycles beginning with the initial Boreal sea inundation and deposition of the Bullhead Group. Tangible evidence for this initial transgression is still lacking. The Bullhead cycle was followed by the early- to mid-Albian *(Arcthoplites)* Moosebar/Gates cycles (Stott, 1984) which flooded most of the interior basin. The Moosebar Formation deposits are equivalent to the Wilrich/Falher nomenclature. Four subsequent cycles developed after the mid-Albian including: the late mid-Albian *(Gastroplites)* Hulcross transgression (Stott, 1984) which is equivalent to the Harmon Formation; the late Albian Joli Fou cycle (equivalent to the Kiowa-Skull Creek cyclothem of Kauffman *et al.*, 1977); a regressive Hasler - Lepine cycle of late Albian *(Neogastroplites)* age, presented by the Goodrich-Sikanni sandstones (Stott, 1984); and the final cycle of latest Albian - earliest Cenomanian age is the Cruiser - Sully package and Dunvegan delta. These latter sediments are equivalent to the Shaftesbury and Dunvegan Formations of northern Alberta. By late Albian time, the connection between the Boreal Sea and the Gulf embayment was completed (Stott, 1984).

The third clastic wedge includes beds of latest Cenomanian to early Maastrichtian time. The first cycle is termed the Blackstone, which culminated with the deposition of the Cardium Formation (Stott, 1975; 1984). The second marine phase is recorded in the Muskiki, Bad Heart and Puskwaskau Formations. The Santonian - early Campanian regression and transgression cycle is marked by the Chungo Sandstone of the Smoky Group (Stott, 1984). The Nomad cycle of early to mid-Campanian age is represented by the Belly River Formation. The last marine incursion, occurring in late Campanian to early Maastrichtian time, is represented by the Wapiti Formation

The deposition and nature of sediment within these cyclothems are often related to eustatic sea level rise and fall (Kauffman, 1984; Haq *et al.*, 1987). A change in sea level worldwide and the respective increase or decrease in absolute water volume are considered to be related to either periods of large scale continental glaciation or a product of plate tectonics.

Depositional sequences and tract boundaries are belived to be globally synchronous (Haq *et al.*, 1987). As such, a globally uniform sea level curve was erected to correlate transgressive-regressive packages to sea-level fluctuations (Haq *et al.*, 1987). However, criticism of this method has centered on the fact that some depositional events are not recorded on a worldwide basis, and the lack of adequate incorporation of subsidence into the scheme (Caldwell, 1984; Kauffman, 1984). Bergman and Walker (1988), in a discussion of Cardium stratigraphy in the Carrot Creek Field of Alberta, have emphasized that recorded, frequent sequence boundaries were tectonically-controlled rather than the result of global-eustatic fluctuations. Embry (1988) also considers that long term cycles of eustatic sea level changes cannot explain the thin bed nature of some sediment packages and that some sequence boundaries can be correlated to episodes of tectonism. Caldwell (1984) also suggested that the short duration of some cycles and the limited geographical distribution of some units cannot be explained by using eustasy theory alone. Caldwell (1984) designated these short-term depositional cycles, which may have been tectonically-controlled, as "transgressive-regressive couplets".

It would seem more plausible to relate the coarse clastic tongues as products of tectonism in the Cordilleran front (Jeletzky, 1977; Stott, 1984). Stott related the two lowermost clastic wedge assemblages to the two phases of the Columbia Orogeny and the third and last clastic wedge sequence to the initial phase of the Laramide Orogeny. The episodic nature of deformation is indicated by the presence of localized regressive pulses or sub-cycles within each major cycle (Caldwell, 1984; Stott, 1984). Thus, it is quite possible that eustatic sea level changes originating from global tectonics may have been overwhelmed by changes in regional tectonics.

The importance of this discussion, although not to discount the role of eustasy, is that Canadian Cordilleran epierogenic events provide a means of explaining variations in local and regional sedimentologic and stratigraphic patterns. As stated above in the preceeding discussion, the foreland basin (Beaumont, 1981) is a product of downward flexure of the ancient elastic lithospheric margin in response to passive loading of supracrustal rocks during the evolution of the thrust and fold belt (Price, 1981). With continued younger thrusting events, three products would have resulted in the foredeep: (1) reworking of previously buried molasse deposits (Price, 1973), (2) eastward migration of the troughal axis (Stott, 1982, 1984) (Figure 2.4), and (3) initiation of transgressive events (Kauffman, 1984).

Of the three, the initiation of transgressive events is of greatest importance to this study. It is postulated that major sea level rise or transgressive events and extensive basin subsidence result with regional tectonic activity. Tectonic quiescence, on the other hand, may allow progra-

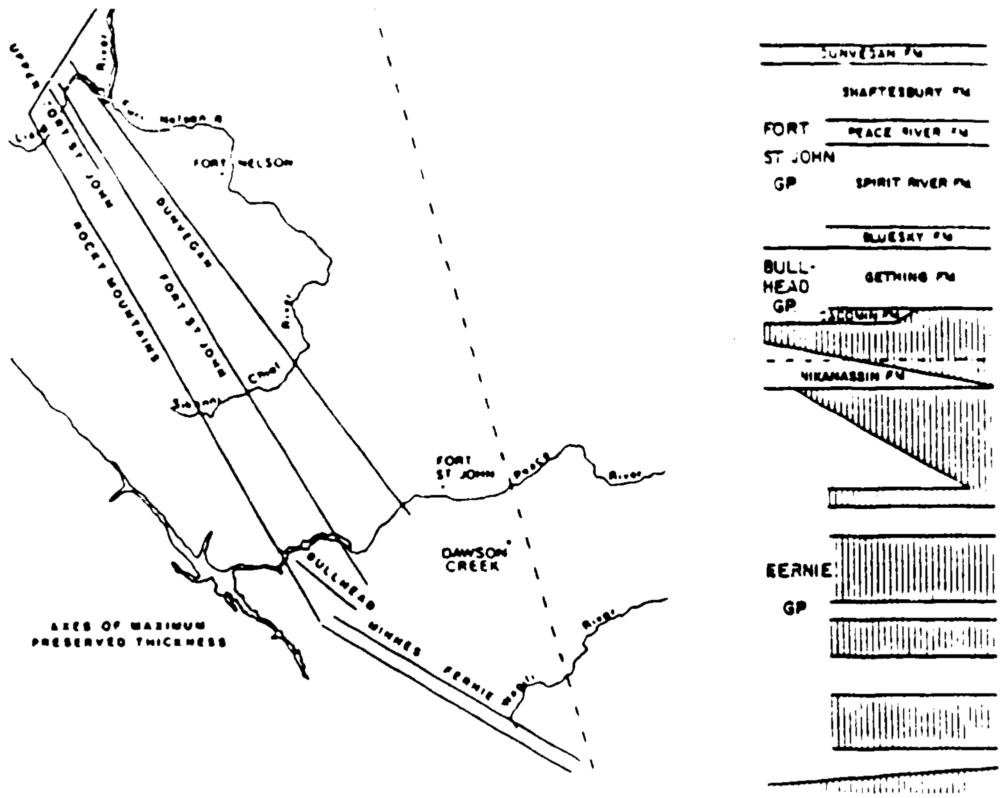


Figure 2.4: Axial trends of maximum preserved thicknesses of successive formations demonstrating the migration of foredeeps with tectonism. The oldest succession is the Fernie-Minnes, succeeded by the Bullhead, then the Fort St.John, and finally by Dunvegan sediments. (modified from Stott, 1982).

dation of eastward thinning clastic sequences and retreating sea levels. Subsidence caused in this stage may only occur by sediment loading (Caldwell, 1984). Therefore, clastic sediments deposited in the foreland basin will contain alternating transgressive-regressive cycles reflecting the tectonic events.

The history of the Bluesky - Chamberlain - Moosebar series may reflect this pattern quite well. The initiation of marine flooding and deposition of the Bluesky Formation may have been associated with a brief pulse of tectonic activity; subsequent structural quiescence then allowed the progradation of the Chamberlain delta. With a renewed period of tectonism and loading, the basin responded by subsiding, causing a rapid inundation of the Moosebar Sea. The latter is represented by the deep water Moosebar shales.

The implications of these theories become apparent when considering resource modelling. Primary targets of rapidly migrating strandlines, stratigraphic pinch-outs, and onlap of shoreline sands onto axial highs are common settings in this model. When in combination with an extensive transgression and deposition in anoxic conditions thick sequences of source rocks are produced. Thus, the potential for hydrocarbon emplacement is greatly enhanced.

2.3 REGIONAL STRATIGRAPHY

2.3.1 INTRODUCTION

The stratigraphic relationship of the Bluesky Formation to its bounding sediments is a complex problem on which the following discussion will focus. The internal stratigraphy and of the Bluesky Formation is presented in Chapter 4.

Since this study is an integration of outcrop and borehole data from the Foothills in conjunction with subsurface data from the Peace River Plains, a revised stratigraphic nomenclature scheme is incorporated to facilitate discussion (Figure 1.4). The Bluesky Formation will include the basal sandstones previously included in the Wilrich Member of the Spirit River Formation in the Peace River Plains area and the succession referred to as the 'Gething Marine Tongue' by Duff and Gilchrist (1981) in the Foothills belt. The Moosebar Formation, following Leckie (1983), will include subsurface strata previously referred to as the Wilrich Member of the Spirit River Formation and also the Buckinghorse Formation. The Chamberlain Member of the Gething Formation (Duff and Gilchrist, 1981) will be informally revised to the Chamberlain Formation. The latter revision is incorporated so as to avoid duplication of the term 'Gething Formation' when defining the stratigraphic sequences. This stratigraphic division is applicable to both outcrop and subsurface data.

The Bluesky Formation is bounded by the continental deposits of the Gething Formation below and the deep-water shales of the Moosebar Formation above. However, in the southern half of the study area between the Sukunka area to the north and the Wapiti Lakes area to the south, the Bluesky Formation is overlain by the coal-bearing Chamberlain Formation. The Chamberlain Formation is then, in turn, overlain by the Moosebar Formation. The stratigraphic relationships of these Lower Cretaceous rocks in the Foothills area have been extensively reviewed by Stott (1968, 1973, 1982), Hughes (1964, 1967), Leckie (1983) and McLean (1982). Subsurface stratigraphy was discussed by the Alberta Study Group (1954), Pugh (1960), Rudkin (1964), and Jackson (1984). Excellent summaries of macro- and micro-faunal zonations for the Cretaceous system were published by Jeletzky (1968) and Caldwell *et al.*, (1978) respectively.

A north-south oriented regional cross-section (Figure 2.5) has been constructed using the base of the Shaftesbury shales as the datum. This cross-section illustrates the extreme thickening of the section from the Cadomin to the Cadotte Formations towards the south, within the foothills, compared to a thinner isopach in the north. Tr thickening trend is considered to be related to a greater amount of basinal subsidence and infilling in the south or, alternatively, to a number of extensive unconformities which bevelled these units. The cross-section also demonstrates the log correlation of the Bluesky Formation across the basin. If one assumes that the pre-Bluesky surface was reasonably flat, then the coarsening upward interval in the south half of the section is stratigraphically equivalent to the Bluesky sediments described by Pugh (1960) in the Peace River Plains.

The sediments of the Bullhead Group and Fort St. John Group represent an overall change in chararcter as the sediments become increasingly more marine from the basal alluvial plain deposits of the Cadomin Formation through the shallow marine sediments of the Bluesky Formation to the deep-water shales of the Moosebar Formation The following discussion is meant to introduce the stratigraphic relationships of the Gething - Bluesky - Chamberlain - Moosebar sequences and to present a general depositional setting prior to-, during-, and after-deposition of the Bluesky Formation.

2.3.2 GETHING FORMATION

The Gething Formation is comprised of interbedded, carbonaceous sandstones, siltstones, mudstones and coals which were deposited in an alluvial-deltaic plain environment during the Early Cretaceous (Stott, 1968). Clastic material was deposited under subaerial to ... aqueous conditions on a low lying, poorly drained plai hich was bordered on the north by the Boreal Sea. Lower sections in the Gething Formation are generally coarser grained and are recognized as upper delta-plain deposits, whereas finer grained sediments in the upper beds of the formation are accumulates of lower delta-plain settings (Stott, 1973). The Gething Formation, on a regional scale, is a composite of several coalescing deltas rather than one large delta (Stott, 1973)

As mentioned previously, the axial high region (hinge zone), in the east-central part of the foredeep, separated the basin into two troughs; the western half was filled with Gething sediments, whereas the eastern half was filled with McMurray sediments (Stott, 1973). The pattern of sedimentation in the southern reaches of the Gething trough was influencod by active subsidence on the Peace River Arch (Stott, 1975). The thickest accumulations of Gething Formation are found adjacent to the north flank of the arch (Stelck, 1975)

The Gething Form[illegible]n flora is comprised mainly of 'Lower Blairmore' nondicotyledorous species of ferns, gin[illegible]es, cycadophytes, and coniferophytes (Bell, 1956). Microflora analysis by Singh (1964) for the [illegible]ville Group, which is in part stratigraphically equivalent to the Gething Formation, indicates that a warm, temperate to sub-tropical, humid climate prevailed in the central region of Alberta. This same type of climate can be postulated for Gething and Bluesky sediments since they were deposited at similar paleolatitudes as Mannville strata.

The age for Gething Formation sediments is still controversial. Traditionally, the Gething Formation has been regarded as Barremian to Aptian based on floral remains (Bell, 1956). However, lowermost beds of the Garbutt Formation in the Liard district, which are equivalent to Gething strata, have yielded the ammonite *Pachygrycia* suggesting an age of Early Albian (Jeletzky and Stelck, 1981). The reader is referred to Stott (1982) and Caldwell (1984) for excellent detailed discussions of Gething dates. The importance of determining an accurate age for Gething sedimentation is necessary for establishing a maximum age for Bluesky deposition.

2.3.3 BLUESKY FORMATION

The initial strong southern advance of the Boreal Sea, which flooded the low-lying Gething delta plain, is recorded by marine sandstones and mudstones of the Bluesky Formation. The Bluesky Formation lies directly on top of the Gething Formation throughout northeastern British Columbia, except in the immediate area of the axial high zone where it unconformably overlies the subcropping Triassic strata. The basal contact consists of a thin polymictic conglomerate band with a sharp, often bioturbated base, overtop of either coals or fluvial deposits of the Gething Formation. The eroded base of the Bluesky Formation may indicate either a beveling effect as a result of Bluesky deposition, or a short hiatus at the end of Gething. The upper contact with the Moosebar Formation is generally conformable and abrupt. The upper contact with the Chamberlain Formation, in the Sukunka-Wapiti River areas, is placed at the first occurrence of carbonaceous shale or sand or a coal seam. For purposes of this study, it is herein considered that the definition of Bluesky strata should be limited to include only marine or marginally-marine sediments, deposited directly as a result of the Early Albian transgression. Chamberlain Formation sediments, on the other hand, are separated from the Bluesky Formation because of their continental origin. In outcrop, the Bluesky occurs as a recessive, talus covered interval between the prominent cliff-forming units of the Gething and Chamberlain Formations.

The Bluesky Formation in northeastern British Columbia is considered to be laterally equivalent to the Bluesky Formation of the Alberta Peace River region, the Glauconitic Member of central Alberta, and the Wabiskaw Member of the Mannville Group in northeast Alberta.

A network of regional gamma-ray log correlations (Figures 2.6-2.8) illustrate the local thickening and thinning of the Bluesky Formation as well as its stratigraphic relationships. Bluesky strata in the Peace River Plains area typically consist of generally coarsening-upward, isolated sand buildups. These buildups can attain a thickness greater than 30 m. Where the buildups are absent, the Bluesky Formation is represented only by a thin conglomeratic veneer. An isopach map of the thickness of the BlueskyFormation in the Peace River Plains area illustrates the localized sandstone buildups (Figure 2.9). Bluesky strata can be traced southward as far as the Wapiti Lake area, which possibly represents the southernmost limit of the transgressing Moosebar Sea in northeastern British Columbia. Taylor and Walker (1984) suggested that the Moosebar Sea eventually advanced as far south as the Ram River area in Alberta. In the Monkman to Wapiti Lake area, Bluesky strata are much thinner than in the Sukunka region, but still retains a consistent lithological character. South of this region it is believed that Bluesky strata are absent and that Chamberlain strata cannot be differentiated from Gething strata. In the Foothills area, the Bluesky Formation consists of a distinctive coarsening-upward, interbedded sandstone and shale succession (Figure 2.6). The coarsening-upward profile is reflected not only by the decrease in mud content (sandier-upwards), but also by increasing grain size and sand bed thickness.

The Bluesky Formation can be dated, approximately, on the basis of its stratigraphic position and lateral associations. Unfortunately the Bluesky Formation lacks diagnostic index fossils which might have been used to date the sequence. Caldwell (1984), however, assigned the Bluesky Formation of the Alberta Peace River Plains to the *Cleoniceras subbeyleyi* ammonite zone, which corresponds to the early Albian *Rectobolivina* sp. subzone of the *Gaudryina nanushukensis* microfaunal zone (Figure 2.10). The *Gaudryina nanushukensis* foraminiferal zone was established by Caldwell *et al.*, (1978). *Cleoniceras* is of early Albian age (Stelck *et al.*, 1956) *Rectobolivina* sp. was later reassigned to early Early Albian age by Caldwell and North (1984) Stelck *et al.* (1956) reported *Cleoniceras* in the basal part of the Loon River shale (equivalent to Moosebar shales) in the Alberta Peace River area. Wall (in Caldwell *et al.*, 1978, p. 504) recorded the presence of *Rectobolivina* sp. within the same horizon. Stelck *et al.*, (1956) also indicated that the Wabiskaw Member belonged to the *Trochammina mcmurrayensis* zone (later revised to a sub-

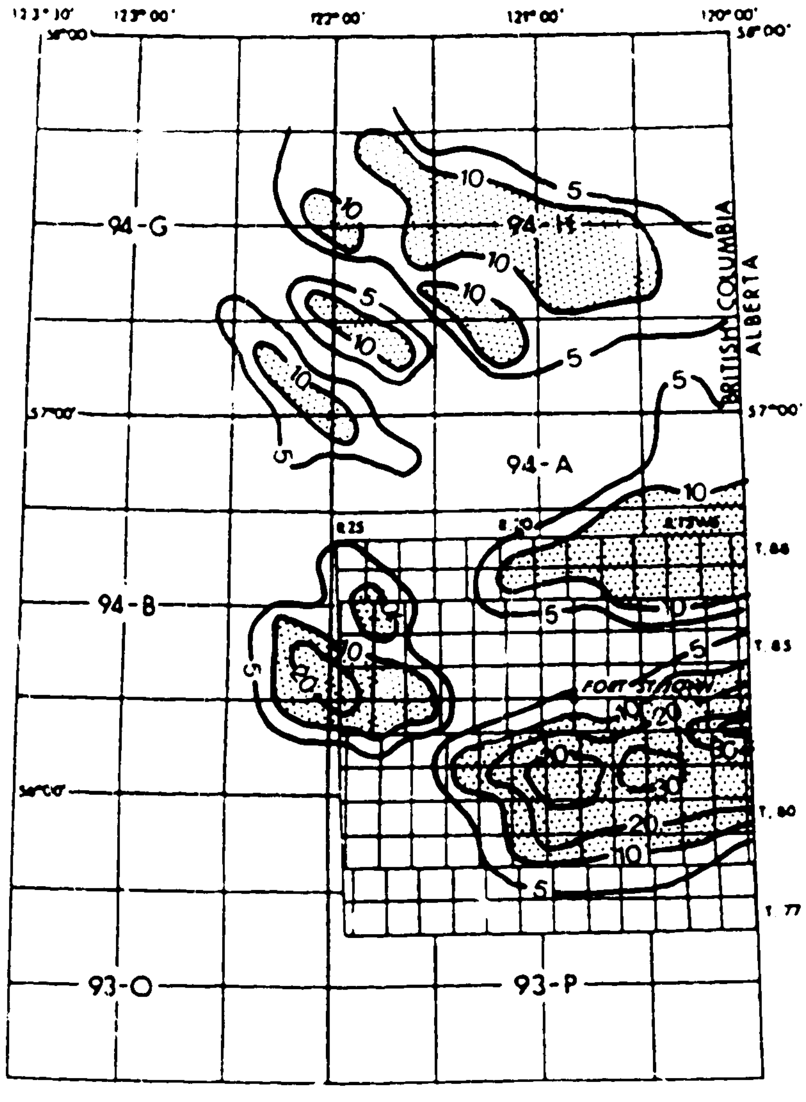


Figure 2.9: Isopach map of the Bluesky Formation in the Peace River Plains subsurface.

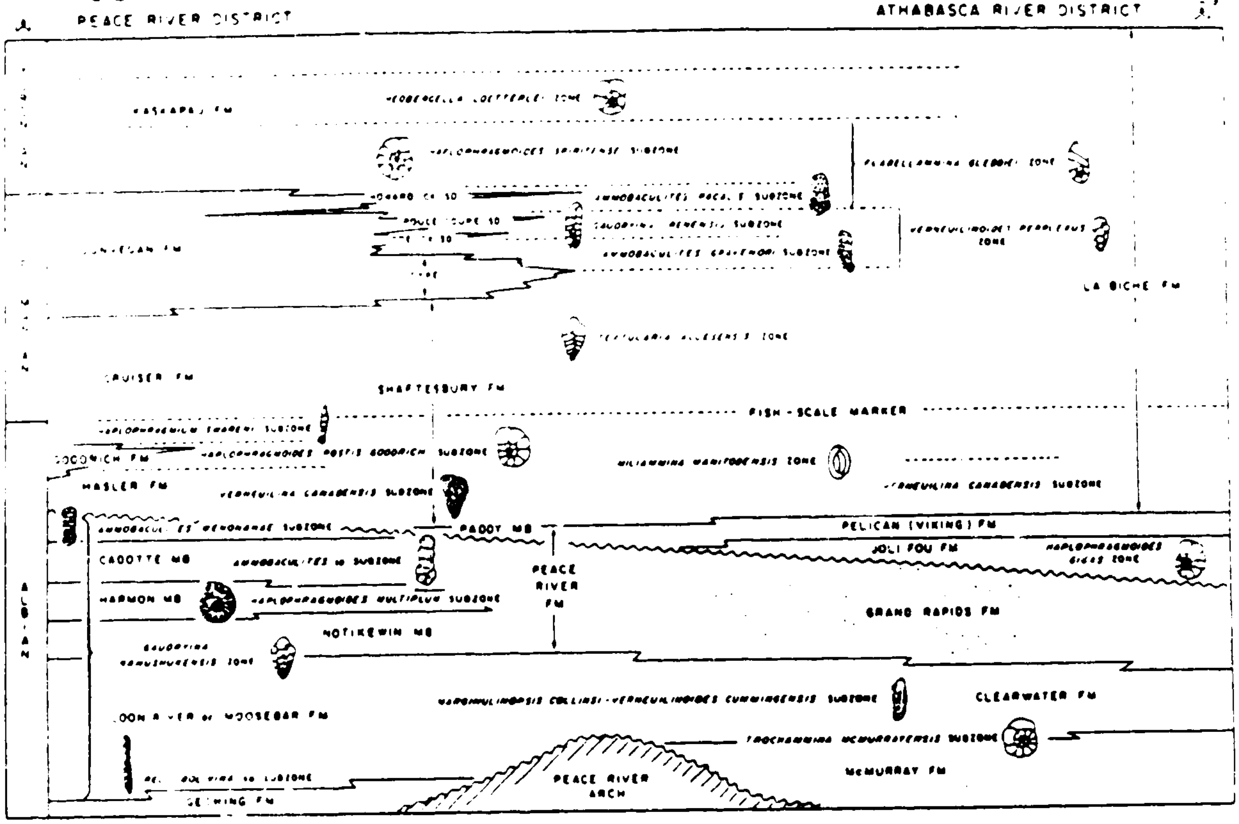
zone by Caldwell *et al.*, 1978). Furthermore, Mellon and Wall (1956, p.11) were able to trace this same microfaunal assemblage into basal shales of the Moosebar Formation in the Pine River area of northeast British Columbia. Thus, the Bluesky Formation is probably no younger than the early Early Albian *Rectobolivina* sp. subzone with an upper limit of lower *T. mcmurrayensis*, since this same assemblage is found in the overlying Moosebar shales.

An implication of this relationship is that the upper parts of the McMurray Formation and the Wabiskaw were deposited later than Bluesky deposition in the Peace River area. This would imply the transgression of the Boreal Sea was moderately diachronous. As indicated previously, the initial transgression of the Moosebar Sea, south of the Peace River, extended into central Alberta, forming extensive estuaries along entrenched drainage systems (McLean and Wall, 1981). The deposits are represented by the Ostracod Zone and upper parts of the McMurrary Formation. This would imply that Bluesky strata of northern British Columbia and Alberta are time equivalent, at least in part, to the Ostracod Member of the Lower Mannville in central Alberta. Thus, the Glauconite Sands, which overlie the Ostracod Member, would be deposited later than the earlier phase of Bluesky deposition.

2.3.4 *CHAMBERLAIN FORMATION*

The Chamberlain Formation, which conformably underlies Moosebar marine shales, typically consists of interbedded sandstones, siltstones and mudstones with two occurrences of commercially viable coal seams. The Chamberlain Formation was deposited as a prograding delta-plain complex at the edge of the Boreal Sea. The Chamberlain sequence is defined by the Chamberlain coal seam (local nomenclature) at the base and a distinctive glauconitic sandstone at the top (Figure 2.11).

A.



B.

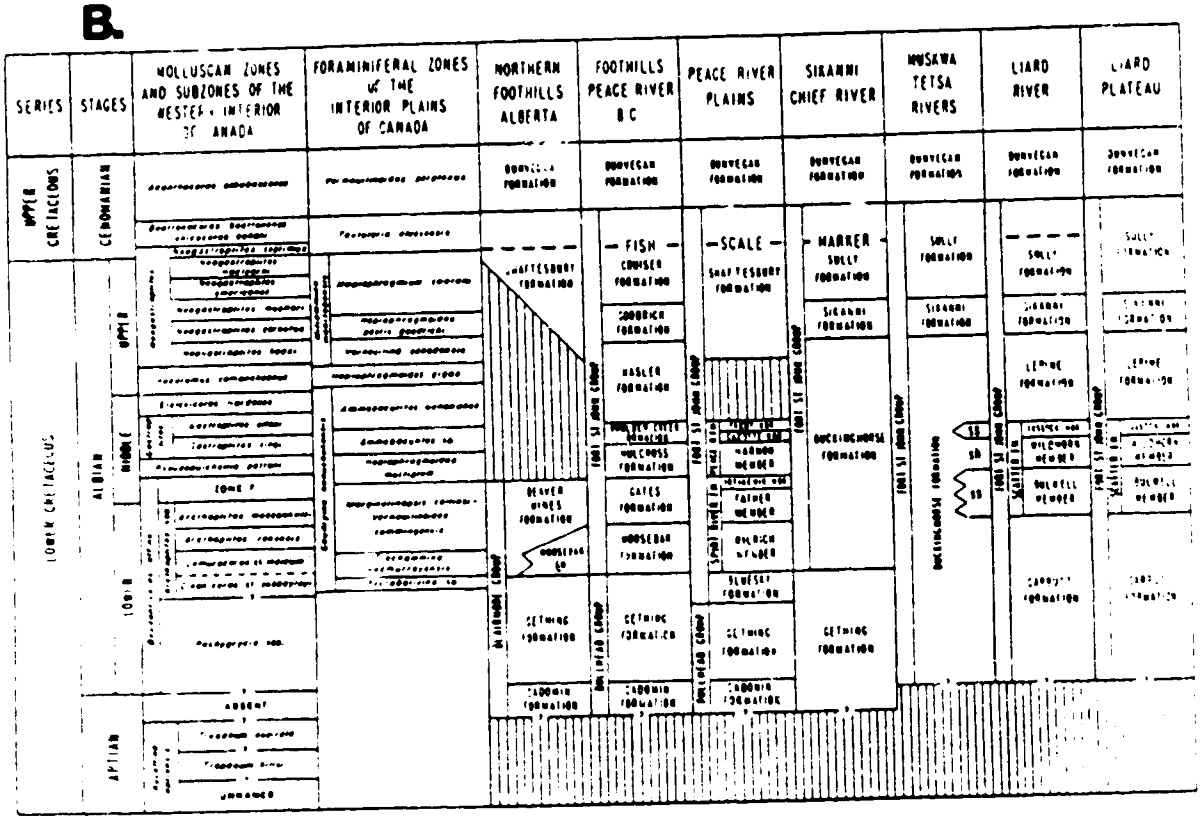


Figure 2.10: Biostratigraphic correlation charts of the A) foraminiferal zones (modified from

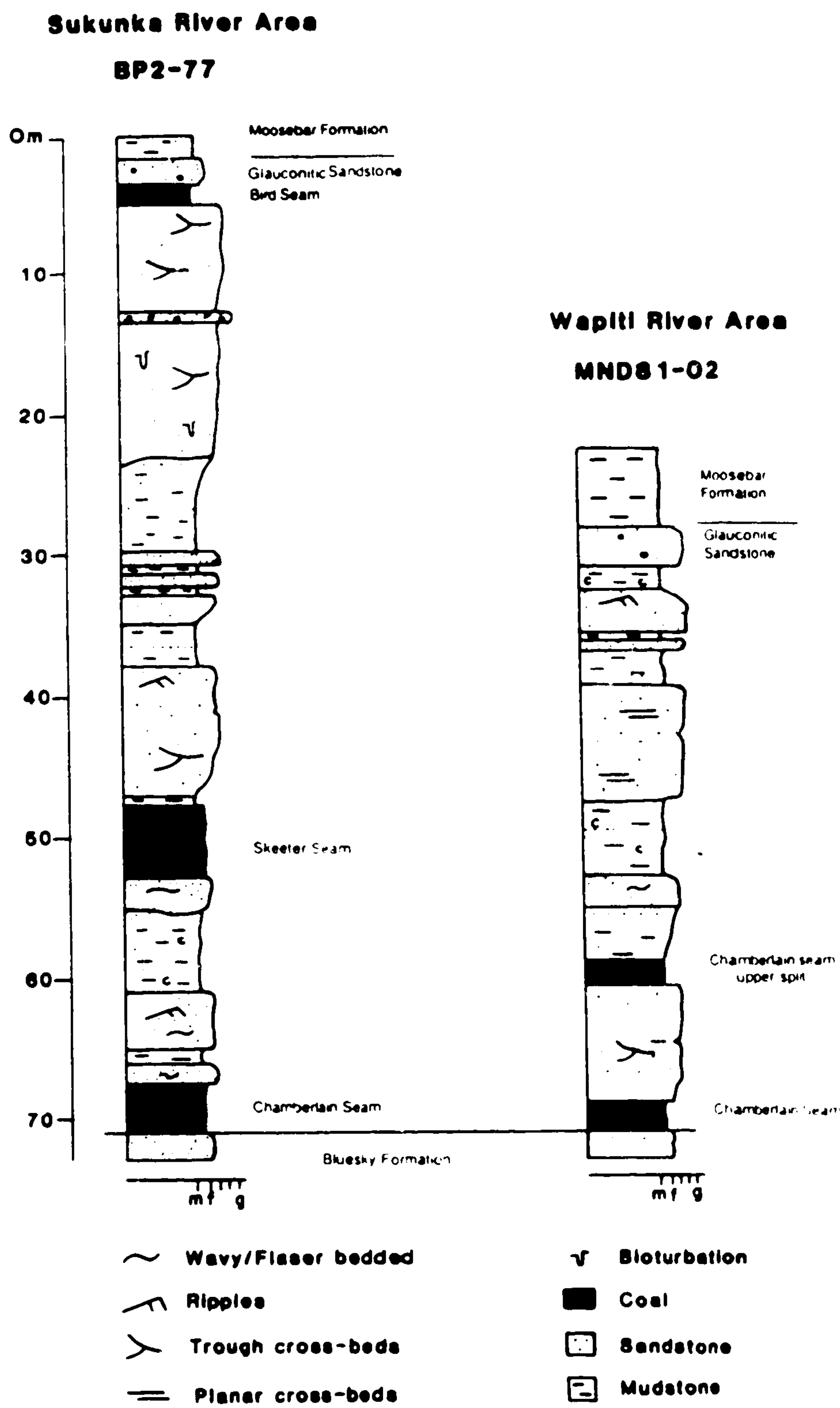


Figure 2.11: Typical lithologic successions for the Chamberlain Formation in the Sukunka and Wapiti Rivers area in the Foothills.

The Chamberlain Formation extends southward from the Sukunka River area into the Belcourt-Wapiti River region and eastwards through the study area into Alberta. Thrust faulting, at the western margin of the inner foothills, has tectonically uplifted and subjected to erosion all Lower Cretaceous sediments. The Chamberlain Formation attains a maximum thickness of 80 m in the Sukunka-Bullmoose area, tapering southward to 30-50 m in the Wolverine River area. In the Monkman area, the Chamberlain Formation is 48 m and at Nekik Mountain, a borehole intersected 35 m of Chamberlain strata. The Chamberlain unit is not recognized north of the Sukunka River, indicating a nearby shoreline. The Chamberlain section is considered to be, in part, equivalent to the highest sandstones of the Bluesky Formation, and also laterally equivalent to lower-most Moosebar shales of the Peace River Plains. The exact boundary of the northern limit of Chamberlain strata cannot be traced due to very few wells in the area. Upper parts of the Gladstone Form n of the Alberta Foothills may prove to be 'erally equivalent to the Chamberlain sequence.

Flora of the Chamberlain Formation also belong to the 'Lower Blairmore' type. Although no analysis has been performed on Chamberlain strata, it is known that angiosperms did not become well established until mid-Cretaceous in the upper Commotion Group (Singh, 1975).

An age for the Chamberlain Formation is suggested to be Early Albian. This is based mainly on stratigraphic position and the dating of lower Moosebar and Bluesky strata. As mentioned in the above discussion, Bluesky strata probably belong to the *Rectobolivina* sp. subzone of early Early Albian age. Moosebar shales have yielded fauna of early Albian age (Stott, 1982). Thus, if Chamberlain strata were the fi of deposition before the complete flooding by the Moosebar Sea, an age of Early A ould not seem unreasonable.

2.3.5 *MOOSEBAR FORMATION*

Sediments of the Moosebar Formation in northeastern British Columbia have been described as offshore marine shales (Stott, 1982) at least 300 m thick in the Peace River area, which thin southwards to 43 m at Mount Torrens. The Moosebar shales were deposited within the neritic zone of the transgressing Boreal Sea. The shales are dark grey, containing foraminifera, abundant, small, horizontal, pyritized burrows and bands of sideritic concretions. Numerous thin beds of bentonites have also been described in the Moosebar shales (Kilby, 1985).

Contemporaneous flooding by the Boreal Sea is represented by the Clearwater Formation in northeast Alberta, the Loon River Formation in north-central Alberta, and the Cummings Member of the Mannville Group in the McMurray area.

The deep water shales of the Moosebar Formation fall within the *Arcthoplites* sp. zone (Ammonitida), considered to be Early Albian in age (Stott, 1968). Micro-faunal assemblages, including species of the *T. mcmurrayensis* (Mellon and Wall, 1956) and *Marginulinopsis collinsi-Verneuilinoides cummingensis* subzones, also serve to date the Moosebar shales as early to middle Albian. Both subzones belong to the *Gaudryina nanushukensis* zone of Caldwell *et al.* (1978).

The faunal assemblages associated with the Moosebar Formation indicate an open marine to marginal marine affinity since no freshwater fossils have ever been recorded (McLean and Wall, 1981). Water depths for the seaway probably did not exceed 100 m. Trace fossils recorded in this rock commonly include *Planolites*, *Helminthopsis* and *Chondrites*.

The contacts with the Bluesky Formation, in the Peace River Plains, and with the Chamberlain Formation south of the Sukunka River, are both very sharp and conformable. The abruptness of the lithologic changes, from the sandstones of the Bluesky Formation to sandstones and coals of the Chamberlain Formation to deep-water shales of the Moosebar Formation, suggests an abrupt change in water depths as a result of the rapid transgression of the Moosebar Sea.

2.3.6 *SUMMARY*

Up to this point, the discussion has centered on fitting the Bluesky Formation into a regional context. It has been suggested that both the Bluesky and Chamberlain successions should be given formational status within northeastern British Columbia. The Bluesky Formation is underlain by non-marine strata of the Gething Formation and, in turn, overlain by the Moosebar Formation in the Peace River Plains. In the Sukunka-Wapiti River region, the Bluesky Formation is immediately overlain by the Chamberlain Formation (Figure 2.12).

The precise age of the Bluesky is as yet undeterminable due to the paucity of diagnostic marine fossils. The Bluesky Formation, however, is suggested to be of early Early Albian age, occurring in the *Cleoniceras* biostratigraphic zone. This zone is coeval with the microfaunal subzone of *Rectobolivina* sp. and possibly in part to the *Trochamminia mcmurrayensis* subzone of Caldwell *et al.*, (1978). As a result of contemporaneous transgression into central Alberta, Bluesky strata may be time equivalent to Ostracod Member deposition, but earlier than Glauconite Sand or Wabiskaw Member deposition

The Bluesky-Chamberlain deposits represent a transgressive-regressive event as part of a series of larger depositional cycles. These cycles were probably to a great degree influenced by active tectonism on the structural front, rather than by pure sediment loading and subsidence.

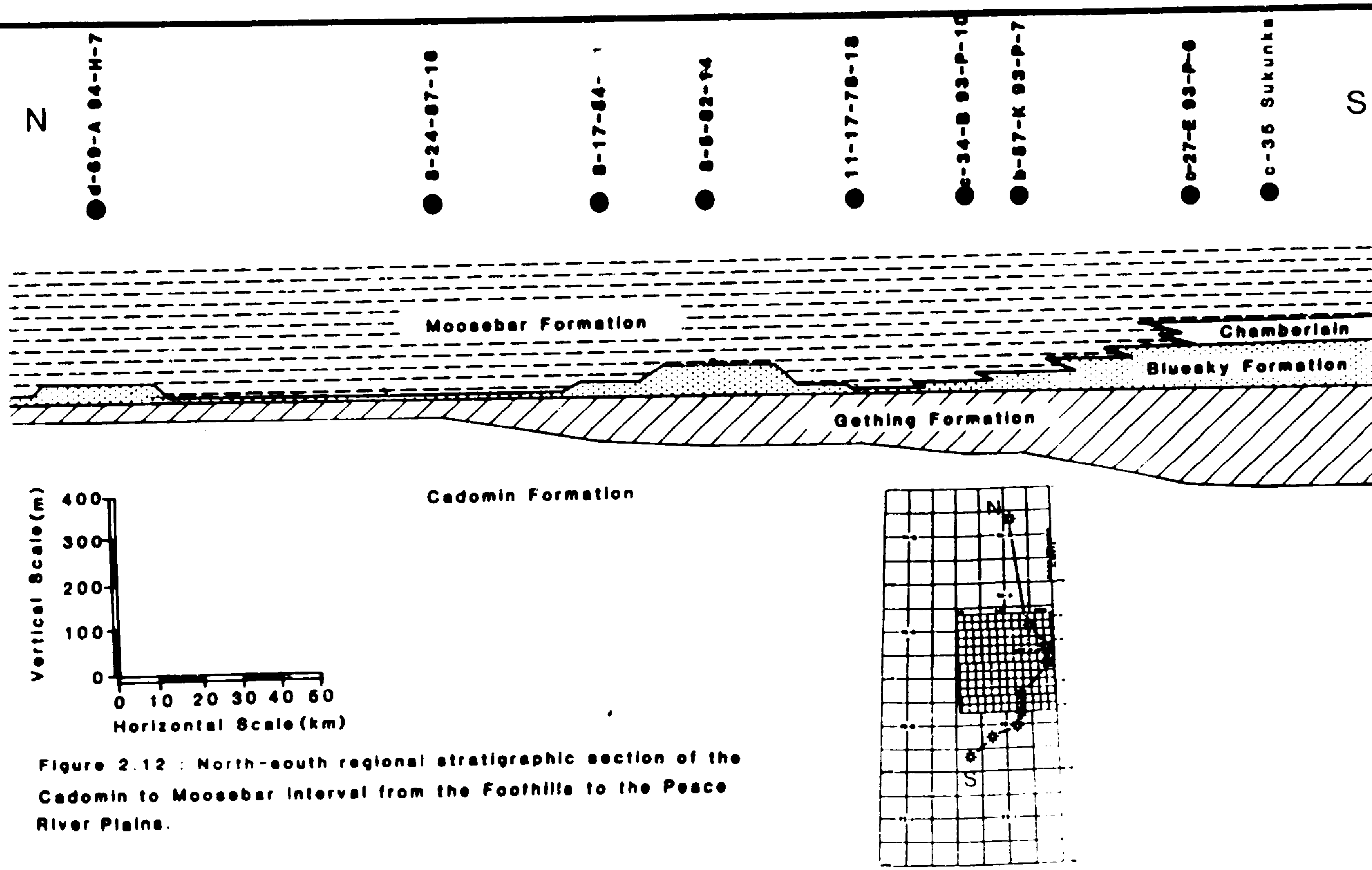


Figure 2.12 : North-south regional stratigraphic section of the Cadomin to Moosebar interval from the Foothills to the Peace River Plains.

Transgressive events are considered to be a response to loading in the foredeep. The Moosebar Formation is considered to represent one such occurrence of complete flooding of the entire basin caused by tectonic loading.

3.0 SHELF PROCESSES AND PRODUCTS

Sedimentological studies of modern continental shelves over the past few decades have provided new insight into the physical and biological processes and produc ts that are prese t in the shallow marine environment. One of the objectives of these studies is to supply geologic models for use in the recognition and interpretation of ancient shelf deposi . Conv ertain aspects of ancient shelf deposits aid in interpreting the mechanisms of sedimentation on the modern continental shelf (Howard, 1972; Harms *et al*., 1975). However, because of some fundamental differences between modern and ancient settings, the task of adequately interpreting ancient rocks has been somewhat limited. The basic difference between the ancient foredeep deposits and modern deposits are found in the basin configuration and the contained sediments. Modern continental shelves are fairly narrow, average 75 km from the shoreline to the self-slope break (>200 m water depth) and face deep ocean b Figure 3.1). In a typical coastal profile relatively new coastal sediments pass seaward into "relic" sediments of once exp terraces, valleys and channels and "palimpsest" sediments of the shelf (Swift *et al*., 1971). hese are products of the recent postglacial Holocene transgression. As a result, sediments on modern shelves do not necessarily reflect modern shelf processes and, in fact, these sediments have had less than 5000 years to adjust to their present conditions (Harms *et al*., 1982).

In contrast, the bulk of ancient foredeep deposits, including those of the Bluesky Formation, were formed in broad, shallow intracratonic seas (epicontinental or epeiric seas) which contained no discernable shelf-slope break or deep-ocean basin (> 1000 m). Instead, epicontinental shelves are divided into depth-related features: the littoral zone (nearshore) and the neritic zone (inner and outer offshore). However, it is considered that many of the modern-day hydraulic processes observed the continental shelves, such as tidal and oceanic currents, had operated

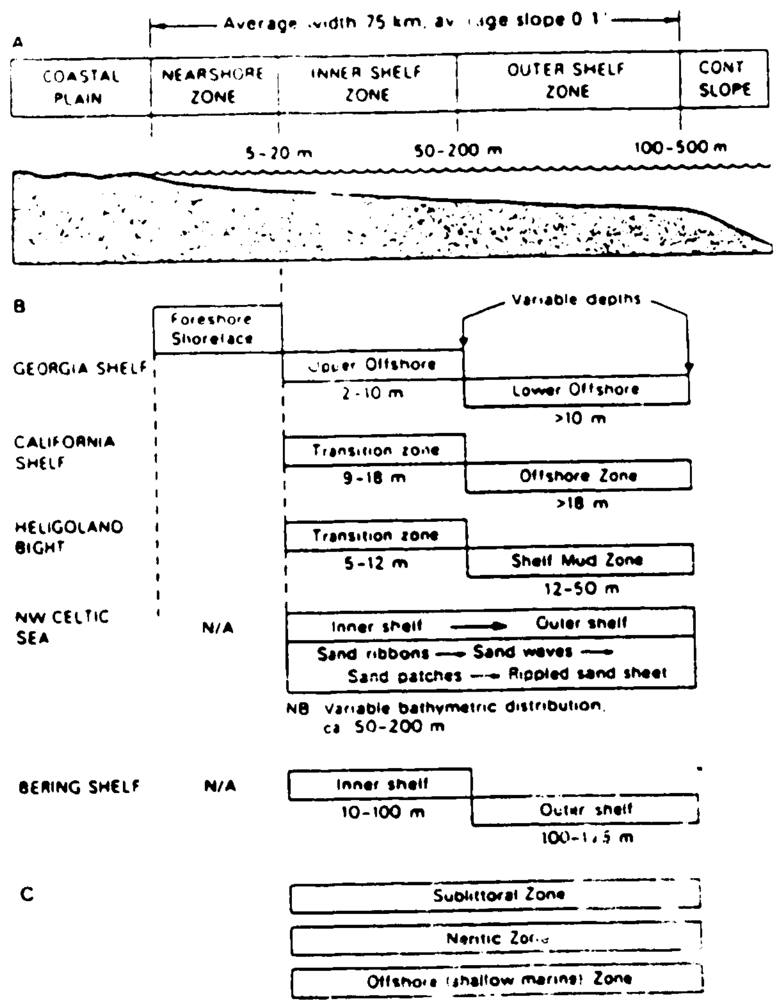


Figure 3.1: Summary of nomenclature used to describe shelf zone and its bathymetric sub-divisions. (A) Idealized shelf profile; (B) Typical depths and terminologies used in modern profiles; (C) nomenclature used for ancient sequences (modified from Johnson and Baldwin, 1986).

simarily in the past. Therefore, these models can be used, to some extent, as partial analogues for comparative studies of ancient shelf deposits and their respective hydraulics.

3.1 MODERN SHELF CURRENTS

The following discussion will review the origin and nature of shelf currents influencing the entrainment, transportation and dispersion of material on modern continental shelves. The four main types of shelf currents are: (1) oceanic circulation (semi-permanent currents), (2) tidal currents, (3) meteorological currents, and (4) density currents. Empnasis will be placed on meteorological- and tidal-currents, as the others show little effect on Bluesky depositional processes. As well, bedforms and products of sedimentation on oceanic-, tide-, and storm/wave-dominated continental shelves will be examined.

3.1.1 OCEANIC CIRCULATION CURRENTS

Semi-permanent large scale movements of water in the oceans are known as oceanic circulation currents. The interaction of the atmosphere with the ocean produces two vertically stacked types of circulation: a wind-driven (surface currents) circulation and a thermohaline or density circulation. Wind-driven circulation is the most shallow and stronger of the two types of currents.

Wind-driven currents are directly produced by the shear stress exerted on the ocean surface by prevailing global wind systems. The global wind patterns are controlled primarily by the differential distribution of heat between equatorial and the polar regions and by the Coriolis forces produced by the earth's rotation. The Coriolis force will cause the moving air masses to deflect towards the right in the northern hemisphere and towards the left in the southern hemisphere. Thus, the global wind system is separated latitudinally into large eliptical cells of circulation known as the trade winds in the low latitudes and the westerlies in the mid-latitudes (Ross, 1975). Patterns displayed by surface currents, therefore, should reflect the prevailing wind patterns of the globe. However, the currents also suffer a gradual decrease in velocity and change in direction with increasing depths to about 100 m. This results in a spiral-shaped current pattern known as the Ekman spiral, in which the net motion of the entire mass of moving water ends up flowing at right angles to the wind direction (Ross, 1975). Examples of wind-driven currents are the Agulhas current and the Brazil current.

Deep water thermohaline circulation currents result from density variations and temperature layering in an ocean. Specifically, these currents are generated by sinking dense, cold water in the polar regions slowly flowing towards the equator. These currents are considerably slower (1-2 cm/sec) and more complex in flow directions than surface currents (Gross, 1977). They are also diverted and deflected by sea floor topography.

Major oceanic currents, created by effects of wind, temperature and density variations, generally lie oceanward of the shelf edge. However, it has been suggested that these currents actively interchange waters of the shelf area and also encroach upon the shelf itself (Johnson and Baldwin, 1986). These large-scale meanders onto the shelf occur seasonally, such as when the Agulhas current impinges on the African shelf during the winter months.

3.1.2 TIDAL CURRENTS

Oceanic tides, which are the cyclical daily rise and fall of sea level, result from the gravitational attraction exerted on the surface water of the Earth by the Moon, and to a lesser extent by the Sun. Tidal ranges and movements are usually complex due to the positional changes of the Moon in relation to the Earth, the Earth-Moon system around the Sun, an uneven distribution of water on the Earth's surface and by irregularities in the morphology of oceanic basins.

During a tidal cycle, the gravitational attraction of the moon pulls the collective ocean mass on one side of the Earth into a tidal bulge. This is counterbalanced on the opposite side by a tidal bulge resulting from a centrifugal force generated by the orbital motion of the Earth and Moon about each other (Ross, 1975). Tides may be diurnal, which are characterized by one high-water mark and one low-water mark during a tidal day; semi-diurnal tides have two high- and two low-water marks; or mixed tides, which demonstrate various combinations of both diurnal and semi-diurnal tides. In addition to the daily tide cycles are fortnightly variations in the strength and range of the tides (Gross, 1977). When the Sun and Moon align the co-gravitational effects cause a maximum or "spring-tide"; when the Sun and the Moon are at right angles a relatively weak "neap tide" occurs. Additional considerations that will affect tide-generating forces involves seasonality and the changing distances of astronomical bodies. Tidal ranges are greater when the Earth is at its closest distance to the the sun (perihelion) and when the moon is at periagee in its orbit around the earth. In both cases, the tide-generating forces are at a maximum due to the closeness of the bodies.

Tidal currents are produced when tidal waves, generated in the deep ocean basin, are displaced onto the continental shelf. The strength of the tide depends upon the oscillation period of the deep basin, and is greatest when this oscillation period matches that of the principal tide-producing force (Johnson and Baldwin, 1986). The oscillation period of a basin is determined by its morphology and water depth. Generally, continental shelves are too small and shallow to generate their own tidal effect. Thus, enclosed seas or those with only a small aperture to an open ocean will usually be tideless or weakly tidal.

In the open ocean the Coriolis force causes the standing waves to be converted to progressive rotating waves which change direction constantly. Within a partly enclosed sea, such as the North Sea, the tidal currents follow this rotary pattern around an amphidromic point or an axis of zero tidal range (Houbolt, 1968). When the current approaches the shoreline, the shoaling effect and the constraints of the coastline converts the rotary currents to alternating flood and ebb currents. The intervening slack-water periods occur between changes in the tidal current direction.

3.1.3 METEOROLOGICAL CURRENTS

One of the most dynamic circulation currents on the continental shelf is the meteorologically-induced current. Generally, these are the most effective sediment transporting agents on open shelves when tidal influence is negligible (Johnson and Baldwin, 1986). Four main types of currents result specifically from meteorological forces: (1) wind-driven currents; (2) oscillatory and wave-drift currents; (3) nearshore, wave-induced longshore and rip currents; and (4) storm-surge.

Wind-driven currents are large scale surface currents generated from the application of wind shear stress on the water surface. Since atmospheric disturbances, such as mid-latitude pressure cells, directly control the magnitude and [illegible]tion of wind driven currents, their occurrence are temporal. Nearbed velocities up to 80 cm s at depths of 50-80 m on the Pacific continental shelf have been recorded during winter storms (Smith and Hopkins, 1972). These currents are generally unidirectional.

Oscillatory and wave-drift currents result from the orbital motion of water particles as a wave propogates. The wave form itself is induced by pressure differentials between the wave crest and trough as the wind passes over [illegible] air-water interface. In deep water, water particles move in circular orbits with orbital diameters matching the wave height at the surface. The orbital diameters progressively decrease with depth till they are virtually non-existent at a level beneath surface equal to one-half the wavelength (Figure 3.2). As a wave shoals, though, the water particles follow a progressively more elliptical motion which flatten with depth (Clifton, 1976). At the sea floor, the motion is essentially a flat horizontal plane. A wave-drift unidirectional current arises from the asymmetry in velocity of the water particles. Even weak unidirectional currents contain sufficient energy to transport sediment after entrainment (Clifton, 1976).

Two types of currents associated with nearshore circulation are generated as waves approach and break the shoreline. These are: (1) longshore currents and (2) rip-currents. Longshore currents are generated when waves obliquely approach the shoreline. These flow parallel to the shoreline and are confined almost entirely to the foreshore zone (Komar, 1976). Rip-currents are narrow, high-velocity, seaward oriented currents intimately associated with longshore currents (Figure 3.3). Rip-currents are induced from variations in water level from differences in wave height, standing waves or bottom topography (Elliot, 1986). Longshore currents may convert into rip-currents in zones of low water levels or when they flow around the edge of a nearshore bar (Elliot, 1986). Rip-current velocities up to 1 m/s are common and are effective transport agents during storm events (Elliot, 1986).

Perhaps one of the most significant meteorologically-induced flow is the storm-surge current. During storm events, higher amplitude waves deepen the wave base which causes enhanced shoaling, oscillation, and extreme build-ups of water within the beach prism (storm-surge). This excessive coastal set-up of water creates a seaward pressure gradient and the initiation of relaxation (ebb) flow (Hayes, 1967; Morton, 1981; Nelson, 1982b). The seaward-directed ebb is released and is considered to be deflected to the right by Coriolis force to form a resultant geostrophic flow that parallels the isobaths (Morton, 1981) (Figure 3.4). Consequently, sediment

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Figure 3.2: Path of water particle motion in a shoaling wave. (A) deep water wave and (B) a shallow water wave. Parameters which describe waves are: Wavelength (L) is the horizontal distance between successive wave crests; wave height (H) is the distance between wave crest and trough; water depth (H), orbital diameter (do) is the maximum horizontal distance of excursion of water particles as a wave passes (circular motion in deep water, an an elliptical motion in shallow water; ripple spacing (λ); ripple height (η); and average distance from ripple crest to leading trough (β) (from Clifton and Dingler, 1984).

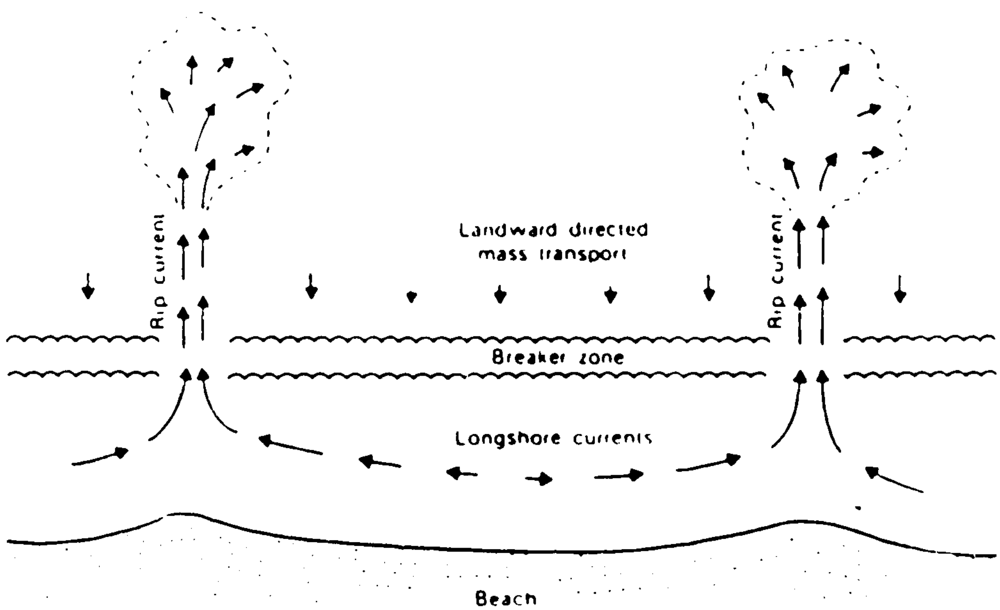

Figure 3.3: Wave-induced nearshore circulation system of longshore currents and seaward-directed rip currents (modified from Elliot, 1986)

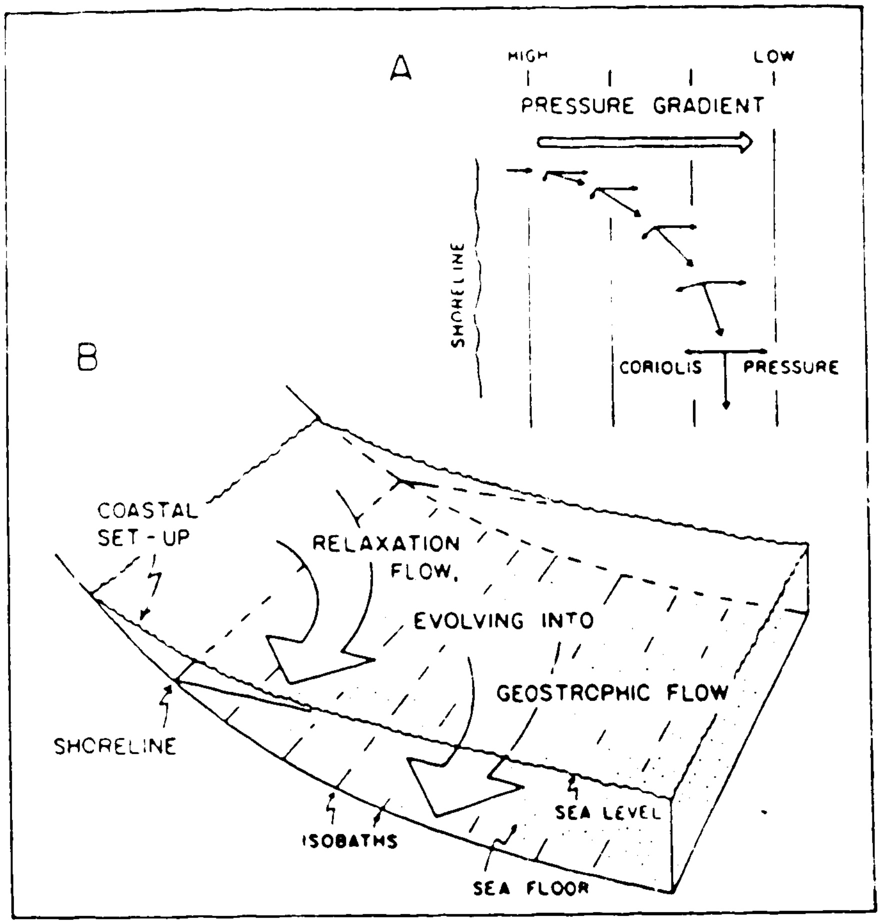

Figure 3.4 Relaxation flow and the generation of storm deposits following the storm surge. Returning bottom waters are deflected to the right (northern hemisphere) by coriolis force to form a geostrophic flow parallel to isobaths (modified from Walker, 1984).

entrained by the storm waves will be deposited parallel to the shore rather than seaward. This storm process of entraining and transporting nearshore sands to the offshore is suggested to be more effective in strongly tidal seas (Johnson and Baldwin, 1986).

3.1.4 SEDIMENT GRAVITY FLOWS

Sediment gravity flows or resedimentation processes are mechanisms that move sediment downslope over the sea floor from shallower to deeper waters by gravitational forces. The four main types of sediment gravity flows are: (1) turbidity currents; (2) grain flows; (3) liquified flows; and (4) debris flows. In turbidity currents, sediments are kept in suspension by the upward component of fluid turbulence which is maintained by the drag of the fluid at its upper and lower boundaries (Middleton and Southard, 1984). Mixing of the flow with the ambient fluid, loss of sediment by deposition and flow separation will eventually stop the turbidity current. Grain flows are visco-elastic flows which are maintained by dispersive pressures generated from grain to grain collisions. These are generally considered to be small-scale sand avalanches and are incapable of depositing a single sand bed greater than a few centimeters. Liquified flows are the failure of a quasi-stable sediment downslope which is internally maintained by upward-moving pore fluids. The grains become suspended and the fabric strength approaches zero. Deposition occurs quite rapidly and the total flow freezes as one complete unit. Debris flows are highly concentrated matrix-supported, viscous sediment dispersions which advance as slow laminar flows. The flow freezes when excess pore pressures are dissipated or when the shear strength of the flow becomes large (Middleton and Southard, 1984). Rarely do these four types of flows occur separately in nature, but probably interact in concert within a flow event.

3.2 MODERN CONTINENTAL SHELF SEDIMENTATION

Sedimentation on modern shelves is not just simply controlled by the hydraulic processes active in the basin. Instead, there are several interactive factors which influence the distribution and magnitude of shelf sediments, including: 1) the rate and type of sediment supply, 2) the type and intensity of the hydraulic system, 3) sea-level changes, 4) climate, and 5) animal-sediment relationships (Johnson and Baldwin, 1986).

In the following sections, three classic situations of modern shelf sedimentation are reviewed, including: oceanic current-dominated, tide-dominated, and wave- and storm-dominated shelves. As well, a brief review of wave- and tide-dominated shoreline sedimentation is provided. These will be only briefly discussed and illustrated; however, the references cited should direct the reader to more complete reviews.

3.2.1 OCEANIC CURRENT-DOMINATED SHELF SEDIMENTATION

Oceanic currents play a significant role in shelf edge sedimentation on continental margins with extremely narrow shelves and those which are situated to receive deflected boundary currents. As mentioned previously, oceanic currents, such as the Gulf Stream, are able to meander, eddy onto and interchange waters with the continental shelf. When these oceanic currents impinge onto the shelf they are responsible for the development of large migrating bedforms. Approximately 3% of continental shelves are oceanic-current dominated (Walker, 1984). Powerful boundary currents have been known to influence sedimentation off the African and South American shelves. However, the best documented example of a powerful boundary current strongly influencing sedimentation is the Agulhas Current along the southeast African continental margin (Flemming, 1980).

The southeast African shelf is microtidal (< 2 m), 700 km long and ranges in width from 10 km in the north to 40 km in the south (Flemming, 1978). The general sand distribution on the shelf is illustrated in Figure 3.5. Flemming (1980) indicated that the relative narrowness of the shelf pl the anomalously steep continental slope (12^{o}) allows the Agulhas Current to impinge upon the shelf. According to Flemming, once on the shelf, the Agulhas Current is able to meander nearshore, entrain and transport sand southwards to the outer zone of the shelf.

The outer zone of the shelf is dominated by the Agulhas Current (Figure 3.5) and contains a mobile sand stream on which a variety of longitudinal and transverse bedforms are imprinted. Large sandwaves or dunes, ranging up to 17 m in height with wavelengths of several hundred meters, are replaced offshore by sand ribbons and patches (Johnson and Baldwin, 1986). Walker (1984) suggested that the associated high face angle (25^{o}) in the sandwave may form thick (> 2 m) sets of angle of repose cross-bedding. Walker also indicated that since the Agulhas Current only reworks relict shelf material and does not supply new sediment, it will form only a deposit no greater than a few meters thick.

3.2.2 *TIDE-DOMINATED SHELF SEDIMENTATION*

Approximately 17% of the world's continental shelves are swept by strong tidal currents which produce a wide variety of bedforms (Walker, 1984). Most of these shelves experience diurnal tidal cycles with ranges greater than 3 m and surface current speeds ranging from 60 to > 100 cm/s. An exceptional tidal range of > 15 m (with current speeds 1-2 m/s) is found in the Bay of Fundy (Thurman, 1978). Tidal sedimentation bedform morphology has been examined in the Gulf of Korea (Off, 1963), the Georges Bank (Twichell, 1983), the New England seaboard (Kumar and Sanders, 1975), and the Bay of Fundy (Klein, 1970). One of the best documented and most

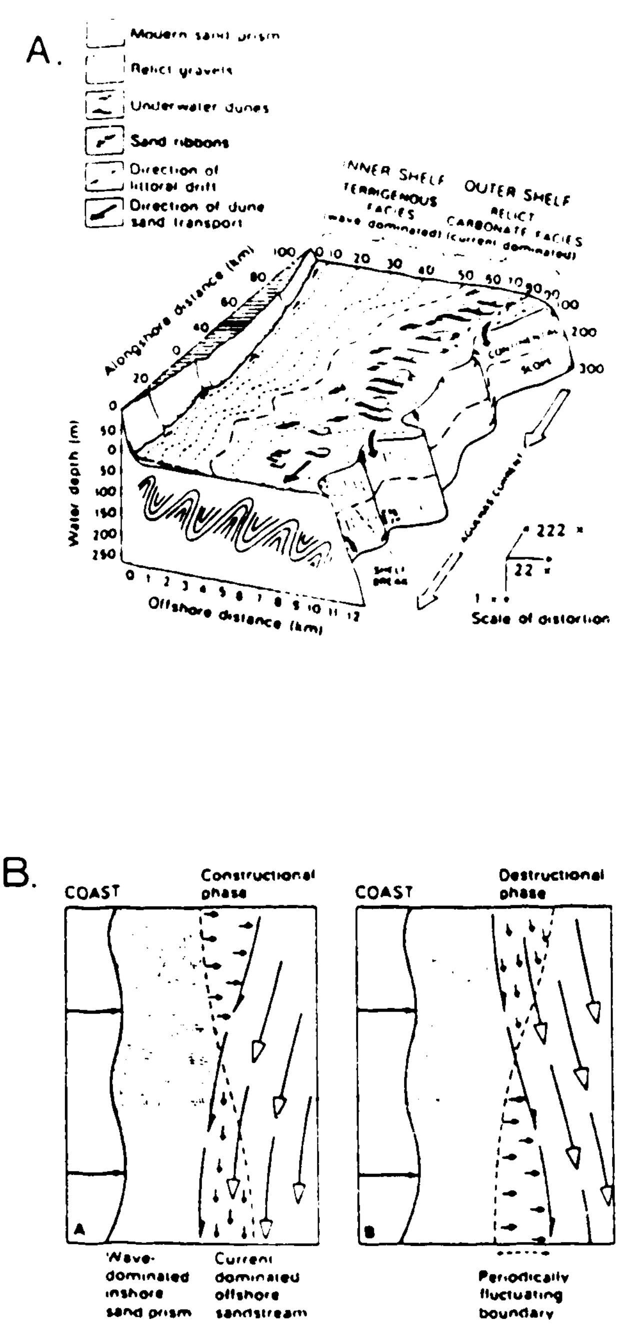

Figure 3.5 (A) Distribution and orientation of bedforms along transport paths on the S.E. African shelf; (B) Shelf sedimentation controlled by migration of the Agulhas oceanic current. When the oceanic current sweeps inshore it erodes seaward edge of the inshore zone (modified from Flemming, 1980).

striking examples of a tidally-dominated shelf and coast is the North Sea (Belderson and Stride, 1966; Houbolt, 1968; and Stride, 1982).

Nearly all tidally-produced bedforms, such as sand ridges and sand waves, are current-oriented either longitudinally or transversely even though tidal currents are bidirectional, rectilinear or rotary in flow (Terwindt, 1971). It has been suggested that they develop unidirectional transport paths either because: (1) of an asymmetry in strength of ebb and flood current velocities, (2) the ebb and flood tidal currents may follow separate paths, (3) other types of currents such as wind-driven currents may enhance one tidal direction, or (4) the lag effect associated with a rotary tide delays the entrainment of sediment (Johnson and Baldwin, 1986).

The morphology and distribution of bedforms within a tidally-dominated area is empirically related to the near-surface mean tidal current velocities (Stride, 1982) and by the sediment supply (Belderson *et al.*, 1982). This is analagous to the flow regimes associated within fluvial settings where certain current speeds have associated characteristic bedforms. Five main sedimentary features generally result from differing tidal current velocities, including: sand ribbons, sandwaves, sand patches and mud zones (Figure 3.6).

Furrows and scour hollows are current-parallel seafloor erosional features which are confined to areas with tidal currents greater than 150 cm/s. Furrows can be up to 8 km long, 30 m wide and 1 m deep, and can be found in gravel, sand or mud floors (Belderson *et al.*, 1982). Examples of scour hollows greater than 28 m deep have been recognized in the Irish Sea (Belderson *et al.*, 1982).

Sand ribbons are current-parallel bedforms developed in zones of tidal currents in excess of 100 cm/s (Kenyon, 1970). These consist of fine bands of sand up to 200 m wide by 15 km long

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Figure 3.6 Distribution of bedforms produced by tidal currents on a continentalshelf (A); at times of low sand supply (B); and with high sand supply (C). Bedforms are aligned parallel with mean spring peak near surface tidal current velocities in cm s (from Belderson, Johnson, and Kenyon, 1982).

and generally less than 1 m thick. Sand ribbons are typically found in the hydraulic zone between the erosional and sand wave zones (Belderson et al., 1982). These features are usually transistory and are often reworked into sandwaves when current speeds or sand supply change with time (Kenyon, 1970).

Sandwaves are flow-transverse bedforms with straight crests and well developed lee-slopes. Sandwaves vary in height between 3 and 15 m high with wavelengths between 30 to 500 m (Johnson and Baldwin, 1986). Terwindt (1971) has documented sand waves up to 1250 m in length in the North Sea. The sand waves range from symmetrical to strongly asymmetrical in form (Reineck and Singh, 1980). The asymmetrical cross-section has been suggested to be related to the asymmetry of the tidal flow with the lee side facing downstream to the strongest tidal current (Johnson and Baldwin, 1986). Lee-slopes on asymmetrical forms range from about 17^{0} to 35^{0} but are most commonly greater than 20^{0} (Belderson et al., 1982).

The sandwaves occur where current velocities are greater than 65 cm/s in the hydraulic zone between the sand ribbon zone and the proximal sand patch zone (Figure 3.6). Smaller sand waves represent an end-member and require a slightly lesser current velocity for formation (Belderson et al., 1982). Internally the sandwave is composed of low angle set boundaries with superimposed high angle cross-stratification (from the smaller-scale megaripples) and rare mud drapes. Sandwaves have also been termed giant ripples and dunes (Reineck and Singh, 1980). These features have been known to occur in a variety of tidal environments with well documented examples from the Rhine-Meuse Estuary (McCave, 1971), the German Bight (Reineck, 1963) and the southern North Sea (Belderson et al., 1982).

Sand patches occur in areas of relatively weak (< 50 cm/s) tidal currents (Figure 3.6). These are widespread features and are aligned more or less flow-transverse or longitudinal to the peak tidal flow (Johnson and Baldwin, 1986). It is considered that these oscillatory velocities are insufficient to move the sand and thus must be enhanced by storm flow. The sand patches were first recognized by Kenyon (1970) in the Celtic Sea where they tended to be thin, tabular, well-sorted sand fields with an arcuate to ragged shape. These sand patches can occur up to 3 m in thickness (Belderson *et al.*,

At the ends of the tidal current transport path are zones of mud accumulations (Johnson and Baldwin, 1986). The deposition of mud along continental shelves is influenced by wave effectiveness, suspended sediment concentration and tidal current shear stress (McCave, 1971). Mud is generally found in a wide variety of settings, but usually it accumulates in moderately deep water (> 30 m) where wave effectiveness is low. However, if excessively strong tidal currents are present, as in the St. Georges Channel, mud is swept clear from the sea bed even at depths of over 100 m (Johnson and Baldwin, 1986).

The largest bedform most distinctive of tidally active seas are the 'tidal sand ridges'. The best examples of these occur in the North Sea where they can be up to 50 km long, 6 km wide and 40 m high (Belderson *et al.*, 1982). These are linear features with their long axis oriented at a small oblique angle (0-20^{o}) relative to the strongest tidal current flow direction. Some tidal sand ridges are strongly arcuate or sinuous in plan view and most have asymmetrical cross-sectional profiles. When this occurs, the steeper slope faces obliquely downstream to the net sand transport direction. Therefore, this feature can generally be used as an indicator of flow direction.

Due to the inequality of the tidal currents producing these structures, the internal morphology is quite complex. The internal structure of the sand ridge is comprised of a series of low angle master bedding planes (< 10^{o}) separated by small-scale cross-stratification produced by obliquely migrating sandwaves and megaripples (Houbolt, 1968). The migration of these smaller bedforms is the result of tidal currents which entrained sand converging towards the crestline of the ridge.

Sand ridges of tidally-dominated seas can occur either as clusters in offshore and estuary settings or singly as nearshore coastal features on the lee of islands or submerged shoals (Belderson *et al.*, 1982). Because their present location is considered to be geologically anomalous it has been suggested that these sand ridges represent relict features from a earlier lowstand nearshore wedge which are now being modified under present conditions (Swift, 1975).

Other sedimentary structures that are more or less diagnostic of tidal regimes are mud drapes, reactivation surfaces and herringbone cross-stratification. Mud drapes are distinctive features in tidal deposits and are usually well-developed in modern subtidal environments such as estuaries. The thin mud layers are considered to be produced during slack water periods in the tidal cycle. Double mud drapes with a thin parting of sand are suggested to be diagnostic of two slack periods with the intervening sand deposit resulting from subordinate tidal current (Johnson and Baldwin, 1986). Reactivation surfaces are suggested to be formed by erosion of the bedform by the weaker reversing tidal current (Klein, 1970). Herringbone cross-stratification is probably the [illegible] diagnostic feature for tidal deposits. This bedform consists of co-sets of oppositely dipping foresets laminae which preserves the bidirectionality of the tidal currents. This type of cross-bed suggests that both ebb- and flood-tidal currents were effective in transporting bed-load

3.2.3 *TIDALLY-DOMINATED SHORELINE SEDIMENTATION*

The influence of tidal processes is also recorded in the shorelines of tidally-dominated shelves. The most significant environments in terms of its areal distribution and lithofacies development, is the tidal flat setting. Tidal flats form extensive stretches of shoreline in meso- and macro-tidal coasts, lagoons, estuaries and tidally-influenced deltas. A wide variety of physical and biogenic structures occur in response to tidal and wave activity. Most tidal flats are usually featureless plains dissected by channels where sediments accumulate by vertical accretion and progradation of the flat and channels. In general, during a tidal advance, waters enter the channels overtop the banks and inundate the adjacent flats. During slack water, the waters retreat via the tidal channels and re-expose the flats.

Tidal flats are subdivided into supratidal, intertidal, and subtidal environments. Supratidal flats occur above mean high tide level, and are commonly covered with salt marshes characterized by interlaminated silts and clays. Rootlets, mud-cracks and biogenic structures extensively disrupt the laminae (Elliot, 1986). Intertidal flats are comprised of a mud-dominated zone near the high water mark and a sand-dominated zone near the low water mark. The intertidal flats are characterized by prolific flaser, wavy, and lenticular bedding patterns, reflecting a mixture of strong tidal currents and waves alternating with slack water times (Elliot, 1986). Dissecting the intertidal flats are highly sinuous tidal channels and creeks which deposit minor sand bar sequences along it's inner banks.

The intertidal flats pass laterally into the subtidal zone. The subtidal zone is generally featureless but can contain numerous extensive channel bars and shoals. This zone contains medium and large-scale cross-beds produced by channels and fine grain sands with ripple lamination produced in the shoals and bars. Progradation of a tidal flat will typically produce a fining-upward sequence consisting of coarser-grained material from the subtidal zone at the base to mixed sand and mud beds from the low intertidal sand flats, and capped by mud-dominated deposits of the high intertidal and supratidal mud flats (Klein, 1970; Elliot, 1986).

Estuaries occur in association with tidal flats, river deltas and barrier islands and are subdivided into two types: the tidally-influenced riverine estuary and the drowned paleo-valley. The tides, in the estuary, firstly affect the degree of salt and fresh water mixing and secondly, the tidal currents may influence sediment transport and deposition. The nature of the deposited sediments is influenced by the changing velocities of the tidal currents, the reversing nature and symmetry of the ebb-flood cycle, and the variation of tidal currents along the length of the estuary (Off, 1963).

In tidally-influenced estuaries, the lower reaches, comprised of partially to completely mixed circulation, pass upstream to tidally-influenced meandering channels before reaching the fluvially-dominated river system. While sand is normally concentrated in the central part of the estuary, silt and mud accumulate along the margin on the fringe of the intertidal flats. Sand is deposited as either transverse or longitudinal-oriented subtidal bars or ebb- and flood-oriented deltas (Elliot, 1986). Sedimentary structures associated with the estuarine bars range from current ripples to sand waves. Herringbone cross-beds are common features in current ripple lamination. Bioturbation patterns are also influenced by the changes in salinites of the estuary and, as such, show almost as much variation in distribution as the bedforms (Ekdale *et al.*, 1984).

Several sedimentological studies of modern estuaries have been provided recently, including a study of the Dutch Rhine delta (Terwindt, 1971) and the Georgia coast (Howard and Frey, 1980). A review of the typical physical characteristics of estuarine sediments has been summarized by Elliot (1986) and Greer (1975). Reviews of the biological aspects of brackish water environments is found in Howard and Frey (1975, 1980) and Ekdale *et al.*, (1984).

3.2.4 WAVE- AND STORM-DOMINATED SHELF SEDIMENTATION

In many shallow marine environments where tidal currents are weak, meteorologically-induced currents dominate the hydraulic regime and sedimentation patterns on a daily basis Wind- and wave-induced currents are also characteristically strongly seasonal, with their most intense period falling in the winter months. Also, shelves which face the prevailing westerly winds (ie. Bering Sea, Washington-Oregon shelf) are subject to more intense wave action than those shelves in the lee of prevailing winds (Atlantic shelf, Gulf of Mexico). Since most of the shelf off North America is wind storm-dominated, these are now almost as well understood as tide-dominated seas (Swift *et al.*, 1971). Numerous studies concerning storm and wave-dominated shelves have been completed on the Oregon-Washington shelf, the eastern United States Atlantic shelf, the Bering Sea and the Gulf of Mexico. Excellent studies of beach profiles and process-response to wave action have also been provided by Clifton (1976), Davis (1978), Howard and Reineck (1981), and Elliot, (1986) Detailed reviews of high energy barred coastlines and barrier islands have been provided by Davidson-Arnott and Greenwood (1976), McCubbin (1982) and Reinson (1984).

Wave- and storm-dominated shelves and coastlines are modified by wave-drift (wave surge) and direct wind-driven currents, which are usually most intense during the winter months. Normally during summer, wind-induced currents are generally only capable of reworking the sediment surface on the inner shelf Silts and muds are deposited from suspension and biogenically reworked on the middle and outer shelf areas. During fair weather conditions, waves are relatively low amplitude, long period swells with a shallow wave base. As the wave shoals and interacts with the shallowing shelf, oscillatory flow is gradually replaced by asymmetrical landward flow of higher energy. The bedforms on the shelf bottom reflect this changing wave regime and symmetrical ripples produced in the oscillatory wave zone grade up-slope to asymmetrical wave ripples in the shoaling w[illegible]e zone followed by flat beds produced in the breaking, surf and swash zones (Clifton, 1976) (Figure 3.7).

Conversely during stormweather conditions, the intensity of the wave action can increase greatly from strong winds or oceanic storm waves [illegible]ich lowers wave base and er[illegible]es and transports sand and silt, both as bedload and in suspension, across the shelf to the shelf-edge. R[illegible]tary tidal curre[illegible], although sometimes weak, can enhance this process. Significant modification of nearshore sediments result from storm-surge currents and relaxation (ebb) currents (Hayes, 1967; Morton, 1981; Nelson, 198[illegible] [illegible] 38).

Examples of storm-generated beds have be[illegible]mented [illegible] many shelf areas, such as the German Bight (Aig[illegible]eineck, 1982), Norton Sound area (Nelson, 1982a), Gulf of Mexico (Hayes, 1967), [illegible]Kumar and Sanders, 1976) and California Shelf (Howard and Reineck, 1981). [illegible]y a storm sequence consists of an eroded, sculptured base, a ba[illegible]al lag deposit of mud clasts, followed by horizontal to low angle lamination, wave ripple cross-lamination and topped by a bioturbated interval (Johnson and Baldwin, 1986). Individ[illegible]l storm beds range in thickness from a few centimeters [illegible]few decimeters. The bioturbated interval marks the occurrence of opportunistic organisms taking advantage of the waning currents associated with the stor[illegible] event.

Pr[illegible]mal-distal trends have also been recorded in storm-generated beds. Variation in the storm layers across the shelf are the result of decaying currents and sediment supply. Proximal storm sands are generally thicker, contain basal lags, and display parallel to low angle lamination. Distal sto[illegible] sands are finer-grained, thinner demonstrate grading with many rhythmic interbeds [illegible]f mud laminae. A summary of proximal to distal trends in storm deposits (Figure 3.8) has been provided by Aigner and Reineck (1982). An excellent example of these proximal-distal trends is

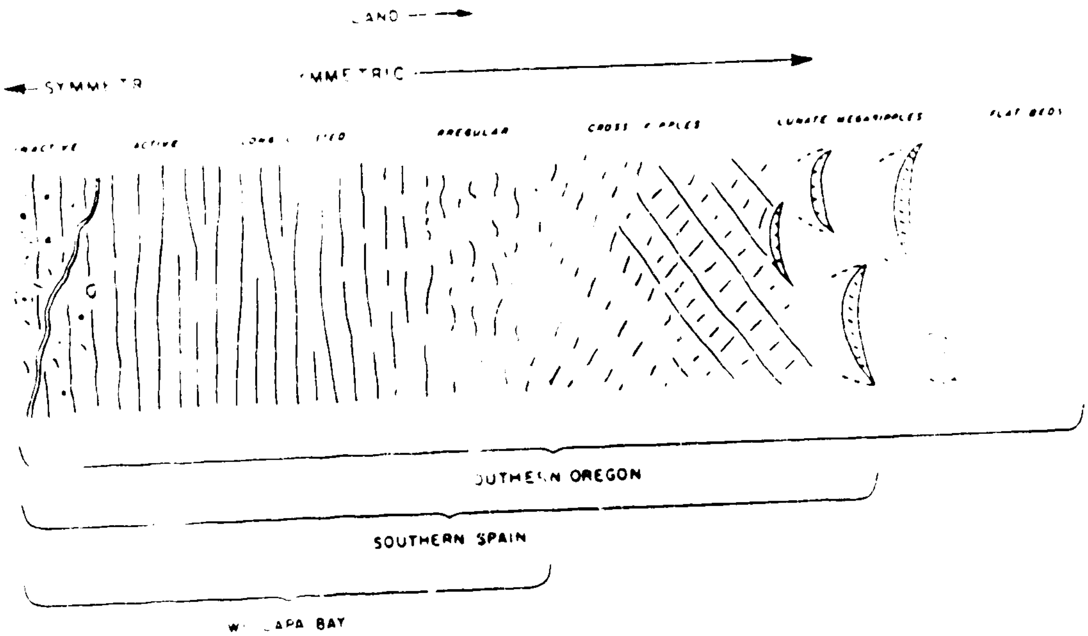

Figure 3.7: Bedforms produced by shoaling waves across the beachface. In a landward direction, inactive storm-generated ripples in the offshore-transition area are bioturbated during fairweather conditions; these pass into symmetrical ripples followed by asymmetric bedforms to lunate megaripples as wave shoals in the shoreface; flat or plane-bed flow conditions occur in the breaker, surf and swash zone of the foreshore (modified from Clifton, 1976).

found in the texturally graded ielf - prodelta sequence of the Yukon Delta in the Norton Sound area (Howard and Nelson, 1982; Nelson, 1982a). Graded storm layers in the sound, typically demonstrate a progressively seaward decrease in grain size and thickness of sand beds and characteristically display waning flow sequences which mimic turbidites (Howard and Nelson, 1982). The amount of bioturbation in the storm deposits also increases towards the distal end-member (Nelson, 1982a).

Hummocky cross-stratification is considered to be the most diagnostic feature of storm sedimentation. This sedimentary structure, as described by H ms *et al.*, (975), has been widely reported in wave-dominated shelf sediments and is accepted to reflect storm-wave deposition above storm wave base. These b s are composed of alternating concave and convex laminae and are typically interbed d with bioturbated beds. The precise mec ism of formation is still under considerable debate as to whether purely oscillatory or combined flows are responsible

t and Bourgeois (1982) suggested that hummocky cross-stratification is produced by the scouring and draping action of storm-generated, large scale oscillatory currents. Harms *et al*. (1982) and Duke and Leck 86) also suggest that fine sands can be moulded into hummocks and swales by purely high oscillatory flows. On the other hand, Swift *et al*. (1983), Aigner (1985), and Nottvedt and Kreisa (1987) argue that astrong unidirectional component combined with oscillatory flow would form hummocky cross-stratification. Leckie (1988) indicated that hummocky beds are intimately associated with coarse grain ripples which are considered to be formed by oscillatory flow combined with a shore-normal unidirectional component. Swift *et al*. (1983) have named these sea-floor features 'hummocky megaripples'. Duke (1985) provided numerous case examples of hummocky cross-stratification produced by intense winter storms and tropical hurricanes.

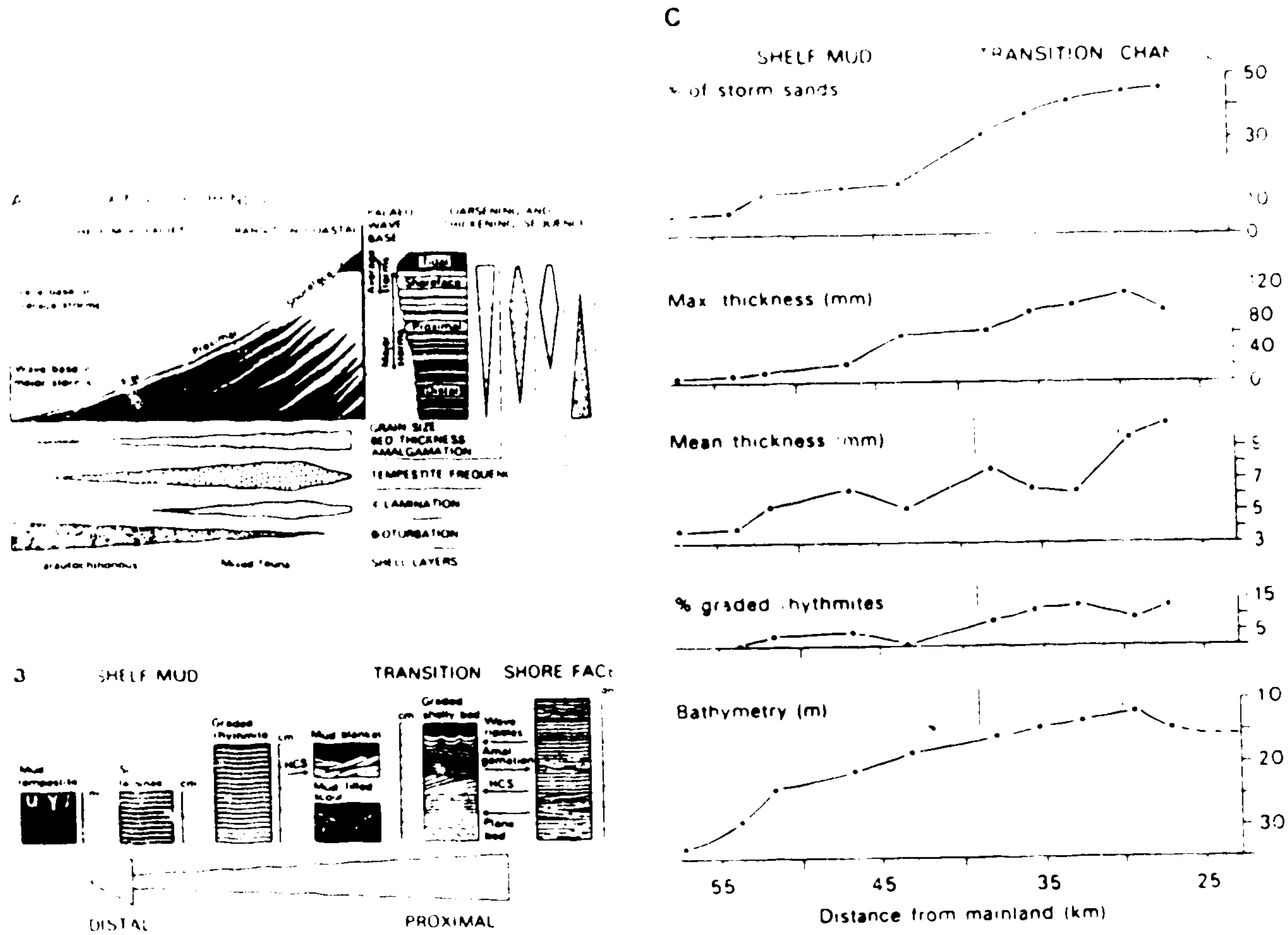

Figure 3.8: Proximal-distal trends of storm deposits on the shelf and nearshore of the German bight. (A) Proximality trends for bedforms and bioturbation. (B) Lateral variations in storm sequences from the shoreface to the shelf. (C) Quantitive description of storm deposits (modified from Aigner and Reineck 1982)

3.2.5 WAVE- AND STORM-DOMINATED SHORELINE SEDIMENTATION

A number of subenvironments, each with distinctive characteristics, are commonly used to describe a wave-dominated shoreline, including the offshore-transition, shoreface zone, foreshore and backshore (Figure 3.9). The offshore-transition zone extends from mean fairweather wave base to mean storm wave base, and is subject to periods of intense wave action during storm events (Johnson and Baldwin, 1986). Normally during fairweather periods, silts and muds are deposited from suspension and biogenically reworked. However, during storms, bottom sediments are subjected to strong wind-driven- and storm surge ebb-currents which generally deposit graded and cross-laminated sand beds. Thus, fairweather deposits of bioturbated silts and muds alternate with silt-sand beds of storm origin.

The shoreface zone (Figure 3.9) extends from mean low water level to fairweather wave base and is sub-divided into three zones: upper, middle and lower. In normal conditions, the upper shoreface is subjected to breaking and surf currents. The middle and lower shoreface is subjected to progressively weaker oscillatory and shoaling waves with depth. Generally, along the shoreface profile, with increasing water depths, bioturbation gradually increases, grain size decreases and bedforms change from flat beds to asymmetric ripples followed by symmetric ripples. During storm events storm-surge currents erode the beach prism and upper shoreface and redeposit the sediment on the lower shoreface or beyond by relaxation (ebb) currents. Thus, the upper shoreface is generally sand-dominated, with little distinction between fairweather and storm events except for periodic lags (Johnson and Baldwin, 1986). The lower shoreface consists of interbedded fairweather bioturbated silts and storm deposits of parallel-laminated or hummocky sands. It should be noted that bioturbation is a sensitive indicator of energy conditions in the beach profile, in that it occupies distinct zonation. Burrow forms reflect the behavior of the animal

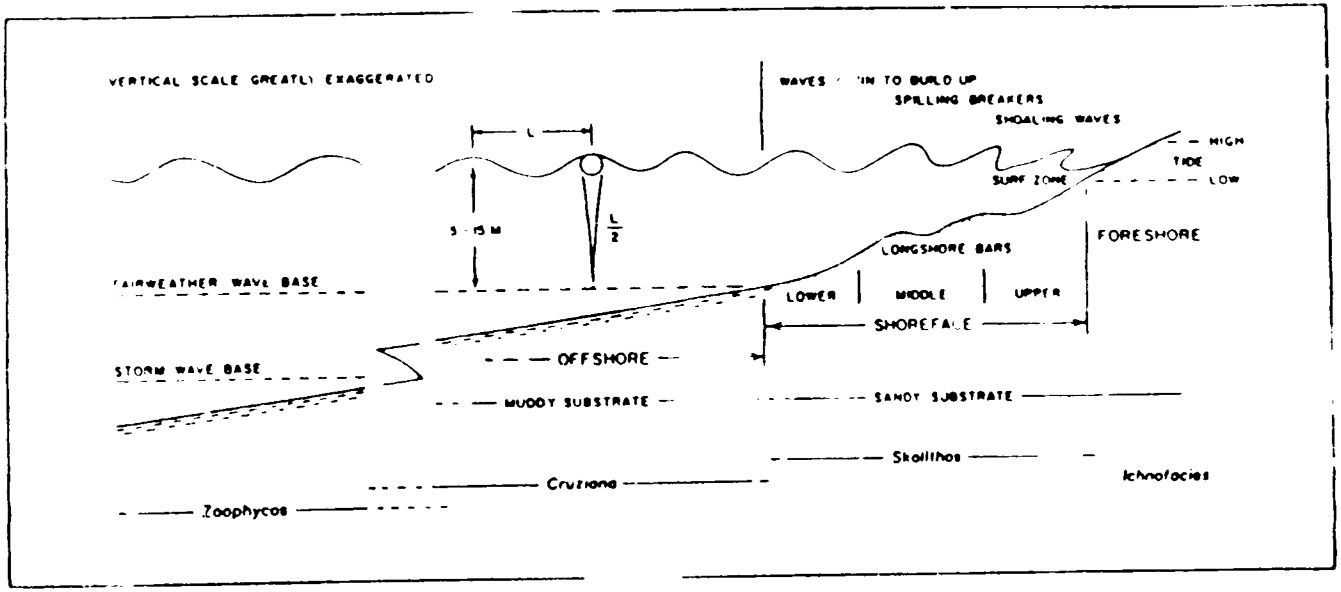

Figure 3.9: Beach to offshore profile locating sub-enviroments, fairweather- and storm-wave base and ichnofacies occurrences. Fairweather wave base commonly lies at about 5-15 m water depth (modified from Walker 1984).

to the prevailing energy conditions. This condition provides the basis for the ichnofacies concept (Seilacher, 1967a; Frey and Pemberton, 1984; Ekdale *et al.*, 1984).

The next sub-environment along the shoreline profile is the foreshore which is located in the intertidal area between the mean high level and the low water level marks. Bedforms reflect the high energy processes of breaker, surf and swash flows with predominantly planar-laminated and current laminated sand beds (Thompson, 19 ; Davis and Fox, 1975). Bioturbation is usually sparse due to the harsh and rugged living conditions of the surf zone (Ekdale *et al.*, 1984). In low energy settings such as the Atlantic coast of eastern United States, symmetrical and asymmetrical wave ripples, rather than planar-laminated beds, are common. Ridge and tunnel topography, which is a series of asymmetric ridges separated by shallow troughs, is often superimposed onto the foreshore. The ridges are suggested to be caused by swash-backwash of the shoaling waves in concert with aggradation of the foreshore (Davis *et al.*, 1972).

The final sub-environment is the backshore area which is located at the mean high water mark. This zone is dominated by sand emplaced during storm events and becomes reworked by wind processes and organisms. This is normally the site for vegetation growth.

3.3 SPECULATIONS ON CRETACEOUS SHELF PROCESSES

Numerous studies on the processes of sand sedimentation on modern continental shelves provides a basis for comparison with Lower Cretaceous deposits examined in this study. Investigations of sand bodies on many of the modern shelves (ie. Bering, Oregon-Washington and the North Sea shelves) illustrate the processes and products of differing hydraulic regimes. They have also advanced the concept that sediments on the modern shelves are composed of relict, palimpsest and new origins (Swift *et al.*, 1979). On the Atlantic shelf, although t[illegible] are suggested to have been stranded by the Holocene transgression, they are presently being modified by a variety of present-day shelf currents (Swift and Field, 1981).

During the Mesozoic, the western interior of North America was flooded by ep[illegible] seas which probably contained, as modern shelves do, a variety of hydra[illegible] re[illegible] me[illegible] [illegible]se-quently, a wealth of marine nearshore and offshore deposits accumulated withi[illegible] [illegible]sin under either tidal-, wave- and storm- or oceanic-current systems. The[illegible] are many e[illegible] s of ancier deposits in the literature which are considered to have been deposited under [illegible] process or another. Marine offshore sands in the Jurassic (Oxfordian) of Montana, interpreted as shallow marine shoals, are considered to have been formed by wave-induced currents (Brenner and Davies, 1974). These are characterized by well-sorted coarsening upward sequences with gradational bases. The lower half of the bar contains highly bioturbated sands which grade upward to wave-rippled and trough cross-bedded sands. The Duffy Mountain Sandstone, in Colorado, are offshore bars suggested to be partly created by oceanic currents combined with storm-generated currents (Boyles and Scott 1982). The Jurassic Fernie-Kootenay deposits in Western Canada are p[illegible]stulated to [illegible]ave been emplaced by the transport of sands from a wave [illegible]ominated delta to the offshore by storm-surge ebb density currents (Hamblin and Walker, 1979). Rapid transgression and shoreface erosion are suggested to have emplaced thin conglomeratic veneers within the Cardium Formation of Alberta (Eyles and Walker, 1988). Bars in the Cretaceous Shannon Sandstones of Wyoming have been described by Spearing (1976) to contain an abundance of tidal structures. A typical vertical profile consists of a coarsening-upward succession of bioturbated shale grading upward to bioturbated sandstones and clean, non-bioturbated cross-bedded sandstones (Spearing, 1976). These are believed to be the products of tidal currents enhanced l[illegible]ally by storm-induced [illegible]urrents. Later these were re-interpreted by Tillman and Martinsen (1984) to be products [illegible]f storm currents enhancing fairweather oceanic currents This combination of currents was said to transport and deposit sands in the form of megaripples and sandwaves. Stacking of these sandwaves and megaripples developed north-south trending sand ridges up to 20m thick (Tillman and Martinsen, 1984).

Although the behavior of tides for the Cretaceous is poorly understood, the Cretaceous seaway of the Western Interior has been suggested by many authors to have experienced micro to meso-tidal regimes (Tillman and Martinsen, 1984). Slater (1985) suggested that tidal systems for the seaway were minor, with tidal ranges of 0.86 m and current velocities up to 10 cm/s. Klein and R... (1978) argued that the Cretaceous epeiric seas were tidally-dominated and that the effects of tides are often overlooked. Their evidence for strong tidal settings is based on the deduction that regular tidal flushings in semi-enclosed basins (lagoons, estuaries) were needed in order to maintain the marine benthic assemblages found in these deposits. They also stated that strong tidal influences are indicated by several forms of bivalves which faithfully recorded neap-spring tidal cycles on their growth rings. In addition, Bridges (1982) stated that since epicontinental seaways in the Jurassic and Cretaceous were connected to the Arctic Ocean, tidal waves propagated within this open ocean would have been displaced onto the North American cratonic shelf, thus, forming significant tidal currents within the basin. However, because of the vast distances travelled by the tides, twice-daily and daily tidal waves probably would have produced reflected waves, which, in turn, leads to the development of amphidromic systems containing central points of zero tidal range (Figure 3.10). Thus, strong reversing tidal currents with a specific tidal range would have developed parallel to- and adjacent to- the western coast of the seaway (Bridges, 1982).

The strength of tides is also said to be significant when certain basin morphologies are encountered. High velocities for tidal currents have been recorded in re-entrants such as Norton Sound (Nelson, 1982). When the tide passes through a narrow aperture, the smaller cross-sectional area acts as a 'throttle point', thereby increasing the tidal current velocity greatly

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Figure 3.10: Speculative reconstruction of the paleotidal regime in the Western Interior Seaway during upper cretaceous times. Tidal ranges are indicated by dashed lines and amphidromic systems are indicated by solid lines. Amphidromic systems indicate rotary tidal currents around a zero node (Belderson *et al.*, 1982).

Also noted on modern continential shelves is that the widest shelves are characterized by the greatest tidal range amplitude and current velocities. Thus, this should also be applicable to ancient epeiric seas (Klein and Ryer, 1978).

Most of the evidence presented for the presence of tides in the epeiric sea is somewhat speculative. The fact that tides were present hinges on the premise that a sufficient aperture to the Arctic ocean existed to allow propagation of a tidal wave. However, the presence or absence of a suitable connection to the Arctic Ocean during the Cretaceous cannot be proven either way. If we assume that Slater (1985) is correct, in that the Cretaceous seaway was microtidal with ineffective tidal processes, then sediment transport, in particular Bluesky sediments, was accomplished by other mechanisms. However, occurrences of tidally-produced bedforms, such as herringbone cross-stratification, were noted in nearshore tidal channels in Bluesky sediments northwestern Alberta (O'Connell, 1988). Thus, the role of tides during Bluesky time should not be fully discounted. The shoreface features noted by O'Connell (1988) though, were considered to be formed in a wave-dominated setting.

As noted earlier, many of the offshore sand bodies and deltas are considered to have been greatly influenced by wave and storm processes. In particular, storms were a major feature of the hydraulic regime in ancient shelves (Kreisa, 1981; Wright and Walker, 1981; Leckie and Walker, 1982). Marsaglia and Klein (1983) have indicated that the western interior basin of Canada was within a latitude susceptible to frequent winter storms. There are several examples which have documented ancient storm deposits. Leckie (1983) noted hummocky cross-stratification in shoreface deposits of the Lower Cretaceous Gates Formation in northeastern British Columbia. High-angle cross-laminations, in the Wabiskaw 'C' sand of northern Alberta, have been interpreted as hummocky cross-stratification by Ranger *et al.*, (1988). Rahmani and Smith (1988) reported hummocky cross-stratified beds in the shoreface of the Cadotte Formation in the Elmworth area of Alberta.

Recent studies of the effects of storms on modern continental shelves (e.g. Swift and Field, 1981) suggest that storms are primarily resonsible for the bulk transport of sand into the offshore setting. However, it has also been noted that the thickness of these storm layers are in the range of a few centimeters to several decimeters. The layer deposited from Hurricane Carla was less than 10 cm (Hayes, 1967) and layers in the Norton Sound are less than 20 cm (Nelson, 1982b). These storm layers alone seem insufficient to account for the vast quantity of sediment needed for the build-up of extensive offshore bars such as in the Bluesky Formation. Thus, alternative shelf processes are needed to account for the occurrence of large amounts of sand discretely emplaced many kilometers from any paleo-shoreline.

Recent work by Eyles and Walker (1988) in the Cardium Formation of central Alberta have provided an alternative explanation for the extensive linear sandstone accumulations often interpreted as 'offshore bars'. These sandstone accumulations are encased in mudstones and capped by conglomerates whose origin has stirred considerable debate. They have suggested that the sandstones are remnants of a once continuous blanket sand that was shaped into discrete linear sand pods by shoreface erosion during a major transgression. The occurrence of the conglomerates, at the top of the sandstone buildups, are considered to have been provided by coeval coastal input and distributed across the shelf by the transgressing sea (Eyles and Walker, 1988).

4.0 SEDIMENTOLOGY

The sedimentology of the Bluesky Formation within northeastern British Columbia is examined in this chapter. The discussion will focus on detailed descriptions of the lithologies and their depositional settings. The sedimentology will be presented in two parts: (1) detailed descriptions of the lithofacies and (2) interpretation of sedimentary structures and depositional origins of the lithofacies based on the physical bedforms observed.

4.1 LITHOFACIES DESCRIPTIONS

4.1.1 CRITERIA FOR THE RECOGNITION OF LITHOFACIES

The term "lithofacies", as used in this thesis, refers to the lateral and vertical structural and organic aspects of sedimentary rocks (de Raaf, *et al.*, 1965) No connotation in a genetic sense is implied. In the following discussion the dominant lithofacies are described and interpreted as observed in outcrop and core. Core was studied bed by bed to record detailed descriptions of the internal sedimentary structures, thickness, bedding contacts, mineralogy, apparent grain size and biogenic structures. The major criteria for classifying the various beds into a specific lithofacies were the physical sedimentary structures, grain size and the presence or absence of biogenic structures. All measured sections of outcrop, coal company diamond drill core, and petroleum well core are illustrated in Appendix I. Apparent grain size was determined by using the Wentworth grain size scale (Figure 4.1).

The dominant lithologies in the Bluesky Formation are fine- to coarse-grain sandstone, conglomerate, siltstone and mudstone. Eight distinct lithofacies characterize the Bluesky Formation in the study area. For this study, physical and biogenic sedimentary structures have particular

Grain Size Scales

Millimeters		Phi (φ)	Wentworth Size Class	
	4096	-12	Boulder (-8 to -12φ)	GRAVEL
	1024	-10		
	256	-8	Cobble (-6 to -8φ)	
	64	-6	Pebble (-2 to -6φ)	
	16	-4		
	4	-2	Granule	
	3.36	-1.75		
	2.83	-1.5		
	2.38	-1.25		
	2.00	-1.0	Very coarse sand	SAND
	1.68	-0.75		
	1.41	-0.5		
	1.19	-0.25		
	1.00	0.0	Coarse sand	
	0.84	0.25		
	0.71	0.5		
	0.59	0.75		
1/2	0.50	1.0	Medium sand	
	0.42	1.25		
	0.35	1.5		
	0.30	1.75		
1/4	0.25	2.0	Fine sand	
	0.210	2.25		
	0.177	2.5		
	0.149	2.75		
1/8	0.125	3.0	Very fine sand	
	0.105	3.25		
	0.088	3.5		
	0.074	3.75		
1/16	0.0625	4.0	Coarse silt	MUD
	0.053	4.25		
	0.044	4.5		
	0.037	4.75		
1/32	0.031	5.0	Medium silt	
1/64	0.0156	6.0	Fine silt	
1/128	0.0078	7.0	Very fine silt	
1/256	0.0039	8.0	Clay	
	0.0020	9.0		
	0.00098	10.0		
	0.00049	11.0		
	0.00024	12.0		
	0.00012	13.0		
	0.00006	14.0		

Figure 4.1: (after Folk, 1968).

importance. The relative abundance (density) of biogenic reworking was recorded in each of the measured units. The density of trace fossils can range from a few scattered burrows to beds that are completely burrowed. The identification of trace fossils in core is based mainly on their appearance on the outer surface of the core, either split or unsplit surfaces. Criteria which must be considered for the identification of individual ichnogenera are: (1) orientation of the structures, (2) presence of wall lining or ornamentation, (3) distinct geometric pattern, (4) presence of meniscate fill, (5) occurrence of spreite, (6) presence of organic residue, (7) occurrence and type of fill, (8) presence of pellets and, (9) the overall shape and size of the structure (Ekdale *et al.*, 1984).

Generally, drill core provides excellent exposure of traces; however, there are several limitations to the use of core for ichnofossil recognition. Firstly, the lack of three-dimensional exposure limits the lateral interpretation of any changes in the traces or setting which surround the core. Secondly, the density of a trace may be under-represented, especially for vertical burrows since their long axes parallel the core. Finally, different sectional views of a particular trace may resemble that of a completely different ichnogenus, thus, an erroneous identification of an ichnogenus is possible. It should be noted that in some intersections of Bluesky core, severe biogenic reworking has destroyed or obscured all recognizable burrow forms making it difficult, if not impossible, to assign formal ichnotaxa.

Two main types of internal cross-stratification were interpreted based on the association of cross-laminae within the set and the nature of the lower bounding surface. Planar cross-stratification is recognized by the inclined, parallel foresets which maintain a constant angle and parallel bounding surfaces. Trough cross-stratification is characterized by strongly concave upwards laminae with non-parallel, curved bounding surfaces. The latter is recognized only in small- to medium-scale varieties in core. In outcrop, because most of the Bluesky Formation is easily eroded, only the basal conglomerates are generally exposed.

Bedding contacts were described as planar, scoured or gradational. Planar bedding contacts are non-erosional, sharp and form horizontal bed boundaries. Scoured contacts are irregular to wavy, generally with relief greater than 3 cm, and commonly truncate structures in the underlying bed. Gradational contacts record a gradual change from one lithofacies to the next over several centimeters.

4.1.2 SUMMARY OF LITHOFACIES

Eight major lithofacies with four sublithofacies have been outlined for the Bluesky Formation These are:

- Lithofacies 1: Ungraded matrix-supported pebble conglomerate
  - Sublithofacies 1a: Structureless
  - Sublithofacies 1b: Low-angle cross-stratified
- Lithofacies 2: Interbedded mudstone and fine-grain sandstone
- Lithofacies 3: Small-scale cross-stratified medium-grain sandstone
- Lithofacies 4: Massive glauconitic mudstone
- Lithofacies 5: Bioturbated glauconitic sandstone
  - Sublithofacies 5a: Medium- to coarse-grain
  - Sublithofacies 5b: Fine-grain
- Lithofacies 6: Massive to horizontally stratified medium-grain sandstone
- Lithofacies 7: Low- to high-angle cross-stratified medium-grain sandstone
- Lithofacies 8: Silty mudstone

4.1.3 LITHOFACIES 1: UNGRADED MATRIX-SUPPORTED PEBBLE CONGLOMERATE

This lithofacies is comprised of two distinct sublithofacies: a structureless matrix-supported conglomerate and a low-angle cross-stratified matrix-supported conglomerate. This lithofacies is characterized by matrix-supported pebble conglomerates which are generally poorly sorted with a wide range of clast sizes. Clast shapes vary greatly from oblate to equant. Clasts are generally randomly oriented and are composed of mainly chert with minor quartz and quartzite floating in a sand or mud matrix. The sand matrix is typically a medium- to coarse- grain chert and quartz sand (0-2 phi). The clasts are sub-angular to rounded with most being sub-rounded. The largest of t[illegible]bbles average 5 cm (range: 2-15 cm) whereas the average mean pebble diameter is about 1 cm. Most of the chert is white, grey or black with minor amounts of pale green or bluish grey. Clast content is generally less than 40%. In a [illegible]thofacies 1a is composed of a crudely bedded, glauconitic sandstone with minor floating chert pebbles. This [illegible] end-member to the structureless, matrix-supported conglomerate of sublithofacies 1a.

Within sublithofacies 1a, sand matrix-supported conglomerates usually grade upwards into mud-supported conglomerates at the contact with the Moosebar shales. An unique feature of this pebble conglomerate is that although the sand matrix may grade vertically upward into mud, no corresponding decrease in pebble size was noted. This does not occur in sublithofacies 1b. Typically, sublithofacies 1b occurs interbedded with the cross-bedded sandstones of the Peace River plains area. Locally, this matrix-supported conglomerate may be interbedded with zones of clast supported conglomerate.

4.1.4 SUBLITHOFACIES 1A: STRUCTURELESS UNGRADED MATRIX-SUPPORTED PEBBLE CONGLOMERATE

In this sublithofacies structureless, ungraded, matrix-supported pebble conglomerates show little evidence of stratification within the bed (Figure 4.2a). The unit is usually composed of a single, thick veneer of disorganized, poorly sorted pebble conglomerate. Basal contacts are consistently erosive with relief of 2-15 cm (Figure 4.2b). Bioturbation is found exclusively at the contact. A network of sharp-walled, unlined burrows, tentatively identified as *Thalassinoides* extend from the contact into the underlying sediments. The burrows are infilled with pebbles and sand from the overlying conglomerate. Bioturbation is not evident in outcrop due to the friability and poor condition of most outcrops. Rip-up- and coal-intraclasts are commonly found in the lower sections of the conglomerate. Composition of the matrix and clasts are as those described above. Glauconite content averages 5% and occurs as discrete sub-rounded grains. Bed thickness averages 60 cm (range: 10-700 cm). This sublithofacies is commonly found in contact with the underlying Gething Formation or at the top of the Bluesky section grading into Moosebar shales. Between the major sandstone buildups of the Peace River Plains area, this sublithofacies can comprise the complete Bluesky Formation occurrence.

4.1.5 SUBLITHOFACIES 1B: LOW-ANGLE CROSS-STRATIFIED UNGRADED MATRIX-SUPPORTED CONGLOMERATE

This sublithofacies consists of matrix-supported pebble conglomerate with faint low-angle cross-stratification (Figure 4.2c,d), occurring in sets 0.01-2.5 m thick. Although in outcrop this sublithofacies was found only at Willow Creek and Crassier Creek, it occurs commonly in approximately 20% of the petroleum core conglomerate intersections. This unit may vary from a single pebble veneer 1 cm thick to 7.2 m total thickness as found at the Crassier Creek outcrop. Beds, in this instance, are laterally persistent and can be traced for several meters with only slight variations in thickness. Stratification is revealed by laminations which slightly fine upward in pebble or matrix grain size, or by associated lenticles and stringers of planar laminated sandstone. Cross-beds average 0.15 m (range 0.1-4.5 m) in thickness. The cross-laminae generally dip at a low angle (5-15^{0}) which parallel the lower bounding surface. The conglomerate beds are sheet-like units of cobbles, pebbles and coarse sand, and contain rare low-angle gravel filled scours at the top of some beds. Grading and sorting in this sublithofacies is slightly more de-

FIGURE 4.2 CORE PHOTOGRAPHS OF LITHOFACIES

a) **Structureless ungraded matrix-supported pebble conglomerate, sublithofacies 1a:** Crassier Creek outcrop. Pebbles are chaotically scattered throughout with a wide range of clast sizes. There is neither imbrication nor any developed stratification in the outcrop. Up is towards the left of the photograph.

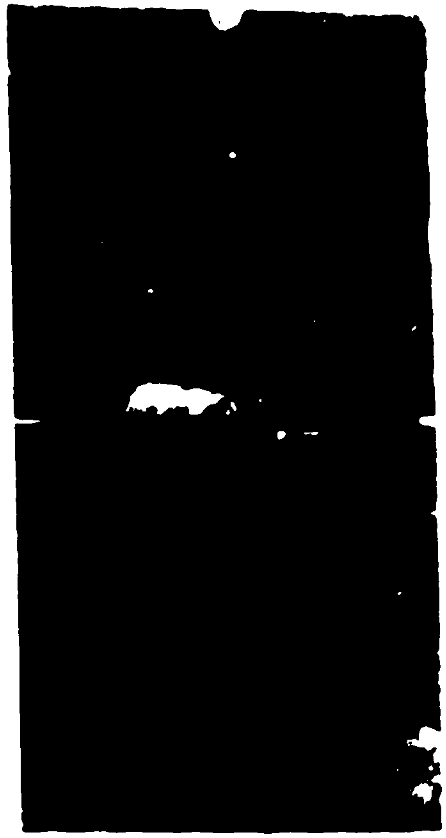
b) **Structureless, ungraded matrix-supported pebble conglomerate, sublithofacies 1a:** Conglomerate in contact with carbonaceous mudstones of the Gething Formation. Contact is scoured with over 15 cm of relief (dashed line). A fragment of the underlying Gething mudstone is found within the conglomerate (hollow arrows). A truncated burrow is noted by a solid arrow. d-67-k 94-H-2, 1046.4 m.

c) **Low-angle cross-stratified ungraded matrix-supported conglomerate, sublithofacies 1b:** Crude stratification in pebble conglomerate is indicated by slight decrease in clast size and crude layering. Packing of clasts is high. b-42-L 94-A-13, 1338.5 m.

d) **Low angle cross-stratified ungraded matrix-supported conglomerate, sublithofacies 1b:** Low angle cross-stratification is indicated by thin lenticles of pebbly sandstone and crude normal grading. d-14-L 94-H-2, 1037.5 m.

NOTE: In all photographs, unless otherwise specified, cores are 8 cm wide. Bottom is the base of the core photograph.

a.



b.

c.

d.

veloped than the above sublithofacies Basal bedding contacts range from planar to scoured with low relief (< 3 cm) Bioturbation is not evident in any of the core intersections or outcrops. Glauconite content averages <2% Composition of the matrix and clasts are as those described above

4 1 6 LITHOFACIES 2 INTERBEDDED MUDSTONE AND FINE GRAIN SANDSTONE

This lithofacies consists of a distinct succession of alternating, thinly bedded, fine-grain sandstone and mudstone couplets in which the frequency and thickness of the sandstone beds increases upward (Figure 4 3a) A typical sandstone-mudstone couplet ranges in thickness from 3 to 20 cm in the lower half of the section, whereas in the upper parts the sandstone beds thicken to several meters ranging from 0 1 to 3 5 m Individual bed thicknesses for the mudstone beds ranges from 0 02 to 3 5 m Total section thickness for this lithofacies averages 10 m (range 2-35 m) The base of the section is dominated by mudstone A sandstone-mudstone couplet consists of a sharp based, cross-laminated sandstone grading upward to bioturbated mudstones. Overall, the sandstone beds are fine-grain, and normally-graded with well-developed, low-angle cross-laminations and rare ripple laminations at the top of the bed Trough cross-laminations are interpreted from small-scale structures in which the cross-laminae dip asymptotically towards the base. The lower bounding surface of the trough cross-bed is also generally concave upwards. Cross-laminae are revealed by grain color variations and fining upward of grain size. The basal contact of the sandstone beds are usually scoured with abundant load structures. Load structures are commonly well defined ball-and-pillow or flame structures. Bioturbation in the sand beds is minor and typically occurs at the top of the bed. Stratification in the mudstones is indicated only by rare, thin silty streaks. Mudstone interbeds show varying degrees of bioturbation. This lithofacies contains the trace fossils *Chondrites*, *Gyrochorte*, *Helminthopsis*, *Planolites*, *Skolithos*, *Teichichnus*, *Zoophycos*, and *Palaeophycus*. The mudstones weather, in outcrop, to rusty rubbly pieces, whereas the sandstones have a platy habit. Body fossils have been identified in this horizon by C.R. Stelck (in Duff and Gilchrist, 1981). A few of those identified are: *Entolium* sp., *Ostrea*?, *Pecten* sp., *Corbula*?, *Entolium irenense*, *Inoceramus dowlingi*, *Tancrodia* sp., *Pleuromya* sp. and *Protocardia* sp. This lithofacies occurs only in the southern half of the study area, within the Pine Pass-Wapiti River region, and is characteristically overlain by sediments of lithofacies 3.

4.1.7 LITHOFACIES 3: SMALL-SCALE CROSS-STRATIFIED MEDIUM-GRAIN SANDSTONE

This lithofacies consists of medium- to coarse-grain sandstone with well-developed, low-angle planar and small-scale trough cross-laminations with rare ripple cross-laminations (Figure 4.3c). Planar laminations are typically low-angle with a slight fining upward from medium- to fine-grain sand. Laminations range from 0.3 to 1 cm in thickness. Sandstones are essentially clean, quartz arenites which are well-cemented and well-sorted. The complete sandstone unit generally grades upward from small-scale cross-stratification at the base to parallel laminated beds near the top. The log signature is characteristically blocky. Commonly, there are thin (2-5 mm) layers of chert pebble/granules and/or zones of angular muddy intraclasts (range: 1-3 cm in length) on the basal section of the parallel laminated beds (Figure 4.3d). Bed thicknesses range from 1 to 12 m, averaging 3 m. Thin stringers of finely macerated carbonaceous matter are also found throughout the unit. This lithofacies is locally bioturbated *(Palaeophycus)* and is characteristically non-glauconitic. The base of this unit is typically a scoured, disconformable surface overlying mudstones of lithofacies 2. The underlying beds are visibly truncated (relief of < 2 cm). Soft sediment deformation, in the form of flame and load structures, was also observed. This unit is overlain by the deltaic deposits of the Chamberlain Formation in the Sukunka-Wapiti Rivers area

FIGURE 4.3 CORE PHOTOGRAPHY OF LITHOFACIES

a) **Interbedded mudstone and fine grain sandstone, lithofacies 2:** Sandstone lenses exhibit normal grading, cross-lamination and sharp, eroded bases. The mudstone beds often exhibit burrows such as *Planolites* (arrows). C-35, (Sukunka), 384.3 m.

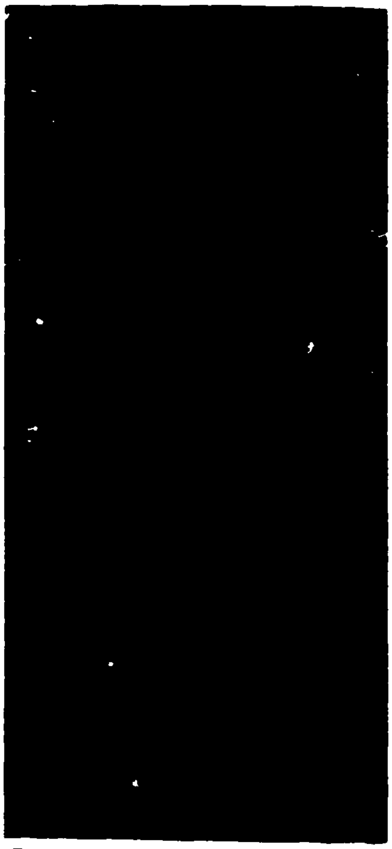
b) **Interbedded mudstone and fine-grain sandstone, lithofacies 2:** Thick beds of cross-bedded sandstones with minor bioturbation commonly occur at the top of this coarsening-upward lithofacies. Low- to high-angle cross-lamination is marked by dashed lines and a possible *Skolithos* burrow is marked by an arrow. C-35, (Sukunka), 378.7 m.

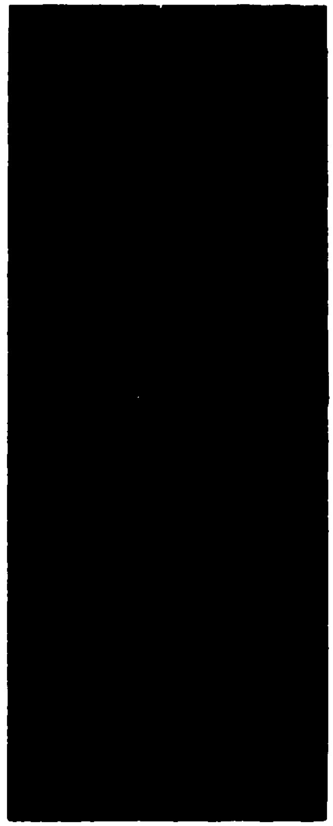
c) **Small-scale cross-stratified medium-grain sandstone, lithofacies 3:** Parallel-laminated clean sandstone with thin layers of chert granules and mud intraclasts. Base of sandstone bed is sharp and eroded C-35, (Sukunka), 305.1 m.

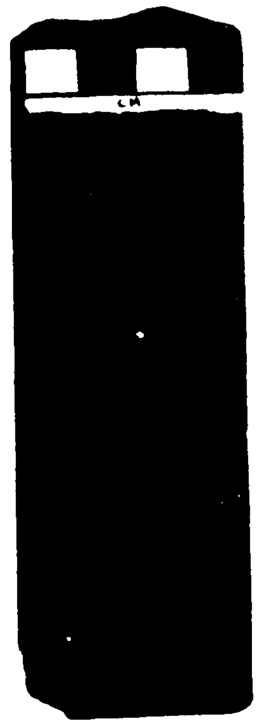
d) **Small-scale cross-stratified medium-grain sandstone, lithofacies 3:** Zone of angular mudstone intraclasts in a cross-bedded sandstone. C-35, (Sukunka), 362.4 m.

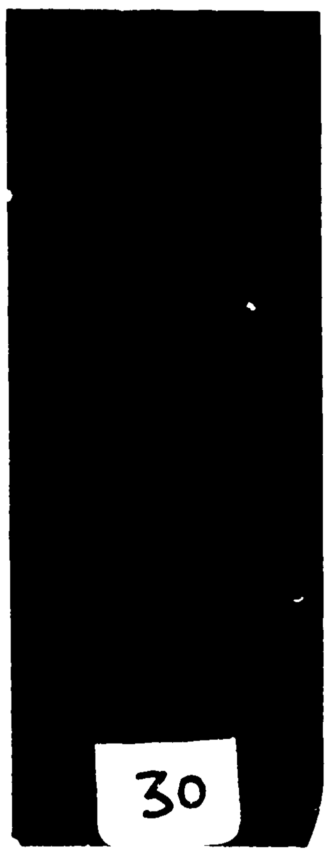
e) **Massive glauconitic mudstone, lithofacies 4:** Outcrop photograph of extremely glauconitic mudstone (G) which exhibits a a dark hue in photographs and a green color in outcrop. Dashed lines mark the top and basal contact of the unit. The unit is 0.9 m wide. The top of the unit is towards the left of the photograph. Sukunka River outcrop.

NOTE: In all photographs, cores are 4 cm wide.


a.


b.


c.



d.

e.

4.1.8 LITHOFACIES 4: MASSIVE GLAUCONITIC MUDSTONE

An extremely glauconitic silty mudstone comprises this lithofacies. Glauconite content varies from scattered grains to about 50 %, where the unit forms a coarse, gritty mudstone. The bed ranges from 0.1 to 2.0 m in thickness and is generally massive to crudely bedded. Stratification, although generally obscure, is indicated by thin sandstone stringers. In some cases, a few scattered floating chert pebbles were found averaging < 1 cm in diameter. The pebbles are random oriented. In outcrop and core, the unit has a distinctive dark green color (Figure 4.3e). Bioturbation was not evident in any of the cores or outcrops.

4.1.9 LITHOFACIES 5: BIOTURBATED GLAUCONITIC SANDSTONE

This lithofacies consists of intensely bioturbated glauconitic sandstones which are divided into two sublithofacies based on the grain size, type of burrows and mud content The sandstones have been greatly modified by biogenic reworking. In general, homogenization by burrowing has completely destroyed any evidence of primary sedimentary structures, giving the sandstones a distinct mottled appearance in core. The lower and upper contacts are gradational over a few decimeters. In some core intersections, the unit is truncated at the top contact by deposits of lithofacies 1 and 6.

4.1.10 SUBLITHOFACIES 5A. MEDIUM- TO COARSE-GRAIN BIOTURBATED GLAUCONITIC SANDSTONE

This sublithofacies consists of medium- to coarse-grain, bioturbated and glauconitic sandstone containing sharp, discontinuous, slightly wavy shaly laminae or mud stringers (Figure 4.4a). The percentage of mud content is generally less than 15%. The bed thickness in this sublithofacies ranges from 0.5 to 8 m. Glauconite content averages 10% (range: 2-20%). Glauconite occurs pervasively as sub-rounded grains or in the lining and fill of the burrows. Physical sedimentary structures are generally completely destroyed by a high density of biogenic structures. Structures, where apparent, consist of rare asymmetric ripple forms. Ichnofossils identified within this sublithofacies consist of dominant vertical and subordinate horizontal burrows (Figure 4.4b). This sublithofacies contains a recurring assemblage of trace fossils consisting of at least 8 ichnogenera, including: *Asterosoma*, *Cylindrichnus*, *Diplocraterion*, *Ophiomorpha*, *Palaeophycus*, *Planolites*, *Skolithos*, and, *Teichichnus*. In addition, numerous indiscernable burrow forms are also present which provide a bioturbate texture. Porosity in this sublithofacies is fair to good (< 10%). This unit generally overlies deposits of sublithofacies 5b.

4.1.11 SUBLITHOFACIES 5B: FINE-GRAIN BIOTURBATED GLAUCONITIC SANDSTONE

This sublithofacies consists of a fine-grain intensely bioturbated, glauconitic sandstone which generally has a higher mud matrix content than sublithofacies 5a The proportion of mud laminae and shale stringers commonly ranges from 10 to 25%. Primary sedimentary structures in this sublithofacies have also been completely destroyed by biogenic reworking (Figure 4.4c). The beds are generally extremely poorly sorted which is probably also a consequence of the intense bioturbation. Preserved mud laminae are commonly wavy, discontinuous and range in thickness from 1 to 3 mm. This sublithofacies is also characterized by a recurring association of ichnofossils. In contrast to the above sublithofacies, this assemblage of traces are dominantly horizontal with abundant grazing and deposit-feeding traces. This assemblage consists of *Asterosoma*, *Chondrites*, *Helminthopsis*, *Palaeophycus*, *Planolites*, *Rosselia*, *Teichichnus*, and *Terebellina*. The bed thickness for this sublithofacies ranges from 0.5 to 6 m. Glauconite content ranges from 5 to 30%. Porosity in the unit is generally fair (< 5%).

FIGURE 4.4 CORE PHOTOGRAPHS OF LITHOFACIES

a) **Medium- to coarse-grained bioturbated glauconitic sandstone, sublithofacies 5a:** Intense bioturbation, dominated by vertical-oriented lined tubes, possibly *Skolithos* (arrows), have destroyed the primary physical structures 6-29-81 15, 10953 m.

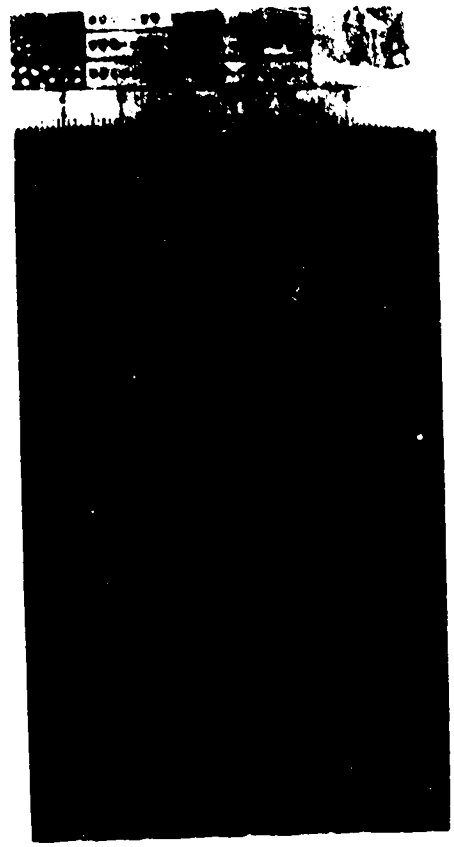
b) **Medium- to coarse-grained bioturbated glauconitic sandstone, sublithofacies 5a:** *Diplocraterion* burrows in a medium-grained, severely bioturbated sandstone. 6-29-81-15, 1094.7 m.

c) **Fine-grained, bioturbated glauconitic sandstone, sublithofacies 5b:** Severely bioturbated shaly sandstone, from the Peace River plains sandstone buildups, exhibiting *Palaeophycus* burrows (arrows) and possibly *Teichichnus* structure (upper left corner of photograph). 7-30-80-14, 1098.2 m.

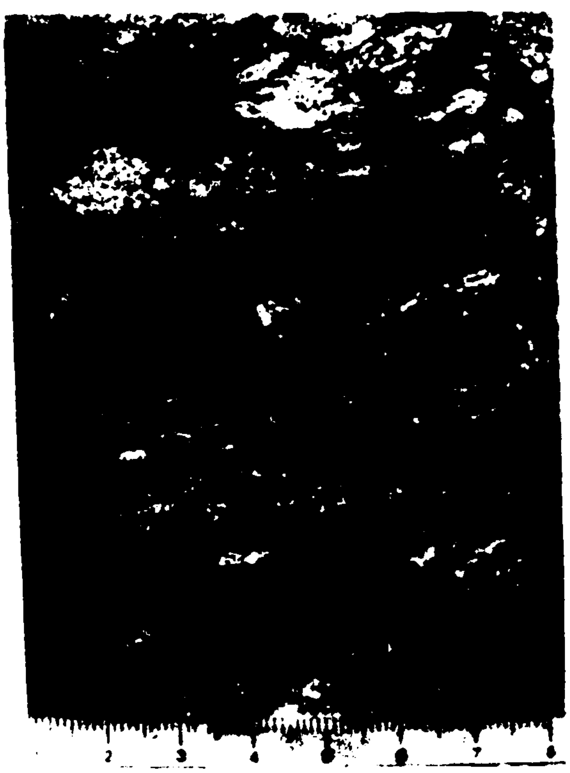
d) **Fine-grained, bioturbated glauconitic sandstone, sublithofacies 5b:** Burrowing is dominated by deposit-feeding and a few dwelling structures such as *Palaeophycus* (arrows). The shift in feeding strategy is typical in the finer grained sandstones of sublithofacies 5b. 6-29-81-15, 1093.3 m.

NOTE: In all core photographs, core is 8 cm wide.

a.



b.



c.

d.

4.1.12 *LITHOFACIES 6: MASSIVE TO HORIZONTALLY-STRATIFIED MEDIUM-GRAIN SANDSTONE*

This lithofacies is characterized by massive to horizontally-stratified, medium-grain sandstones (Figure 4.5a). The laminations when visible are generally planar and parallel and range in thickness from 2 to 7 mm. Laminations are defined by the variation in color of adjacent laminae and slight fining-upward of grain size. Low-angle cross-stratification ($< 5^{o}$) was also recorded in some cores. The bedding contacts are sharp, planar to slightly undulose and possibly erosive. Average bed thickness is 0.3 m (range: 0.01 to 1.2 m). The sandstones are generally well sorted, clean and friable. The grains are commonly angular, composed of clear quartz with white kaolinite booklets occupying the pore spaces. The sandstone has a tan-brown, micaceous appearance in core. Bioturbation is virtually absent throughout most of the bed except a few minor traces (i.e. *Skolithos* and *Palaeophycus*) at the top. Glauconite occurs as scattered, sub-rounded grains and generally is less than 1% in content. Porosity in this lithofacies is generally good to excellent (10-20%). This sandstone is commonly underlain by a thin band of flat-lying, sub-rounded chert pebbles in a coarse-grain sand matrix (Figure 4.5b). The pebble bands average 5 cm in thickness and have a sharp, scoured base.

4.1.13 *LITHOFACIES 7: LOW- TO HIGH-ANGLE CROSS-STRATIFIED MEDIUM- TO COARSE-GRAIN SANDSTONE*

This lithofacies consists of low- to high-angle ($5\text{-}20^{o}$) cross-stratified medium- to coarse-grain sandstone (Figure 4.5c). Commonly, the bedsets contain numerous small, well rounded chert pebbles (5-20 mm in diameter), either as isolated clasts dispersed in the bed, or as discrete layers along distinct bedding planes. This lithofacies is characterized by planar laminations and slightly concave upward curved cross-laminae which are interpreted as trough cross-stratification (Figure 4.5d). Occasionally, low-angle planar cross-stratification displays upward shallowing oppositely-dipping cross-laminae. A distinguishing feature of the slightly concave-upward cross-laminations is their basal low-relief truncation of the previous bedset. Individual cross-lamina are distinguishable by their variation in grain color from adjacent laminations and by a slight fining-up in grain size. Most commonly the sandstones are moderately- to well-sorted and a light-buff-grey with an overall 'salt and pepper' appearance. Bed sets range in thickness from 0.5 cm to 2.5 cm. Glauconite content within these sandstones generally averages < 5 %, which is commonly less than that found in the bioturbated sandstones of lithofacies 5. Bioturbation structures are virtually absent or not apparent. In one occurrence (a-89-K 94-H-6; 1088.4 m) a single *Ophiomorpha* burrow was noted in a cross-bedded sandstone. Reactivation surfaces are not uncommon. Basal bedding contacts are generally sharp, planar and non-erosive. Unit thicknesses average 2 m (range: 0.5 to 9.5 m).

4.1.14 LITHOFACIES 8: SILTY MUDSTONE

This lithofacies consists of dense, dark grey to black horizontally-stratified mudstone and silty mudstone (Figure 4.6a,b). The lithofacies is characteristically bioturbated with burrows of *Planolites*, *Chondrites* and *Helminthopsis*. Extensive bioturbation completely homogenizes some parts of the unit to where no distinct burrow forms or any primary sedimentary structures are discernable. Also recorded were siderite bands ranging from 10-20 cm in thickness and providing the only recognizable stratification. Bedding contacts range from sharp planar to scoured and erosive. Bed thicknesses range from 3 cm to 30 cm. This lithofacies is generally interbedded between two coarse-grain sandstones of lithofacies 5.

FIGURE 4.5 CORE PHOTOGRAPHS OF LITHOFACIES

a) **Massive to horizontally-stratified medium-grained sandstone, lithofacies 6:** *Skolithos* (arrow) burrow in a massive-appearing sandstone. The sandstone bed has a sharp basal contact over a chert pebble conglomerate band. C-16-D 94-H-10, 3313 m.

b) **Massive to horizontally-stratified medium-grained sandstone, lithofacies 6:** Massive sandstone (m) with a basal conglomerate log truncates an extremely bioturbated sandstone layer (b). d-33-C 94-H-11, 1109.7 m.

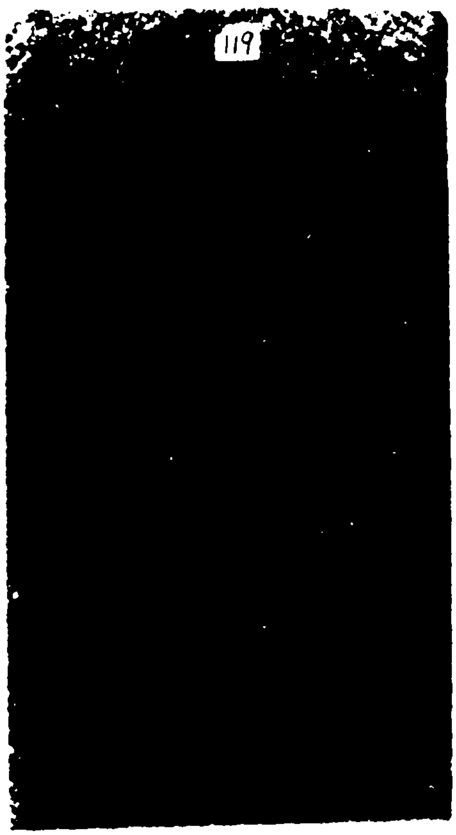
c) **Low- to high-angle cross-stratified medium sandstone, lithofacies 7:** Non-burrowed, high-angle, cross laminations are interpreted to be trough cross-stratification. b-42-L 94-A-13, 1338.5 m.

d) **Low- to high-angle cross-stratified medium-grain sandstone, lithofacies 7:** High-angle cross-laminated sandstone with abundant chert granules and pebbles along bedding planes. d-67-K 94-H-2, 1033.6 m.

NOTE: In all core photographs, cores are 8 cm wide

a.

b.



c.

d.

FIGURE 4.6 CORE PHOTOGRAPHS OF LITHOFACIES

a) **Silty Mudstone, Lithofacies 8:** Dark grey mudstone with thin interbeds of laminated and burrowed silt overlying a chert pebble conglomerate of sublithofacies. b-42-L 94-A-13, 1336.0 m.

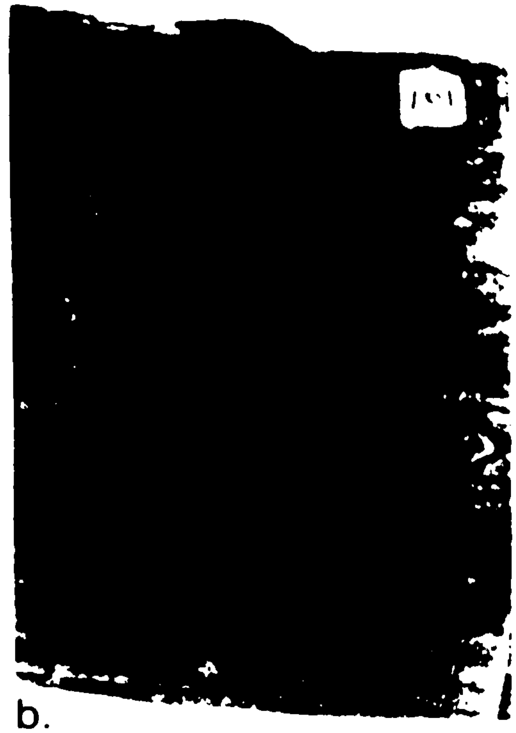
b) **Silty Mudstone, Lithofacies 8:** Bioturbated mudstone with silt laminations and silt-filled *Planolites* burrows (arrows) d-67-K 94-H-2, 1046.3 m.

c) Bioturbated medium-grained sandstones of sublithofacies 5a overlying massive to horizontally-stratified medium-grained sandstones providing a laminated-to-burrowed appearance in core. Arrow points to a large *Skolithos* burrow. Top of core is at the upper right corner of photograph. Core is 5 cm wide. PD 8001, 535 m.

d) **Low- to high-angle cross-stratified medium-grain sandstone, lithofacies 7:** High-angle cross-laminations are interpreted to be trough cross-stratification. d-59-L 94-A-13, 1290.21 m.

NOTE: Core in all photographs, except where noted, is 8 cm wide.

a.



b.

c.

d.

4.2 LITHOFACIES INTERPRETATIONS

4.2.1 SUBLITHOFACIES 1a: STRUCTURELESS UNGRADED MATRIX-SUPPORTED PEBBLE CONGLOMERATE

Sublithofacies 1a which consists of a poorly sorted, ungraded, matrix-supported chert pebble conglomerate containing glauconite, mud and coal intra-clasts is interpreted to be a transgressive lag deposit formed during the initial transgression of the Moosebar Sea. The rapid input of coarse-grain detritus is considered to be related to a major transgression and/or a substantial shift in slope gradient due to tectonic readjustment within the depositional basin. These deposits commonly separate carbonaceous non-marine strata below from bioturbated marine sediments above.

The lag deposits are product results from surf-winnowing processes of shoaling waves as the sea advanced over relict deposits. Oscillatory-flow currents produced by shoaling waves which contain strong asymmetric orbital velocities tend to drive relatively coarse sediment onshore (Komar, 1976). From Figure 4.7, it can be seen that current velocities >40 cm/s are required before grain movement begins in 2.0 mm sand. Therefore, it is considered that exceedingly large clasts, that were not capable of transport, were concentrated on the erosional surface. The paucity of stratification indicates that finer-grain sediments may have been transported and deposited rapidly from suspension as the energy levels waned rather than by bedload tractional transport. With continued transgression, as the water deepens and the shoreline retreats, the effect of waves will diminish and no longer be able to modify the texture of the seafloor sediment. Generally at this point, only fine-grain size sediment was available for deposition in the lag. This may account for the fining upward matrix size commonly observed in the fabric of this lithofacies.

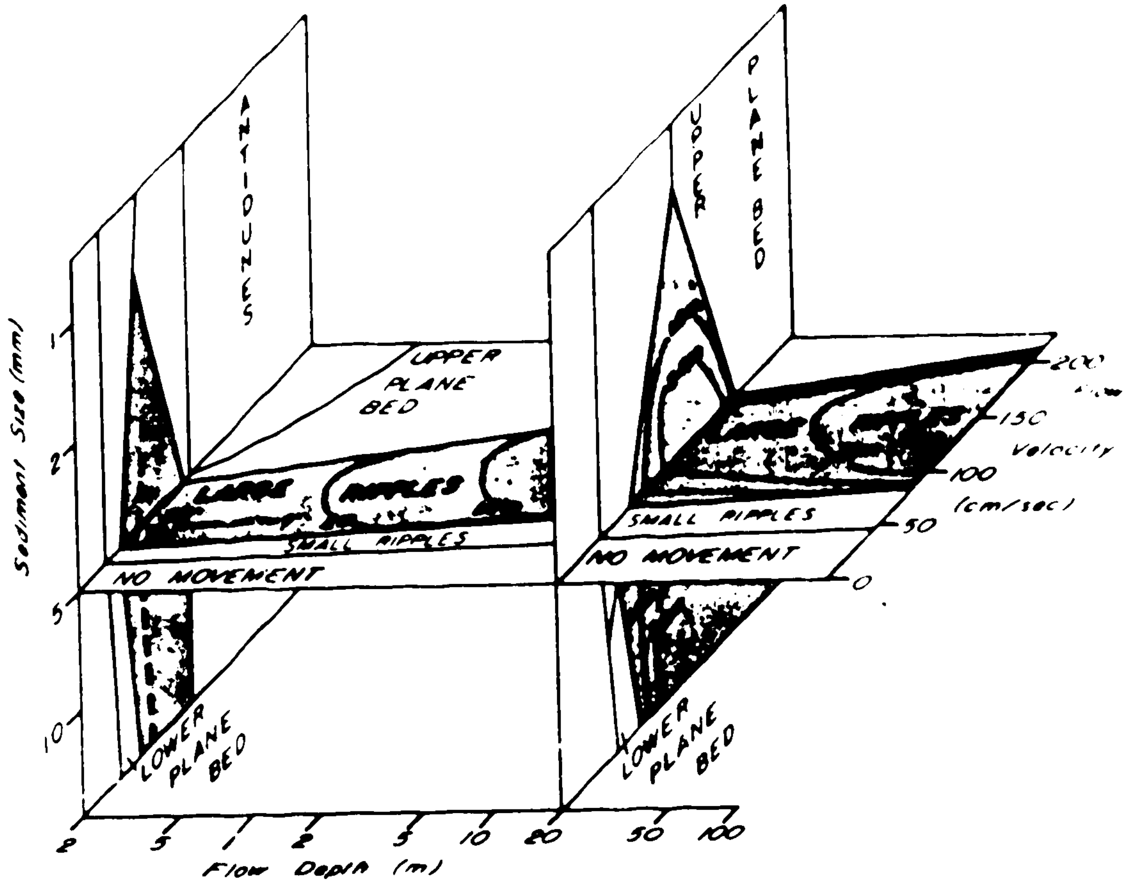


Figure 4.7 Generalized bedform stability plot illustrating relationships between depth, velocity and grain size in unidirectional currents. Positions of boundaries and nature of junctions are approxiamate (modified from Harms et al., 1982)

It is generally considered that transgressive lags represent all that remains from the erosion of underlying sediments (Pugh, 1960) or the remnants of a transgressive shoreline (Sanders and Kumar, 1975). According to Sanders and Kumar (1975) shoreface retreat, which occurs with a relatively slow and continuous rise in sea level, inevitably leads to complete destruction of any nearshore deposits by erosion and wave reworking. The preserved record would only consist of a thin transgressive lag overlain by nearshore marine sediments (Sanders and Kumar, 1975). At this point, it is difficullt to interpret exactly where the sediments for the lag deposits in the Bluesky Formation originated. It has been noted though, that the 'lag' deposits, of the Bluesky Formation, contain material commonly found in the underlying Gething Formation.

4.2.2 SUBLITHOFACIES 1b: LOW-ANGLE STRATIFIED UNGRADED MATRIX-SUPPORTED CONGLOMERATE

The internal stratification of sublithofacies 1b, consisting of low-angle, cross-stratified, matrix-supported conglomerate, indicates that this sediment was tractionally reworked, possibly after the initial depositional event. Intuitively, it would appear that at least moderate to high energy oscillatory currents were responsible for the deposition of this lithofacies. The lateral persistence of these cross-stratified conglomerate beds indicates that the mechanism for deposition acted over a broad area and was not localized at a single locale (e.g. fluvial channel).

Although, the type of current required to produce planar laminations in this sublithofacies is difficult to interpret, long period swells can produce the necessary velocity asymmetry required to generate these bedforms. Clifton (1981) interpreted that cross-stratified pebbly sandstones of the Miocene Cediente Range were produced by large waves with periods of 8 to 12 seconds. Bourgeois and Leithold (1984) suggested that reworked conglomerate beds in the sandstone of Floras Lake are products of long-perod swells forming low-amplitude, migrating nearshore bars.

These nearshore bars are composed of low-angle, cross-stratified conglomerates. Leckie and Walker (1982) also described stratified conglomerate layers from the Cretaceous Moosebar - Lower Gates Formations. They suggested that these conglomerates may have formed by long period oscillatory waves within the shoreface, in depths ranging from 30-100 m. Leckie (1988) has suggested that coarse-grain ripples are common features within broad, open transgressive shelves, where suitable coarse sediment is available. Coarse-grain ripples are considered to be generated by 6-16 second period oscillatory waves within water depths ranging from 3-160 m. Leckie (1988) has documented several occurrences of ancient and modern coarse-grain ripples on transgressive surfaces and shoreface environments.

Thus, it seems possible that these cross-stratified beds have been the result of reworking during higher energy and/or long period oscillatory wave currents which postdate the initial depositional event. The dearth of biogenic structures in this sublithofacies indicates either an extremely high energy setting or environmental conditions to stressful for benthic organisms. The low-angle cross-stratification may be the result of large, low-amplitude migrating bedforms.

4.2.3 LITHOFACIES 2: INTERBEDDED MUDSTONE AND FINE-GRAIN SANDSTONE

This lithofacies which consists of alternating layers of mudstones and fine-grain cross- and ripple-laminated sandstones represent a depositional setting that was subjected to fluctuating energy conditions. The interbeds of mudstone are interpreted as suspension fall-out of mud particulates during hydraulically quiet times. This represents the 'normal' state of sedimentation in a low-energy tranquil muddy substrate environment. Conversely, the fine grain sandstones are considered to be associated with episodic higher energy currents generated possibly during meteorlogical disturbances and carrying coarser detritus. These coarser beds, in most cases, exhibit planar erosive contacts,normal grading and waning flow-sequences of parallel-laminated sands passing upward into starved ripples. Trough cross-stratification commonly replaces the parallel lamination at the base of a sandstone bed. The sedimentary structures in these beds reflect high rates of sediment transport in suspension-laden currents and deposition by traction under progressively decreasing current velocities.

The fine grain size and well-sorted nature of the parallel laminations reflects that these were deposited by upper-flow regime upper plane-bed transport; plane beds of the lower flow regime are not stable within this size range. Figure 4.7 displays the stability fields for bedforms produced by varying flow velocities. For 0.2 mm sand, stable bedforms with increasing flow velocities are: no movement, small ripples, dunes, upper plane bed, and antidunes. Thus, current velocities of about 80 to 120 cm/s are needed to produce planar laminations. Trough cross-stratification which is produced by the migration of megaripples or dunes are formed in current velocities from 60 to 80 cm/s. Current velocities of 15 to 60 cm/s are required to generate ripples in 0.15 mm sand. However, it is possible that some of the cross-laminated interbeds are formed or altered by current or wave reworking subsequent to the initial depositional event. Many of of asymmetric ripples and starved ripples could have been reworked by shoaling waves or a tidal current system.

Several processes, mostly attributed to storms, have been suggested to be responsible for these episodic depositional events in shallow coastal areas, including: density gravity flows from storm-surge ebb (Hayes, 1967), wind-induced geostrophic bottom currents (Morton, 1981), cyclic wave loading during storm events (Nelson, 1982) and density (turbidity) currents (Wright and Walker, 1981). However, it would be difficult to discern exactly which process is responsible for the deposition of this lithofacies as each process is capable of producing similar sequences.

The interbedded nature of this lithofacies is considered to reflect the interaction of storm- and fair-weather processes. The sandstone interbeds, which consist of bedforms produced under waning flow velocities, suggest that these were generated by frequent episodic meteorological disturbances. The mudstone interbeds, on the other hand, represent a return to the 'normal' sedimentological and biological patterns of suspension fall-out in a tranquil, low-energy environment. Typically, in a beach to offshore profile, sediments such as these would be located in the offshore zone above storm wave base.

4.2.4 LITHOFACIES 3: SMALL-SCALE CROSS-STRATIFIED MEDIUM-GRAIN SANDSTONE

The sequence of sedimentary structures in this lithofacies, including small-scale cross-stratification grading upwards to planar laminations, indicates increasing flow velocities with time. The small-scale cross-stratification is interpreted to be the result of migrating small-scale ripple forms. Low-angle planar laminations are considered to be a product of higher energy plane bed deposition.

The occurrence of small-scale cross-lamination in moderately well-sorted fine sands indicates that currents at the time of deposition were in lower part of the lower flow regime. From Figure 4.7, the range of current velocities required to produce small ripples in fine sand (0.2 mm) is from about 20 to 60 cm/sec. The tractional movement of sand producing flow-transverse ripples are preserved only as lee-side laminae. Planar laminations in the slightly coarser, well-sorted sands at the top of the succession were probably deposited from upper flow regime currents. Current velocities producing upper plane beds range from 90 to 130 cm/sec for 0.3 mm sand. The organic debris found in several beds was transported into the site of deposition. It is also possible that planar laminations could have resulted from reworking by high energy oscillatory waves. The change in sedimentary structures vertically, from small-scale cross-laminated beds to planar-laminated beds, indicates an increasing current velocity and/or a shallowing of water depths upsection. The few scattered burrows within this lithofacies indicate that deposition occurred in either a physically demanding or ecologically-stressful setting such as that adjacent to a point source of fresh water. The combination of sedimentary structures in this lithofacies, its association with brackish or fresh water, and its stratigraphic position, underlying the Chamberlain delta sediments, suggest that these sediments were influenced by both unidirectional and oscillatory wave currents. Depositional settings which record the effect of both types of currents are commonly located in the subaqueous zone of the delta front (Coleman and Gagliano, 1965).

4.2.5 LITHOFACIES 4: MASSIVE GLAUCONITIC MUDSTONE

The extremely high content of glauconite in this lithofacies poses an interesting problem of origin. The specific environment of formation which is favorable for the development of glauconite minerals can be ascertained from their occurrences on the present continental shelves. Glauconite is generally not a reliable depth indicator as it occurs on recent shelves at depths from 60 to 500 m (Odin and Matter, 1981). Glauconite has also been recovered at depths exceeding 800 m from submarine highs (Odin and Matter, 1981).

Glauconite most commonly occurs in semi-indurated sandy-muds and muds which underlie the present shelves. These sediments typically have extremely low organic contents and are subjected to low rates of deposition (Bornhold and Griesse, 1985).

The formation of glauconite on shelves is the result of relatively long, quiet conditions at the sediment water interface (Odin and Matter, 1981). The turbulent waters of the shoreface are generally unfavorable to the development of glauconite because the high rate of detrital input does not provide the necessary residence time for glauconite formation before burial (Odin and Matter, 1981). At depths greater than 50 m, into the offshore areas towards the shelf edge, energy levels are reduced and even though substrates are continually shifting and winnowing, the rate of new detrital influx is low (Odin and Matter, 1981). Therefore, within this zone sediments accomodate the formation of glauconite. Thus, it is considered that the development of glauconite is characteristic of open-marine environments, when the requisite parameters are met beyond the zone of high detrital input (Odin and Matter, 1981).

The chemical development of glauconite occurs at the sediment water interface in the transition zone between the strongly oxidizing conditions above and the reducing conditions below (Bornhold and Giresse, 1985). Iron, which is mobilized at depth under reducing conditions, migrates upwards into increasingly oxidizing and acidic conditions closer to the sediment water interface allowing the precipitation of glauconite within the pore space (Bornhold and Griesse, 1985). Glauconization of fecal pellets is also well developed even though they have a high initial organic content. Microbial activity is apparently very intense, thereby reducing the content of carbon to lower levels than the surrounding matrix allowing glauconization (Bornhold and Giresse, 1985).

Marine transgressions generally provide favorable conditions for the development of glauconite on a regional scale (Odin and Matter, 1981). Nearshore clastics which are stranded upon transgression are highly susceptible to glauconite formation because of the lower energy conditions and low input of sediment as the shoreline retreats (Odin and Matter, 1981).

Thus, the occurrence of silty mudstones and the high content of glauconite suggests that this lithofacies was deposited in a relatively quiet-water marine environment, associated with extremely slow rates of sediment accumulation. It is quite possible that this lithofacies was produced as the result of a significant increase in water depths.

4.2.6 SUBLITHOFACIES 5a: MEDIUM- TO COARSE-GRAIN BIOTURBATED GLAUCONITIC SANDSTONE

This sublithofacies, which consists of poorly sorted, muddy, medium- to coarse-grain sands, are interpreted to be the result of moderately strong tractional bottom currents. Although, the intense degree of bioturbation makes it difficult to reconstruct the bedform patterns or their kinematics of formation, the overall characteristic of the sediments allows some inference of th physical processes involved. The occurrence of asymmetric ripple forms suggests that oscillatory flow currents were affecting the loose granular bedload. The absence of many of the tidally-suggestive structures, such as reactivation surfaces, bidirectional cross-stratification, and mud drapes along foresets, suggests that tidal influence was probably not as significant as wind- and wave-induced currents.

According to Clifton *et al.*, (1971) wave-formed ripples are somewhat analagous to ripples formed by unidirectional currents, in that both are typical of lower flow power. Asymmetric ripples formed in 10 second period waves require current velocities from 20 to 80 cm/s to form in 0.35 mm sand (Figure 4.8).

The occurrence of a community of infaunal suspension-feeders which require an agitated water column to suspend food also supports the suggestion that moderately-powered currents were probably responsible for depositing this sublithofacies. Although deposit-feeding structures are present, they are not as abundant.

It has been noted that both physical and biogenic structures such as these are common, but not exclusive, to the shoreface environment (Howard and Reineck, 1981; Eckdale *et al.*, 1984). The dominance of ichnogenera such as *Skolithos* and *Palaeophycus* in a sandy substrate indi-

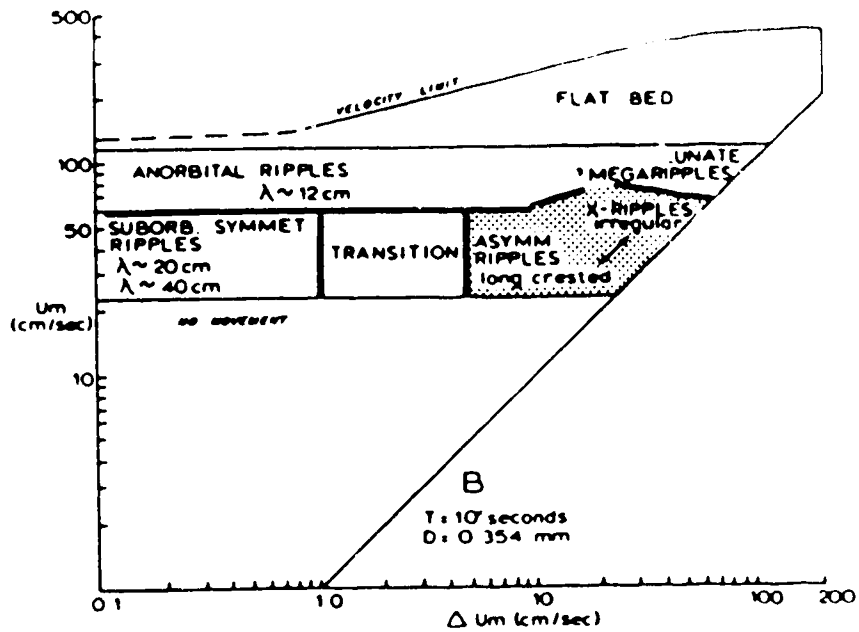


Figure 4.8: Bedform stability plot for 0 35 mm diameter sand in 5-10 ... ...riod oscillatory waves (modified from Clifton, 1976). Asymmetric ripples will for... ...velocities approxiamately from 20 to 80 cm/s as indicated in the shaded area.

cates a high energy marine environment, such as the shoreface, that is suitable for dwelling and suspension-feeding organisms. However, the presence of deposit-feeding forms, such as *Asterosoma*, in relatively clean sands may indicate an offshore origin. It is possible, then, that this sublithofacies was deposited by the aggradation of a sandbed in an offshore setting which was affected by moderately-powered oscillatory wave-currents. The abundance of bioturbation and the paucity of physical sedimentary structures indicates fairly stable substrates and energy levels which were not too physically-demanding to be inhospitable to or destructive to burrowing activity. Current levels must be sufficient, though, to transport medium- to coarse-grain material into the depositional site

4.2.7 SUBLITHOFACIES 5b: FINE-GRAIN BIOTURBATED GLAUCONITIC SANDSTONE

This sublithofacies, which consists of poorly-sorted, muddy bioturbated, fine-grain sandstones, represents a similar setting to that of sublithofacies 5a, but with an overall decrease in energy levels. The wavy and undulating laminations of shaly silt interbedded with fine sands also indicate somewhat fluctuating energy conditions. Muddy silts can accumulate from suspension fall-out in a low-energy environment, whereas the fine sands can be emplaced by slightly higher energy bottom currents. Such periods of differing energy conditions can occur during temporary rough or storm weather conditions. Alternatively, sand with mud drapes can be deposited from the changing high- and low-energy conditions during a diurnal tidal cycle. However, because of the severity of bioturbation most, if not all, physical sedimentary structures in the sand are destroyed making the interpretation of the effects of waves or tides difficult. The suite of trace fossils in this sublithofacies consists predominantly of deposit-feeding structures and occasional dwelling tubes. This assemblage records a fundamental shift in behavior from predominantly suspension-feeding organisms, in sublithofacies 5a, to deposit-feeding organisms in this unit. This is consistent with the finer-grain nature of this sublithofacies.

The argillaceous and fine-grain nature of this sandstone and the suite of predominantly deposit-feeding structures indicates that this sublithofacies was deposited under conditions of low wave activity. The intensity of bioturbation in the sediment reflects stable conditions under which the resident infauna completely reworks the substrate. The current levels must be low enough to support deposit-feeding infauna, but sufficiently high to sustain the few suspension-feeding organisms. Collectively, the physical evidence of this sediment indicates a low energy offshore marine setting slightly more distal than sublithofacies 5a.

The sandstones of this sublithofacies were deposited by traction and the clay laminae were probably deposited from suspension under decreasing wave intensity. Theso alternating traction and suspension deposition may represent temporary storm weather conditions followed by reinstatement of fairweather conditions.

4.2.8 LITHOFACIES 6: MASSIVE TO HORIZONTALLY STRATIFIED MEDIUM-GRAIN SANDSTONE

The relatively non-bioturbated and coarse nature of this lithofacies suggests that the horizontally-laminated sands are the result of rapid deposition in a decelerating strong unidirectional bottom current. The deposition of this lithofacies was controlled by offshore currents of high flow velocities (upper flow regime) which formed upper plane beds and very low amplitude bedforms. High velocity flows are often jet-like and down-welling in nature, and are characterized by a down current decrease in sand flux leading to rapid deposition of sand from suspension (Swift *et al.*, 1983). Figure 4.7 illustrates that the current strength required to produce upper plane beds in 0.4 mm sand is approximately 120 to 140 cm/s. Under these hydraulic and grain size parameters, upper plane beds are the only stable bedform developed.

Of the many sedimentary processes, flows which are able to entrain large volumes of sand in suspension and capable of high velocities are commonly attributed to the action of meteorologically-induced storm-wave surges. Generally, hummocky cross-stratification is considered to be diagnostic of shelf storm deposits (Harms *et al.*, 1975; Leckie and Walker, 1982; Swift *et al.* 1983). However, the geometry of the internal stratification of this lithofacies lacks the undulations and/or curved variable dip laminations and erosional upper contacts which characterizes hummocky cross-stratification (Harms *et al.*, 1982, Duke, 1985).

Storm sand layers in modern sedimentary environments, which contain parallel laminations, have been noted in the Gulf of Mex[illegible] (Hayes, 1967), in the German Bight of the eastern North Sea (Aigner and Reineck, 1982), [illegible] California coast (Howard and Reineck, 1981), Sapelo Island along the Georgia Coast (Howard and Reineck, 1972), and also, the Norton Sound of the Bering Sea (Nelson, 1982b). Storm sand layers in these examples typically show lateral and vertical trends in sedimentary structures as a result of decreasing energy conditions and distance from shore (Johnson and Baldwin, 1986). Generally, the coarsest and thickest sequences of parallel laminations are found in the proximal parts of these sand layers. Following these storm-surge depositional events, normal fairweather processes return, including bioturbation, which may account for the obscuring of ripple lamination at the top of the storm bed.

Although, in this instance, storm-surge ebb currents are the most plausible explanation for this lithofacies, planar stratification can be generated from several current conditions. Harms *et al.*, (1982) state that planar lamination can be produced by traction from a strong, steady upper flow regime unidirectional current such as those found in fluvial settings. Anotner possible hydrodynamic situation responsible for producing planar lamination is high velocity oscillatory flow. These multi-orbital velocities are sufficiently large (Clifton, 1976). These bedforms are usually found in the foreshore and upper shoreface environment in a typical beach to offshore profile

(Clifton *et al.*, 1971). However, this lithofacies lacks the discrete wedge sets and erosional surfaces that characterize the parallel laminations found in these settings. In addition, this lithofacies does not contain any indication of emergent conditions such as rootlets. The massive appearance commonly seen in this lithofacies may be the result of rapid deposition from suspension with no subsequent tractional transport or a uniformity of grain size (Reineck and Singh, 1980).

It is considered that bottom-returning storm-surge ebb currents are responsible for producing beds of this lithofacies. Evidence for this is supported by the interbedded nature of this lithofacies with bioturbated sandstones of lithofacies 6, providing a laminated to burrowed appearance (Figure 4.6c), inclusion o escape burrows, and the occurrence of thin pebble bands at the base of the parallel-laminated sands (Figure 4.5a,b). Laminated-to-burrowed sequences have been observed in ancient sequences from the Star Point and Blackhawk Formations of Utah (Howard and Frey, 1984).

4.2.9 LITHOFACIES 7: LOW- TO HIGH-ANGLE CROSS-STRATIFIED MEDIUM- TO COARSE-GRAIN SANDSTONES

The internal stratification of this lithofacies reveals a rich assemblage of primary structures consisting of low- to high-angle planar- and trough-cross-stratification. The low-angle planar cross-lamination usually occurs as parallel sets of laminations dipping at shallow angles. The medium- to high-angle cross-bedding has been interpreted as trough cross-stratification. These bedforms suggest that they were generated under conditions of continuous and moderately-powered asymmetrical oscillatory flow. The strongest onshore- and offshore-directed water motions in the nearshore are the oscillatory motions due to waves. If the waves are shoaling, the onshore pulses reach higher maximum velocities but are of shorter duration than the offshore pulses (Clifton, 1976). The foreshore and shoreface of the nearshore are dynamic zones dominated by shore-normal oscillatory motion related to primary incident waves (Clifton, 1976).

Bedforms produced by shoaling and breaking waves consist of low-angle, oppositely-dipping wedge-shaped sets bounded by low-angle truncation surfaces (Thompson, 1937; Clifton *et al.*, 1971). Planar-laminations are commonly generated by sheet flow over a flat bed in swash-backwash processes in the foreshore zone.

Seaward of the zone containing planar-laminations is typically a zone of medium-scale trough cross-stratification. These structures are produced by the migration of lunate megaripples (Clifton, 1976). Trough cross-stratification is common in medium- to coarse-grain sands under conditions of intense asymmetric flow generated by long period waves. Dunes or megaripples are large scale features, with spacings of several meters to tens of meters and heights of a few decimeters (Harms *et al.*, 1982). They are three-dimensional forms with a characteristic lunate shape (Harms *et al.*, 1982). They are similar, in profile, to asymmetric ripples produced under unidirectional flow. According to Clifton (1976) lunate megaripples will not form in sands finer than 0.2 mm (fine sand). Based on flume work, Harms *et al.*, (1982) concluded that with increasing flow velocities, the height and spacing of megaripples increase until the transition to upper plane bed flow. The stability field plot of megaripples shows that current velocities of about 55 to 120 cm sec are needed to form these bedforms in 0.5 mm sand (Figure 4.7). Clifton (1976) indicates that the foresets will be oriented shore- parallel or in the direction of a dominant longshore current.

It is considered that the dearth of trace fossils in this lithofacies is the result of the physically-demanding processes limiting the occupation of the substrate by benthic organisms. However, it is possible that the traces may have been destroyed by reworking of the sediment. Similar sequences of physical structures were noted in the foreshore and upper shoreface zone on the California coast (Howard and Reineck, 1972). The occurrence of stacked trough cross-stratified sets represents a laterally migrating megaripple field, probably formed by relatively strong unidirectional bottom currents (longshore drift) driven by wave action and/or semi-permanent currents. This lithofacies is considered to have been deposited in a subtidal high energy marine setting dominated by moderately-powered asymmetrical oscillatory currents. Although occasional planar-laminated beds were encountered, the lack of rooted horizons or paleosols indicates non-emergent conditions of the preserved sediments.

4.2.10 *LITHOFACIES 8: SILTY MUDSTONE*

Unstratified silty mudstones which comprise this lithofacies are generally regarded to represent deposition by suspension fall-out during periods of diminished energies or current quiescence. The sediment is too fine for current-produced structures to have developed, and is thought to have been deposited by settling in slow-moving (<0.1 m/s) bottom currents. Continuous sedimentation over a period of time from hemipelagic fall-out results in the vertical accretion of massive silts and clays. Once deposited, these sediments do not suffer any further transport and thus, form a muddy substrate. Vertical settling of the finest particles under these conditions is extremely slow, and deposition rates range from 0.06 to 6.0 cm per annum (Johnson and Baldwin, 1986). Fine-grain sands and silts were probably introduced during storm weather conditions by storm-surge ebb flow currents into the normally tranquil environment.

The lack of original layering in the mudstones is due to the extensive bioturbation by indigenous organisms. The presence of *Helminthopsis* burrows in mudstones indicates that this lithofacies was deposited in an offshore marine setting.

Mudstones in ancient sequences are considered to have accumulated in the offshore area of epicontinental seas below storm wave-base. However, in Bluesky cores, this mudstone is most often interbedded between bioturbated or laminated sandstones which are not typical of very low energies. This association and the mode of deposition are somewhat problematical. Rather than attribute the sequence of sediments to a transgressive-regressive cy of nearshore clastics and offshore deeper-water muds, it would seem more plausible to relate this series to a temporal shift in depositional styles. The deposition of mudstones can result either during a temporary shortage of coarser sediment, a change in the hydraulic parameters or in the troughal areas between major accumulations of sandstone. Alternatively, during tidal cycles ud drapes can accumulate during the slack water periods. However, these mud drapes are generally thin and cannot account for the relatively thicker (commonly >10 cm) silty mudstone deposits found in the Bluesky cores.

4.3 FACIES SUCCESSIONS

To thoroughly describe the vertical lithofacies succession with the Bluesky Formation, two generalized facies successions are required, one for the Rocky Mountain Foothills area and, the second, for the subsurface in the Peace River Plains area. Figures 4.9 and 4.10 illustrate the typical vertical relationships of the lithofacies associations for each area. Although these represent a typical vertical succession, they do not reflect the variability in the deposits. For example, some cores do not exhibit the same starting point and some lithofacies are repeated in the same cored intersection. C her cores exhibit a reversal of the typical coarsening-upward pattern and some cores are completely lacking an expected lithofacies in the succession.

4.3.1 PEACE RIVER PLAINS AREA

The Peace River Plains area typically consists of a coasrening-upward series of interbedded bioturbated shaly sandstones, cross-bedded clean sandstones and conglomerates (Figure 4.9). The facies succession normally includes lithofacies 5 to 8, with the conglomerates of lithofacies 1 occurring as discrete interbeds.

As mentioned above, some lithofacies types do not occur in some ed intersections, depending upon the location of the core in the study area. For example, in the most seaward sandstone buildups, lithofacies 7, the high-angle cross-bedded sandstones, are infrequent; whereas, in the more landward intersections this lithofacies is common. The relative proportions of these lithofacies also varies considerably; lithofacies 8 is consistently the thinnest, but either lithofacies 5 or 7 can be the thickest. The study area contains localized thick buildups of sandstone which show lateral changes in lithofacies along and perpendicular to the strike of the sand bodies. In the areas between the thick buildups, the formation lacks the development of bioturbated (lithofacies 5) or cross-bedded (lithofacies 7) sandstones, consisting only of poorly-sorted conglomerate deposits (sublithofacies 1a). The thickest buildups of sandstone (>20m) occur within Townships 78-83, Ranges 13-20 W6 and within NTS Block 94-H-7. The average buildup thickness is approxiamately 9 m, whereas the conglomerate deposits are only 3m.

The base of the vertical succession is consistently marked by an erosional surface which truncates the underlying Gething Formation. This truncation surface marks a major shift in depositonal conditions from mainly alluvial to marine sedimentation. Although this contact is broadly conformable with bedding, it has local irregularities with relief up to 15 cm. Poorly sorted conglomerates of sublithofacies 1a generally overlie the erosion surface. The contact is frequently

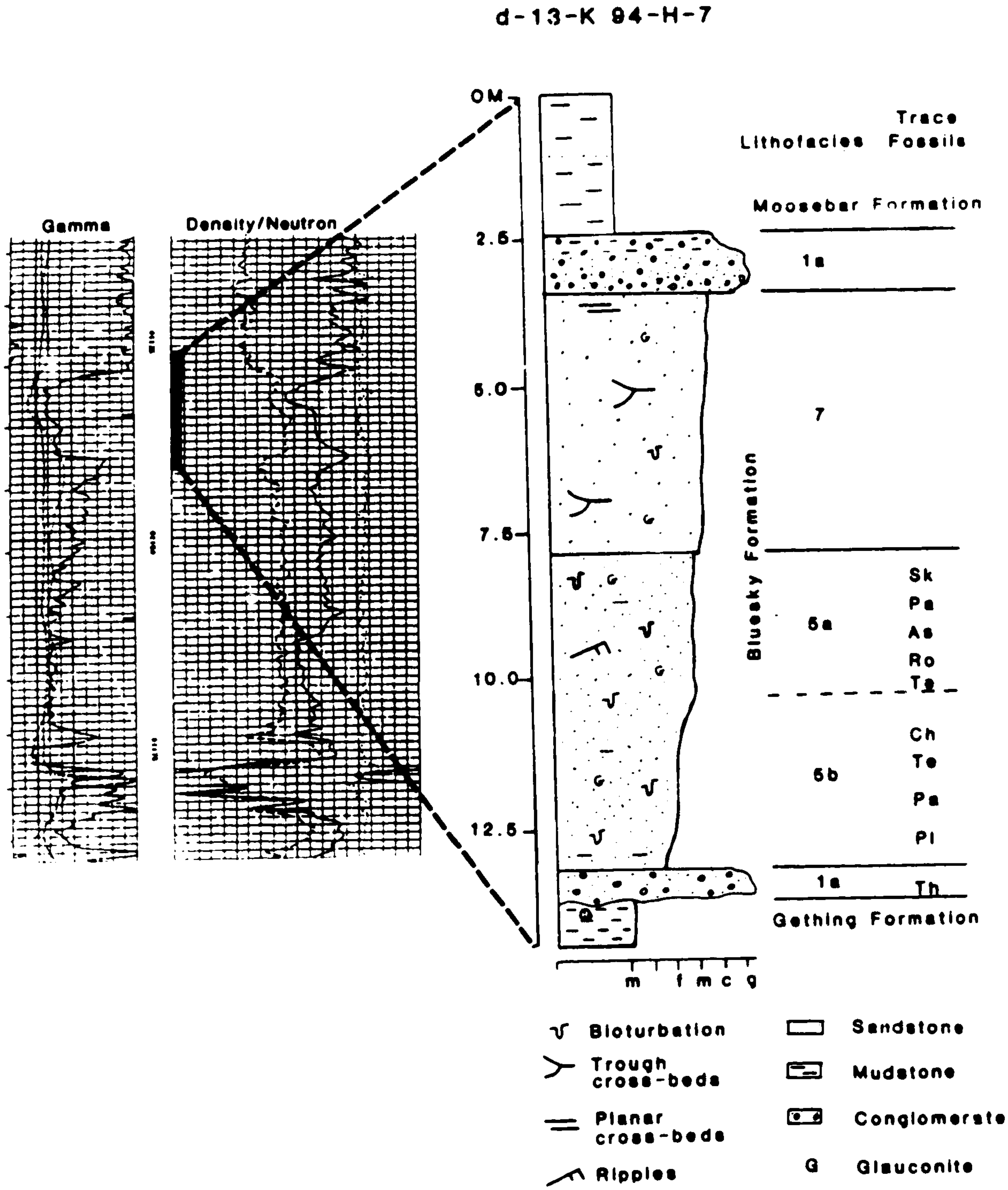


Figure 4.9 : Typical vertical facies succession for the Bluesky Formation in the Peace River Plains subsurface.

bioturbated and *Thalassinoides* burrows are present. These sharp-walled, unlined burrows are incised into the carbonaceous mudstones of the Gething Formation and, are passively-infilled by the overlying conglomerate. This implies that the burrows were emplaced by benthic organisms in a saline setting prior to the deposition of the conglomerate. The poorly-sorted nature of the conglomerate indicates a fairly rapid input of coarse-grain detritus.

The basal conglomerates are generally overlain by interbeds of sublithofacies 5a and 5b. Occasionally, low-angle stratified conglomerates (sublithofacies 1b) cap the basal conglomerates. The basal section of the sandstones is generally formed by the more shaly bioturbated sandstones with abundant deposit-feeding burrows. These gradually clean- and coarsen-upwards to deposits of sublithofacies 5a. The sandstones of sublithofacies 5a appear to have been influenced by higher wave activity or shoaling, indicating a slight shallowing of water depths. The bioturbated sandstones are occasionally punctuated by clean, massive sandstones of lithofacies 6, providing a laminated-to-burrowed appearanc[illegible]re. The bioturbated sandstones are gradationally overlain by clean, cross-bedded sandstones of lithofacies 7. Cross-laminated sandstones often contain interbeds of low-angle cross-stratified conglomerates (sublithofacies 1b).

This lithofacies succession indicates, for the most part, progradation and shallowing of water depths as the sandstone layers accrete. The depositional continuum can be interupted, though, by a reversal of the progradation, accompanied by a shift back to bioturbated sandstones from laminated sandstones. This may indicate either a deepening of the water or a shift in sediment supply and abandonment of the sandstone buildup. The succession is normally capped by a poorly-sorted conglomerate before passing abruptly into Moosebar Formation shales.

4.3.2 ROCKY MOUNTAIN FOOTHILLS AREA

The vertical succession for the Bluesky Formation in this area consists of a coarsening-upward series of interbedded sandstones and mudstones deposited in a completely different setting than that of the Peace River Plains subsurface. The base of the vertical succession in this area is also marked by an erosional surface which is commonly bioturbated with the presence of *Thalassinoides* burrows and overlain by conglomerate deposits of sublithofacies 1a. These conglomerate deposits are fairly widespread, occurring consistently in drill core throughout the Foothills and in several outcrops. The thickness of the conglomerate appears to be fairly uniform with localized thicks scattered throughout the study area.

The poorly-sorted conglomerates are typically abruptly overlain by the coarsen or sandier-upwards interbedded sandstone and mudstone of lithofacies 2 (Figure 4.10). The thickness of this unit varies considerably, with thinning north of the Sukunka River area, thickening in the area between the Sukunka and Wolverine rivers and thinning south of this area. It is difficult to distinguish sedimentologic thickening from tectonic thickening within the Foothills and, as such, some cored intersections may be tectonically thickened by low-angle thrusting.

This lithofacies, in the southern half of the Foothills, tends to show thicker, planar laminae and coarser grain sizes. Parallel and lenticular silt laminae become more common and tend to be very thinly bedded in the northern half of the region. The coasrening-upward succession in the Pine Pass and Crassier Creek areas is capped by the glauconitic mudstones of lithofacies 4, before being abruptly overlain by Moosebar shales. In the Sukunka region, the glauconitic mudstones are followed by a continuance of the interbedded sandstone and mudstone lithofacies.

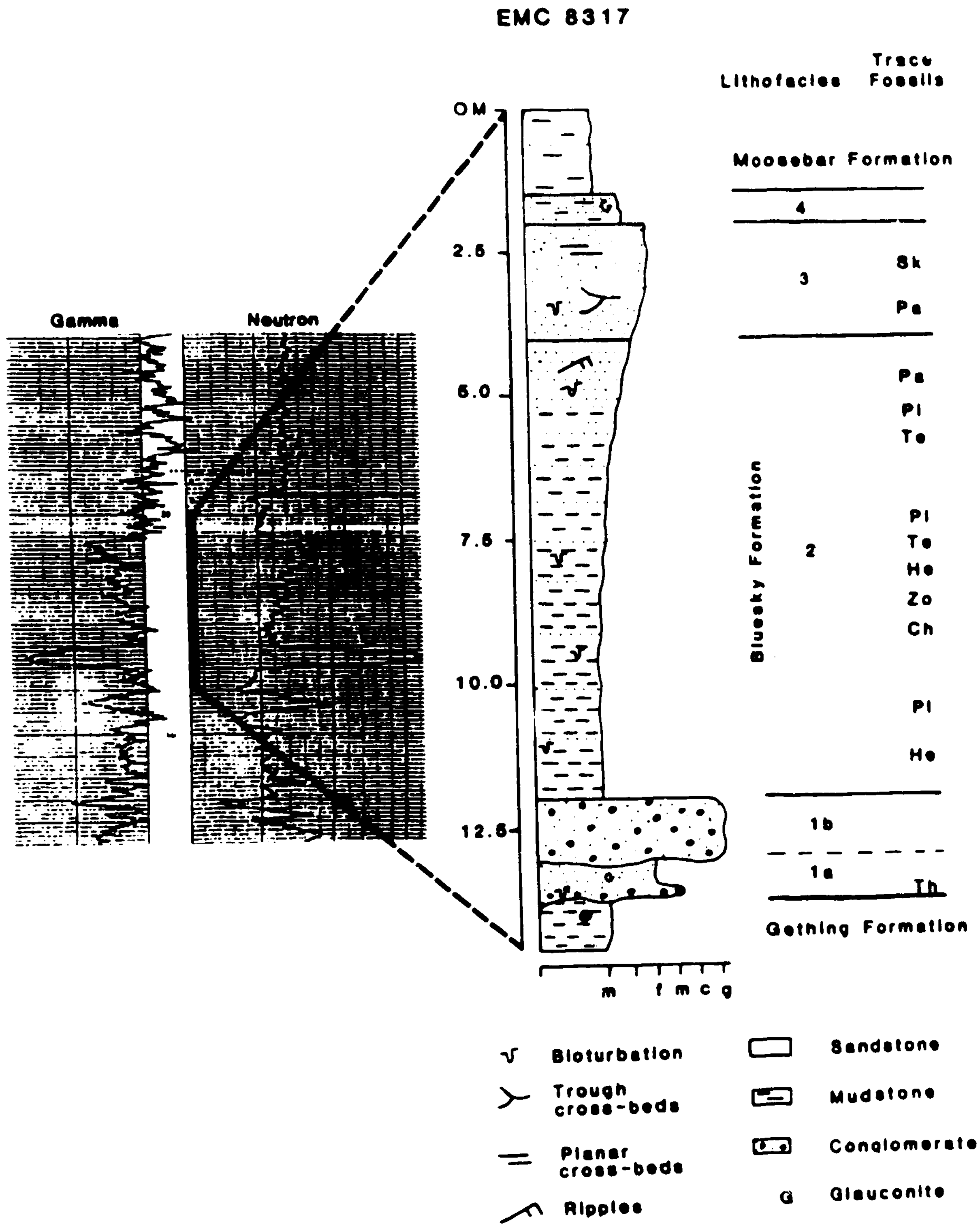


Figure 4.10 : Typical vertical facies succession for the Bluesky Formation in the Foothills area near Pine Pass.

Progressing upwards within the succession, the small scale cross-laminated sandstones (lithofacies 3) gradationally overlie, and sometimes interfinger with, lithofacies 2. In the landwards portions, towards the south, the sandstones are characterized by cleaner sediments, fewer interbeds of silt or mud, and exhibit a variety of well-developed small-scale cross-laminae, current ripples, and planar laminations. Towards the seaward positon, in the north half, the sandstone beds become thinner, muddier and exhibit more bioturbation. The complete succession in the northern half of the Foothills region averages less than half of the total thickness in the southern half. The cross-sections in Figure 2.6 illustrates the thinning of the Bluesky Formation in a northwards direction south of Township 77.

The small-scale cross-stratified sandstones are typically overlain by the fresh-water deposits and coals of the Chamberlain Formation. The complete sequence in the Foothills, in general, records a shallowing upward of water depth as the shoreline of the Chamberlain delta progrades northward.

5.0 ICHNOLOGY

5.1 INTRODUCTION

In a marine environment the habits and habitats of tracemakers are considered to be dependent upon the local environmental parameters, such as temperature, salinity, food supply and substrate consistency and level of water agitation (Ekdale *et al.*, 1984). Seilacher (1967a) noted that many of these ecological parameters which influence the abundance and distribution of macro-invertebrates tend to change progressively with increasing water depth and, as such, can be divided into distinct zones. In most respects this zonation by Seilacher (1967a) is fundamentally a powerful tool; however, it must be emphasized that the abundance and distribution of benthic organisms is related to a host of ecological parameters rather than solely bathymetry (Rhoads, 1975).

Trace fossils are the preserved record of the behavior and functional morphology of tracemakers as they adapt to specific environmental conditions. In short, particular assemblages of trace fossils appear to be characteristic of specific environments, occurring wherever and whenever the requisite ecological parameters are met. As such, the practical value of trace fossils lies with their use in paleoenvironmental analyses when applied cautiously within the full line of physical and biological evidence (Frey and Pemberton, 1985).

5.2 GENERAL PALEOECOLOGICAL SIGNIFICANCE OF TRACE FOSSILS

community of environmentally-related traces is known as an ichnocoenose. The preserved record of that ichnocoenose is known as the ichnofacies (Ekdale *et al.*, 1984). Seilacher (1967a) designated six fairly distinct and recurrent ichnofacies, which are found in a variety of typical marine environments. These ichnofacies are composed of trace fossil associations that represent specific environmental niches and(or) bathymetric zones (Frey and Pemberton, 1985).

One of the key aspects in attaching some form of paleoecological and geological significance to trace fossils is to interpret the type of behavior the trace records. Seilacher (1967b) has classified the type of behavior displayed by organisms into six major categories: resting traces *(Cubichnia)*, grazing traces *(Pascichnia)*, feeding burrows *(Fodinichnia)*, locomotion traces *(Repichnia)*, dwelling structures *(Domichnia)* and escape structures *(Fugichnia)*. The recognition of the type of ethological structure is most useful in environmental analysis, as well as in the identification of the general trophic group. It must be remembered that not only do the types of trace fossils vary widely because of the diversity in species, but that the same organism may produce different traces under different conditions (Frey, 1978).

In addition to the type of behavior the trace fossil records, the ichnologist can attach some ecologic significance to the preservability of traces, sediment type, salinity, temperature and energy levels. Preservation of lebensspuren are primarily influenced by the prevailing energy conditions and rate of sedimentation in the depositional zone. Traces are much less likely to be preserved in the high-energy conditions of the upper shoreface setting than the quieter conditions of the mid- to lower-shoreface (Frey and Pemberton, 1985). Criteria such as strength of construction and subsurface extent of the lebensspuren also influence the selective preservability of an assemblage of traces.

As mentioned earlier, a critical aspect of the environment to a benthic organism is the stability and consistency of the substrate (Purdy, 1964; Ekdale *et al.*, 1984). Different organisms require certain types of substrates, whether it is soft or firm, shifting or stationary and, indeed, some organisms display different behavioral responses to varying substrate stabilities (Ekdale *et al.*, 1984). Of the seven recurring ichnofacies, the traces of the consolidated substrates, including the hardground (*(Trypanites)*, firmground *(Glossifungites)* and woodground *(Teredolites)* ichnofacies are dominantly controlled by the substrate type moreso than the water depth or other environmental parameters (Ekdale *et al.*, 1984). The soft-ground ichnofacies (ie. *Skolithos*, *Cruziana*, *Zoophycos*, and *Nereites*) are influenced primarily by the environment that surrounds them.

Sediment grain size and composition seem to have little effect on the distribution of tracemakers except in extreme conditions (Frey and Pemberton, 1985). The common association of one particular trace to a class of grain size is generally the result of a mutual response to hydraulic conditions rather than an animal-sediment affinity (Purdy, 1964; Frey and Pemberton, 1985). Sediment textures are limiting factors only in certain cases of infaunal and epifaunal bioturbation. Settings such as gravel fields and stagnant muds are inhospitable to these forms. Extremely turbid waters in a muddy environment generally do not support filter-feeders (Ekdale *et al.*, 1984). Figure 5.1 illustrates the relative proportion of dwelling, feeding and grazing structures in differing substrate types as observed in the Bluesky cores. For example, the majority of dwelling structures occur in medium- to coarse-grain sandstones, whereas the bulk of feeding and grazing structures occur in mudstones.

Variations in salinity and temperature of the water are important factors in the lifestyles of some organisms. A common response to high stress conditions such as these is to buffer the fluctuations by intrastratal living. Sediment cover is an effective screen for slowing down the exchange of pore waters (Rhoads, 1975). Thus, infaunal organisms tend to be more abundant than epifaunal organisms in such high stress environments. The brackish-water environment, in general, also contains different and fewer taxa than the fully marine environment (Ekdale *et al.*, 1984). Howard and Frey (1975) established animal-sediment trends for the Georgia Estuaries which distinctly demonstrated a lower diversity of biota for these brackish-water settings.

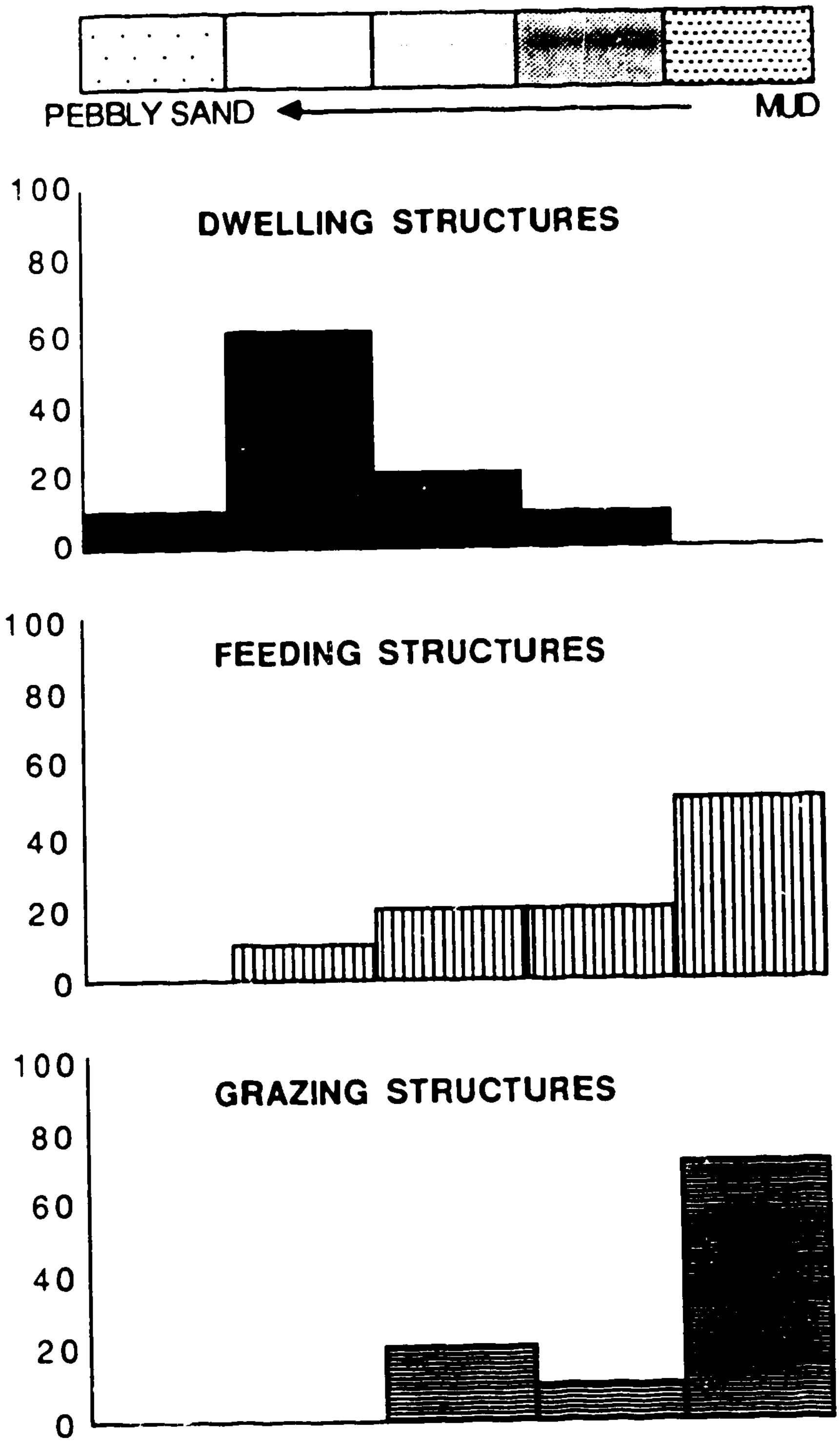


FIGURE 5.1: FREQUENCY DISTRIBUTIONS OF THE RELATIONSHIP BETWEEN TROPHIC GROUPS AND SUBSTRATE TYPE.

A final consideration, but probably the most important, to the ecologic significance of trace fossils involves the role of waves and currents (Frey and Pemberton, 1985). Their primary role is in food and sediment distribution, and aeration of sediments and water (Frey and Pemberton, 1985). Above wave base, the strength of wave action determines the abundance and type of trace as well as the preservability of the trace; but below wave base the type and number of species is determined by the available current and the food it distributes. This type of energy-depth dependenco is best portrayed in a typical beach to offshore setting. Higher energy conditions are found in the shallower zones, and thus are dominated by physical sedimentary structures rather than biogenic. Towards deeper water and lower energy conditions in the offshore, the sediments become increasingly bioturbated. As a result, most physical sedimentary structures are destroyed in the offshore (Howard and Frey, 1984; Ekdale *et al.*, 1984).

The above discussion contains just a few of the parameters that are crucial to ichnology and its use in sedimentology; more complete reviews of the paleoecological significance of trace fossils are provided in the references cited within the discussion.

5.3 DESCRIPTION OF ICHNOFACIES

The occurrence of ichnofauna in Bluesky core fall into 3 distinct, somewhat intergradational ichnofacies. These divisions are made follow . the universal ichnofacies classification of Seilacher (1967a). A fourth yet unnamed ichnofacies consisting of a brackish-water assemblage is also discerned in Bluesky cores. The occurrence of a brackish-water suite of trace fossils has also been noted by Ekdale *et al.*, (1984) and Wightman *et al.*, (1987).

The four ichnofacies represented in the Bluesky Formation are the high energy *Skolithos* ichnofacies, the lower energy *Cruziana* ichnofacies, a stratigraphically limited *Glossifungites* ichnofacies and the unnamed brackish-water ichnofacies (Ekdale *et al.*, 1984). Their respective characteristics and and bathymetric positions are discussed below and also summarized in Table 5.1.

The *Glossifungites* ichnofacies occurs in firm but unlithified substrates, intermediary between the free shifting, softground substrate of the *Skolithos* ichnofacies and the consolidated hardgrounds of the *Trypanites* ichnofacies (Frey and Pemberton, 1985; Pemberton and Frey, 1985) (Figure 5.2). The necessary firmground can be produced in several settings, of which the two most notable are (1) low-energy, protected setting of the intertidal to supratidal which offers periodic prolonged subaerial exposure or (2) in higher energy settings which can exhume unlithified top sediments to expose firm dewatered sediments, typically muds (Frey and Pemberton, 1985; Pemberton and Frey, 1985). An example of the latter is taking place on the present Georgia coast where a transgressing sea is unroofing buried marsh muds

Examples of the occurrence of the *Glossifungites* ichnofacies are generally scarce and poorly documented; however, a notable example is the modern Georgia Coast (Pemberton and Frey, 1985) and an ancient example from the Cretaceous Austin-Taylor contact of Texas (Fursich *et al.*, 1981).

Due to the somewhat limiting nature of the substrate, the suite of biogenic sedimentary structures tend to be dominated by vertical, U-shaped, or oblate borings and dwelling structures of suspension feeders or carnivores (Frey and Seilacher, 1980; Pemberton and Frey, 1985).

Pascichnia and fodinichnia structures are not expected due to the firmness of the substrate and paucity of organics (Ekdale *et al.*, 1984). Also, because of the substrate firmness, lining of the burrows for support is unnecessary (Rhoads, 1975; Frey and Pemberton 1984) and walls are

Table 5.1: General description of ichnofacies and their common trace fossils.

Typical Benthic Environment	Characteristic Trace Fossils
Glossifungites ichnofacies: Firm but unlithified marine littoral and sublittoral omission surfaces, especially semiconsolidated carbonate firmgrounds, or stable, coherent, partially dewatered muddy substrates either in protected, moderate-energy settings or or in areas of somewhat higher energy where clastic, semi-consolidated substrates offer resistance to erosion. Final sedimentary record typically consists of a mixture of relict and palimpsest features.	Vertical cylindrical, U- or tear-shaped borings or boring-like structures or sparsely to densely ramified dwelling burrows, protrusive spreiten in some, developed mostly through growth of animals. Fan-shaped ***Rhizocorallium*** and ***Diplocraterion***. Many intertidal species (eg. crabs) leave the burrows to feed, others are mainly suspension feeders. Diversity typically low, but given kinds of structures may be abundant.
Skolithos ichnofacies: Lower littoral to infralittoral, moderate to relatively high-energy conditions; slightly muddy to clean, well-sorted, shifting sediments, subject to abrupt erosion or deposition. (Higher energy increases physical reworking and obliterates biogenic sedimentary structures, lea[illegible] a preserved record of physical stratification).	Vertical, cylindrical or U-shaped dwelling burrows, protrusive and retrusive spreiten in some, developed mainly in response to substrate aggradation or degradation (escape or equilibrium structures); forms of ***Ophiomorpha*** consisting predominantly of vertical or steeply inclined shafts. Animals chiefly suspension feeders. Diversity is low, yet given kinds of burrows may be abundant.
Cruziana ichnofacies: Infralittoral to shallow cu[illegible] substrates; below daily wave base but not storm wave base, to somewhat quieter offshore-type conditions; moderate to relatively low energy, well-sorted silts and sands, to interbedded muddy and clean sands, moderately to intensely bioturbated, negligible to appreciable, although not necessarily rapid, sedimentation. A very common type of depositional environment, including estuaries, bays, lagoons, and tidal flats, as well as continental shelves or epeiric slopes.	Abundant crawling traces, both epi- and intrastratal; inclined U-shaped burrows having mostly protrusive spreiten (feeding swaths; soft sediment ***Rhizocorallium***); forms of ***Ophiomorpha*** and ***Thalassinoides*** consisting of irregularly inclined to horizontal components, scattered vertical cylindrical burrows. Animals include mobile carnivores and both suspension and deposit feeders. Diversity and abundance generally high
Brackish Water Ichnofacies: Marginal marine, littoral, intertidal to subtidal brackish water conditions; intimate facies relationship between non-marine and marine beds; water salinity subject to erratic fluctuations (eg. input of fresh water from rivers, rains); typical settings are estuaries, bays, delta platforms, tidal flats.	Vertical and horizontal burrows common to both ***Skolithos*** and ***Cruziana*** ichnofacies; abundant crawling, feeding trails, associated with deep lined vertical burrows, reduced diversity of traces; forms typcially found in marine environment; generally reduced sizes of traces constructed by organisms which employ non-specialized feeding strategies.

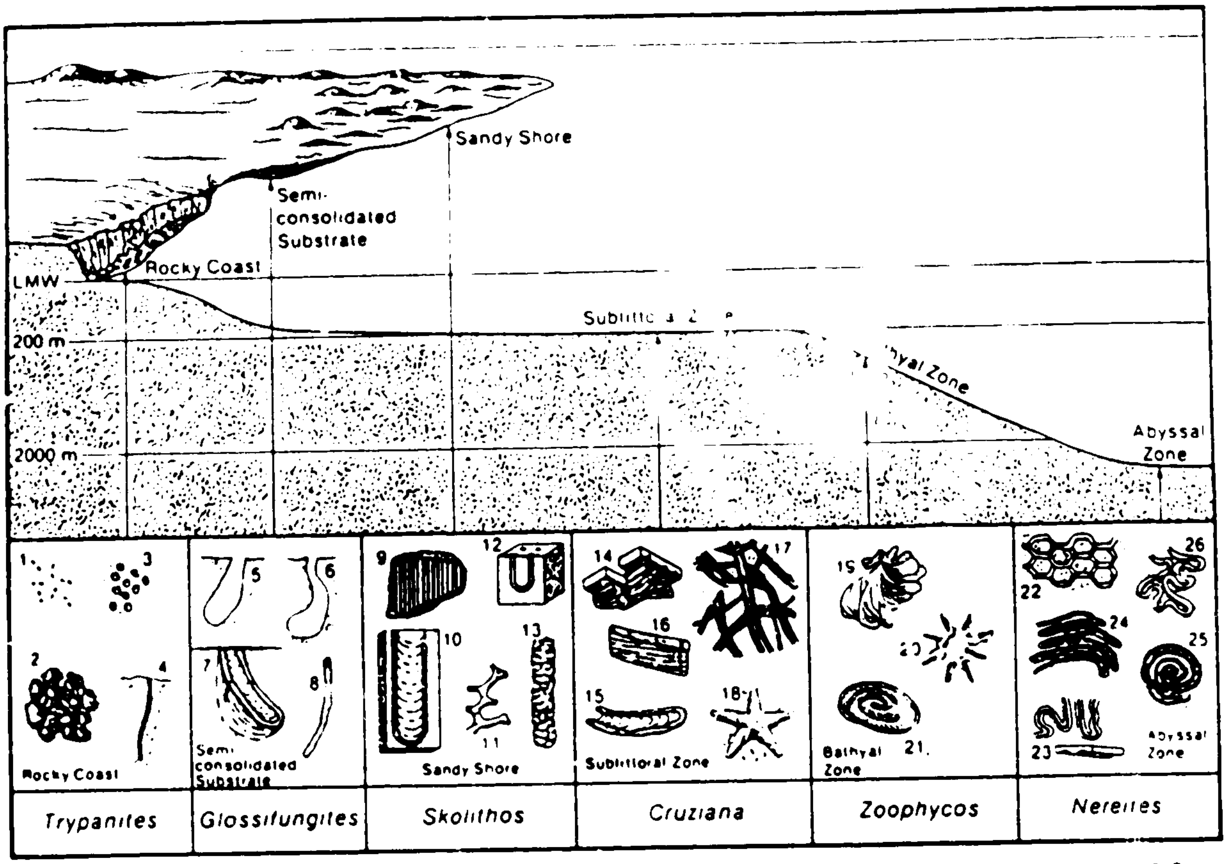


FIGURE 5.2

Recurring marine ichnofacies set in a representative but not exclusive suite of environmental gradients. Local physical, chemical and biological factors ultimately determine which traces occur at which sites. Typical trace fossils include: 1) Caulostrepsis, 2) Entobia, 3) echinoid borings, unnamed, 4) Trypanites, 5-6) Gastrochaenolites or related ichnogenera, 7) Diplocraterion, 8) Psilonichnus, 9) Skolithos, 10) Diplocraterion, 11) Thalassinoides, 12) Arenicolites, 13) Ophiomorpha, 14) Phycodes, 15) Rhizocorallium, 16) Teichichnus, 17) Crossopodia, 18) Asteriacites, 19) Zoophycos, 20) Lorenzinia, 21) Zoophycos, 22) Paleodictyon, 23) Taphrhelminthopsis, 24) Helminthoida, 25) Spirorhaphe, 26) Cosmorhaphe. (Modified from Crimes, 1975; Frey and Seilacher, 1980.) Bathymetric terms are those reiterated by Ager (1963), mainly from the Treatise on Marine Ecology and Paleoecology. Not to scale.

sharply defined and may be sculpted (Frey and Pemberton, 1984). Thus, common traces to the *Glossifungites* ichnofacies are *Thalassinoides*, *Rhizocorallium*, *Skolithos* and numerous polychaete burrows (Pemberton and Frey, 1985). Diversity is generally low, but burrow density may be high (Frey and Seilacher, 1980).

The *Skolithos* ichnofacies is characterized by freely shifting, well-sorted arenaceous substrates that are associated with relatively high wave and current activity. (Seilacher, 1967a; Pemberton and Frey, 1983; Frey and Pemberton, 1985). Within this setting, abrupt variations in the rate of deposition, scouring and physical reworking are quite common (Frey and Pemberton, 1985) due to the changing nature of the hydraulics in the setting. This type of setting is common to the foreshore and shoreface of beaches, bars and spits (Frey and Pemberton, 1985). The high energy is also reflected in the sedimentary structures, where primary physical structures tend to predominate over biogenic ones (Howard, 1975; Ekdale *et al.*, 1984).

Because of the intensity of the current and wave action and associated rapid fluctuations in temperature and salinity, the foreshore and shoreface tend to be very demanding and exacting environments (Crimes, 1975; Ekdale *et al.*, 1984). Thus, the majority of tracemakers that can tolerate these conditions tend to be infaunal suspension feeders. Their response to these conditions is to anchor themselves into the substrate by producing long, reinforced vertical burrows which act as permanent domiciles. Thus, a variety of vertically-oriented agglutinated worm tubes, branching pelletoid-lined, and various U-shaped domichnia as well as some repichnia structures will be found (Ekdale *et al.*, 1984). Trace fossils such as *Skolithos*, *Diplocraterion*, *Ophiomorpha*, *Monocraterion*, and some forms of *Arenicolites* are typical to this setting (Ekdale *et al.*, 1984). Usually, the diversity is low but abundance of any specific burrow may be high (Frey and Seilacher, 1980). To compensate for the freely-shifting substrates, and to help buffer themselves from wide fluctuations in salinity and temperature, the organisms will burrow deeply into the substrate where conditions are not so variable (Rhoads, 1975).

The *Skolithos* ichnofacies, which characterizes the nearshore zone above fairweather wave base, is differentiated from the *Glossifungites* ichnofacies by the softground substrate. Whereas, *Skolithos* ichnofacies can be separated from the *Cruziana* , *Zoophycos* and *Nereites* ichnofacies by the assemblage of traces contained in each (Eckdale *et al.*, 198-). Following Seilacher's (1967a) model, the *Skolithos* ichnofacies grades seaward into the *Cruziana* ichnofacies.

The *Cruziana* ichnofacies occupies the offshore zone below fairweather wavebase but above stormwave base (Frey and Seilacher, 1980). Generally, the energy levels range from moderate in shallower waters to low energy in deeper, quieter waters of the distal offshore (Pemberton and Frey, 1983). Bedding styles and sediment textures are fairly diverse (Pemberton and Frey, 1983). Thin fine-grained sand containing organic detritus, in laminated to rippled cross-bedded units are interbedded with intensely bioturbated poorly sorted, homogenous beds where any physical stratification is obscured. Distal parts of the offshore are mostly totally bioturbated mudstones.

The *Cruziana* ichnofacies, although common to the low-energy offshore or the neretic zone of continental shelves and epeiric seas (Ekdale *et al.*, 1984), has been noted in marginal marine settings such as estuaries (Howard and Frey, 1975) and lagoons (Fursich, 1975).

Because the overall energy gradient is low, food is available in high concentrations as both suspended and deposited components (Purdy, 1964; Ekdale *et al.*, 1984). Therefore, characteristic organisms are abundant and diverse including both suspension- and deposit-feeders as well as mobile carnivores and foragers (Ekdale *et al.*, 1984). Thus, the *Cruziana* ichnofacies will be represented by diverse behavioral responses including dwelling, resting, feeding, crawling and escape structures (Ekdale *et al.*, 1984). Traces diagnostic of the *Cruziana* ichnofacies includes *Cruziana*, *Teichichnus*, *Rosselia*, *Phycodes*, *Dimorphichnus*, and some forms of *Asterosoma* and *Ophiomorpha*. Burrow forms will range from simple intrastratal feeding patterns to inclined U-shaped burrows having mainly protrusive spreiten. However, because the external stimuli (ie. temperature, salinity, oxygen) are less variable than high energy zones, the burrows will have a pronounced horizontal component rather than vertical. Thus, the assemblage of traces will be dominated by grazing and feeding structures with subordinate, underdeveloped vertical burrows. The chances of preservation for burrows in this setting are quite high since the sediments are largely undisturbed by fairweather shoaling waves or currents.

The brackish water ichnofacies is commonly found in marginal marine environments such as shallow lagoons, bays, estuaries, tidal channels and delta platforms (Ekdale *et al.*, 1984). These environments are generally somewhat restricted,; most having a small or narrow aperture that inhibits circulation with fully marine waters. Salinity gradients are typical and result from: (1) variation in freshwater runoff from land, (2) rainfall, (3) salinity content fluctuations in tidal ranges, (4) changes in wind direction and velocity and (5) evaporation (Dorjes and Howard, 1975; Ekdale *et al.*, 1984). Such drastic changes in the environment will have significant ecologic ramifications on the biota. In short, it creates a very stressful environment for the organisms (Rhoads, 1975). So in order for a benthic community to successfully populate this area, they must be able to accommodate and modify their life strategy.

The fauna which ultimately inhabits this setting is not a combination of true freshwater or true marine species, but one which can be considered to be more of an impoverished marine assemblage (Ekdale *et al.*, 1984). Thus, in coping with the high stress niche, a number of systematic trends in the resident biocoenose have been noted, including: (1) a reduction in the diversity of species, (2) infaunal organisms are more abundant than epifaunal, (3) reduction in overall size of some species, and (4) shift in the normal bathymetric range for a particular species (Wightman *et al.*, 1987).

Thus, the typical brackish-water ichnocoenose will be characterized by: a low diversity of species, ichnotaxa common to marine settings, simple, non-specialized structures constructed by trophic generalists, and a mixed association of vertical and horizontal burrows typical to both the *Skolithos* and *Cruziana* ichnofacies (Ekdale *et al.*, 1984). Typical examples of ichnogenera found in a brackish water assemblage are *Skolithos*, *Monocraterion*, *Thalassinoides*, *Ophiomorpha*, *Chondrites*, *Planolites* and *Palaeophycus* (Howard and Frey, 1975; Ekdale *et al.*, 1984).

Occurrences of brackish water trace fossil assemblages have been noted extensively in modern settings such as the Georgia estuaries (Howard and Frey, 1975), but few have been described for ancient rocks. Two notable exceptions are the McMurray Formation of northeast Alberta (Pemberton *et al.*, 1982; Benyon *et al.*, 1988) and the recently described brackish water assemblage of the Upper Mannville Group of east-central Alberta (Wightman *et al.*, 1987). The paucity of ancient studies is probably due more to inaccurate recognition and documentation rather than a lack of occurrence in ancient rocks.

5.4 ICHNOLOGY OF THE BLUESKY FORMATION

The Bluesky Formation contains a rich and varied collection of ichnofossils. Representatives of at least 15 ichnogenera have been identified, including: *Asterosoma*, *Chondrites*, *Cylindrichnus*, *Diplocraterion*, *Gyrochorte*, *Helminthopsis*, *Ophiomorpha*, *Palaeophycus*, *Planolites*, *Rosselia*, *Skolithos*, *Teichichnus*, *Terebellina*, *Thalassinoides* and *Zoophycos*. As well, a number of bioturbate textures were encountered; unfortunately, due to the intensity of the burrowing, individual forms were impossible to discern

Most trace fossils occur in distinct ichnofossil suites which exhibit conspicuous environmental zonation and represent distinct, sometimes intergradational, ichnofacies. Trace fossil suites of the Bluesky Formation contain dwelling, feeding, crawling and grazing structures probably produced by deposit-feeding-, suspension-feeding-, or scavenging-polychaetes, carnivores, bivalves and crustaceans. Thus, accordingly, each ichnogenus can be assigned to an ethological category, a general trophic group and a probable producer of the trace (Table 5.2). In this way, the occurrence of the ichnofossils and their associations can be used as further clues in the reconstruction of paleoenvironments.

5.4.1 TRACE FOSSIL DESCRIPTIONS

The following discussion includes brief descriptions of the encountered trace fossils in Bluesky Formation cores. These descriptions generally apply to all occurrences of individual ichogenera in the core. For systematic taxonomic descriptions of ichnossils, the reader is referred to Hantzschel (1975). Burrow descriptions are summarized from a number of references including: Howard and Frey (1984), Pemberton and Frey (1982,1983) and Hantzschel (1975) as well as their appearance in the Bluesky cores.

ASTEROSOMA

Star-shaped burrow system consisting of tapered finger-like arms radiating from a vertically-inclined central shaft; arms may exhibit longitudinal striae or wrinkles. The arms are usually circular to oval in cross-section (Figure 5.3a). Internally the burrows consist of concentric

Table 5.2: Ethological and trophic classification of ichnofossils from Bluesky cores.

Ichnogenus	Ethological Classification	Trophic Group	Probable Producer
Asterosoma	fodinichnia	deposit-feeder	annelid
Chondrites	fodinichnia	deposit-feeder	sipunculid/annelid
Cylindrichnus	fodinichnia/domichnia	deposit-feeder	annelid
Diplocraterion	domichnia	suspension-feeder	amphipod
Gyrochorte	fodinichnia	deposit-feeder	annelid
Helminthopsis	paschichnia	deposit-feeder	annelid
Ophiomorpha	domichnia	suspension-feeder	crustacean
Palaeophycus	domichnia	carnivore	annelid
Planolites	fodinichnia	deposit-feeder	annelid
Rosselia	fodinichnia	deposit-feeder	annelid
Skolithos	domichnia	suspension-feeder	annelid
Terebellina	domichnia/fodinichnia	suspension-feeder	annelid
Teichichnus	fodinichnia	deposit-feeder	annelid
Thalassinoides	domichnia	suspension-feeder	crustacean
Zoophycos	paschichnia	deposit-feeder	annelid/sipunculid

laminations of mud and sand Individual ovoid arms range in diameter from 20 to 40 mm. The burrows are preserved as epichnia and endichnia. *Asterosoma*, based on its tubular construction of galleries and the details of sediment processing, has been interpreted to be the feeding burrow of a worm. The burrow is part of the fodinichnia behavioral classification representing a deposit-feeder. Accordingly, the worm seems to have repeatedly probed into the sediment to enlarge the gallery vertically and laterally (Pemberton and Frey, 1983)

Burrows identified as *Asterosoma* have been found only in Bluesky cores from the Peace River Plains area, occurring in bioturbated sandstones of lithofacies 5.

CHONDRITES

Burrows belonging to the ichnogenus *Chondrites* are 'root-like' dentritic patterns of small, regularly-branching feeding tunnels; individual tunnels are uniform in diameter and do not normally interpenetrate or cross. The branches connect to one or few main vertical axes open to the surface. These branching tunnels trend downwards then fan outwards subparallel to bedding in the distal parts. In core, *Chondrites* usually appears as tiny (1-3 mm) dots or oblique, subvertical, small tubes. Generally the infill is structureless and more argillaceous than the surrounding host. This ichnogenus is preserved mainly as endichnia.

The *Chondrites* burrows have been suggested to represent tunnels produced by deposit-feeding sipunculids or polychaete organism which worked from the main axis, connected to the surface, probing deeply and repeatedly into the substrate with its proboscis. Thus, the burrow belongs to the fodinichnia ethological category.

FIGURE 5.3 CORE PHOTOGRAPHS OF ICHNOFOSSILS

a) *Asterosoma* burrow in bioturbated, glauconitic sandstone, lithofacies 5. Core is 3 cm wide in photograph. a-23-A 94-B-8, 922.2 m.

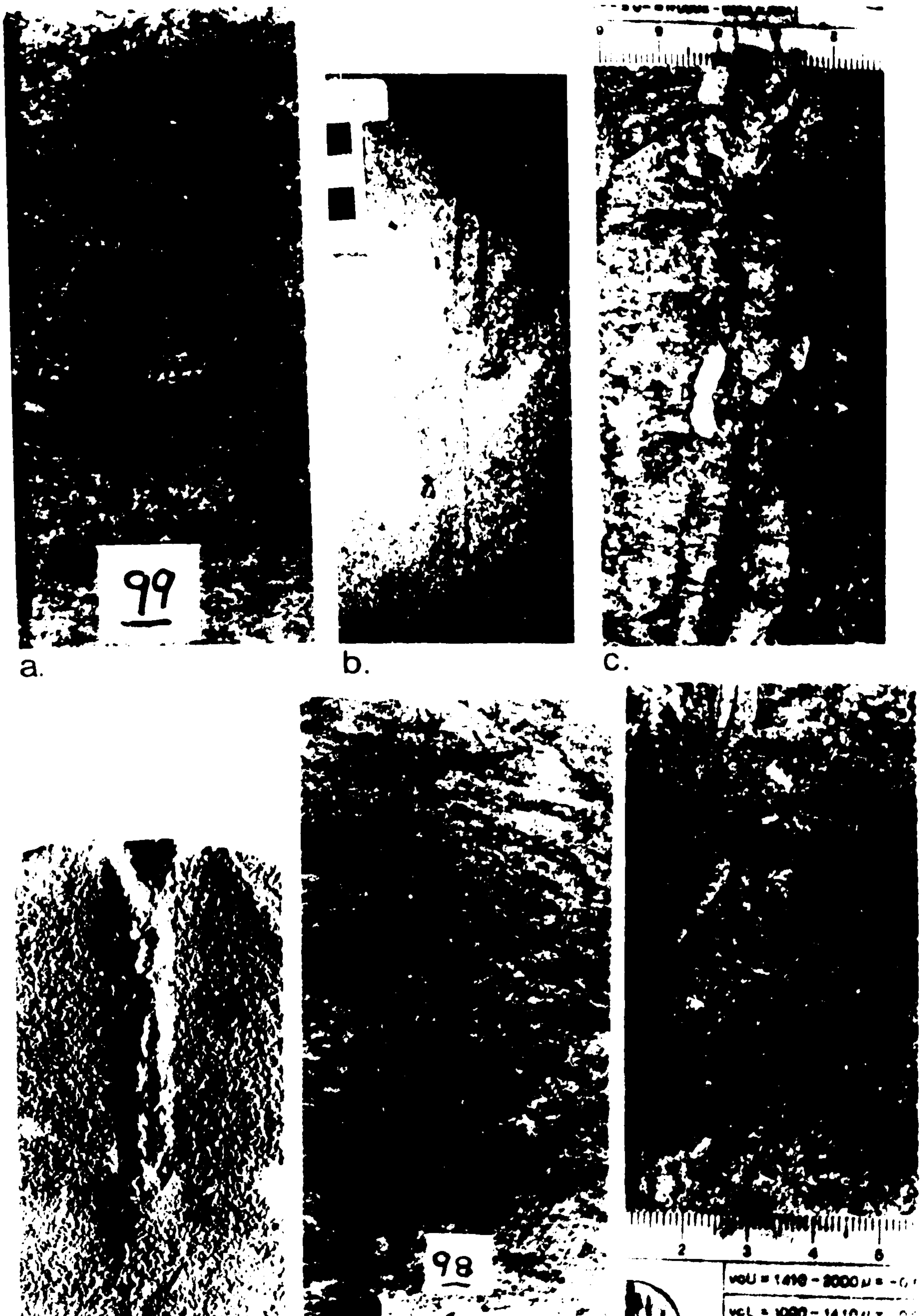
b) *Cylindrichnus* burrow (arrow) exhibiting the multi-lined walls of the burrow. Core is 8 cm wide in photograph. d-33-C 94-H-11, 1105.81 m.

c) *Diplocraterion* burrow (solid arrow) in bioturbated, glauconitic sandstone, sublithofacies 5a. Possible *Skolithos* burrow is indicated by hollow arrow. PD 8001, 533.5 m.

d) *Gyrochorte* in the bedding plane of a mudstone bed, lithofacies 2. Core is 3 cm wide in photograph. GDR 8021, 220.2 m.

e) *Helminthopsis* (arrows) occur as thin, wispy black lines in bioturbated glauconitic fine-grain sandstone. Core is 5 cm wide in photograph. a-23-A 94-B-8, 920.5 m.

f) *Ophiomorpha* burrow in bioturbated sandstones of the Peace River Plains area, PD 8001, 538.4 m.



a. b. c.
d. e. f.

Burrows identified as *Chondrites* occur in the Bluesky cores, in both the interbedded sands and muds (lithofacies 2) of the Foothills region and the bioturbated sandstones (lithofacies 5) in the Peace River Plains area.

CYLINDRICHNUS

This ichnogenus can be described as long, subcylindrical to subconical burrows straight to gently curved with a vertical to sub-horizontal orientation (Figure 5.3b). The walls of the burrow are multi-lined. In core, the burrow can be readily identified by the long, slender tube with concentrically layered walls. The burrows are usually oriented subvertically in Bluesky core. They average 15 cm in length and 1 cm in width at the top, tapering downwards. The burrows are generally preserved as endichnia. The linings of the walls are composed of fine grain sand with thin subparallel mud partings. The fill of the burrow, in Bluesky cores, is identical to the surrounding matrix. The multi-layered nature of the wall has been attributed to the activity of the organism as sediment enters the burrow. In times of slow, continuous sedimentation, the organism presses the intruding sediment to the wall of the burrow. These burrows have been interpreted to be dwelling or combined dwelling-feeding structures possibly occupied by annelids or coelentrates.

These traces have only been encountered in the bioturbated, sandstones of sublithofacies 5a in wells only from the Peace River Plains area.

DIPLOCRATERION

Diplocraterion are U-shaped burrows with spreite (Figure 5.3c). The burrows are vertically-oriented and perpendicular to the bedding plane. These burrows have two clyindrical to funnel-shaped openings to the sediment surface. Retrusive spreiten occurs when the organism migrates toward its apertures, in response to sedimentation. The spreite in this case are on the outside of the "U". Protusive spreiten is formed by the downward migration and lengthening of the burrow, so that the spreite is contained inside the "U". The designation of the type of burrow is dependent upon the position of the spreite relative to the final tunnel occupied by the tracemaker (Ekdale *et al.*, 1984). Protusive burrows are said to be caused by the ret[illegible] the organism when scouring is taking place at the sediment surface. Both retrusive and [illegible]ive spreite may occur in the same trace if alternating period of scour and deposition is occurring. The limbs of the "U" may be parallel or divergent.

In core, *Diplocraterion* are recognized by long slender shafts that which contain protrusive and retrusive spreiten structures. The infill is identical to the surrounding host. No wall linings or ornamentation were noted. The final position of the t[illegible]aker appears as a small pouch-like protrubance into the host sediment. Burrows in the Bluesky core average 20 cm in length and 15 mm in width. These structures are preserved as endichnia. *Diplocraterion* have been interpreted to be dwelling burrows (domichnia) of a suspension-feeding organism. Possible producers of these structures include polychaetes, echiuroids, and crustaceans (Pemberton and Frey, 1983). Possible environments have been suggested to be high energy wave zones (Hantzschel, 1975).

These burrows have only been noted in wells of the Peace River Plains area within the bioturbated, coarse-grain sandstone (sublithofacies 5a).

GYROCHORTE

Gyrochorte are horizontal trails comprised of distinct bi-lobate ridges which are tangentially aligned and separated by a median furrow (Figure 5.3d). The course of the trace may be sinuous, straight or meandering. The traces are unbranched and may intersect itself. The burrow is generally limited to less than 7 mm in width. The burrows are preserved as epichnia, endichnia, and hypichnia

These burrows have been attributed to the crawling-feeding movement of an elongate organism in search of food. Possible producers have been suggested to be annelids or other worm-like organisms, and ?crustaceans. These can be classified as fodinichnia.

In the Bluesky cores, this trace has only been noted in the interbedded sandstone and mudstone lithofacies (lithofacies 2) of the Foothills area. These traces are recognizable only on bedding planes of the mud units. In a vertical section, as seen in core, they can be mistaken for load or flute casts.

HELMINTHOPSIS

These burrows are simple, thin irregular to random meandering trails which never cross or interpenetrate. The burrows are generally horizontal to oblique to the bedding plane. The fill is usually mud size and unlike that of the surrounding host (Figure 5.3e). In the cross-section, the burrows are elliptical to circular and may contain sharp turns. Burrows in longitudinal section appear as thin, dark lines. *Helminthopsis* are usually preserved as endichnia or epichnia.

Helminthopsis is considered to be a grazing trail produced by veriform organisms or worm-like organisms. Thus, these trails can be ethologically classified as pascichnia.

These burrows have been noted in the interbedded sandstone and mudstone lithofacies (lithofacies 2) of the Foothills area and also in the bioturbated, fine-grain sandstones (sublithofacies 5b) of the Peace River area. In lithofacies 2 the burrows are generally restricted to the mudstone beds. These burrows are considered to be a facies breaking trace, but are most frequently found in deeper-water marine settings.

OPHIOMORPHA

Ophiomorpha consists of simple to complex, three-dimensional, cylindrical, dichotomously branching burrow systems distinctly lined with agglutinated pellotoidal sediment. The burrow walls consist of a dense mosaic of discoid, oval, mastoid, bilobate or irregularly shaped pellets pressed to the wall. The pellotoid nature of the lining gives the burrow a mammaliated appearance to the exterior. The burrows may be horizontal to vertical in orientation and where they branch the burrow is swollen. Infill of the burrow system is generally structureless and of similar composition as the surrounding host rock. The diameter of the burrows range from 5 to 12 mm. Individual ichnospecies of *Ophiomorpha* are distinguished by the type of burrow lining. These burrows are generally preserved as endichnia. In core, these burrows are recognized by their circular cross-section and by the two distinct concentric rings enclosing the wall lining (Figure 5.3f).

Ophiomorpha are considered to be the dwelling or combined dwelling-feeding structures of decapod crustaceans, particularly shrimp or shrimp-like animals. Thus, they may be classified as domichnia structures occupied by carnivores or suspension feeders.

Ophiomorpha have only been recognized in the bioturbated, coarse-grain sandstone (sublithofacies 5a) and the high angle cross-stratified sandstones (lithofacies 7), both in the Peace River Plains area.

PALAEOPHYCUS

Palaeophycus is essentially a cylindrical to subcylindrical, distinctly lined burrow oriented horizontal to oblique to the bedding plane; the infill is characteristically the same as the surrounding medium. Wall linings may be smooth or striated on the exterior. Individual ichnospecies may be distinguished by the thickness of linings and the presence and type of ornamentation. *Palaeophycus* may be preserved as endichnia, hypichnia, or epichnia.

In core, the burrows are recognized by their circular to elliptical cross-section with a distinct wall lining. The fill of the burrow is of the same lithology and color as the surrounding host. In a longitudinal view, the burrow appears similar to a small channel-like structure that tapers at both ends (Figure 5.4a) The burrows average 5 to 11 mm in diameter.

Pemberton and Frey (1982) indicated that the *Palaeophycus* burrow is distinguished from the morphologically similar *Planolites* burrow by the presence of a wall lining and the character of the burrow infill. *Planolites* typically has an infill different from that of the host rock. They also suggest that the *Palaeophycus* burrow is passively infilled by gravity-induced sedimentation. Deformation of the burrow is suggested to be caused by the emplacement of the burrow in soupy- or soft-ground, implying the burrow is relatively shallow in depth.

Burrows which are passively infilled and lined are typically interpreted as the dwelling structures of organisms such as polychaetes. The burrow of the modern predaceous glycerid polychaete is commonly used as an analog for the *Palaeophycus* trace (Pemberton and Frey, 1983). Thus, this burrow may be ethologically classified as a domichnion structure possibly occupied by a suspension-feeder (lithofacies 2), the small-scale cross-laminated sandstones (lithofacies 3), and the bioturbated, fine- to coarse-grain sandstones (lithofacies 5).

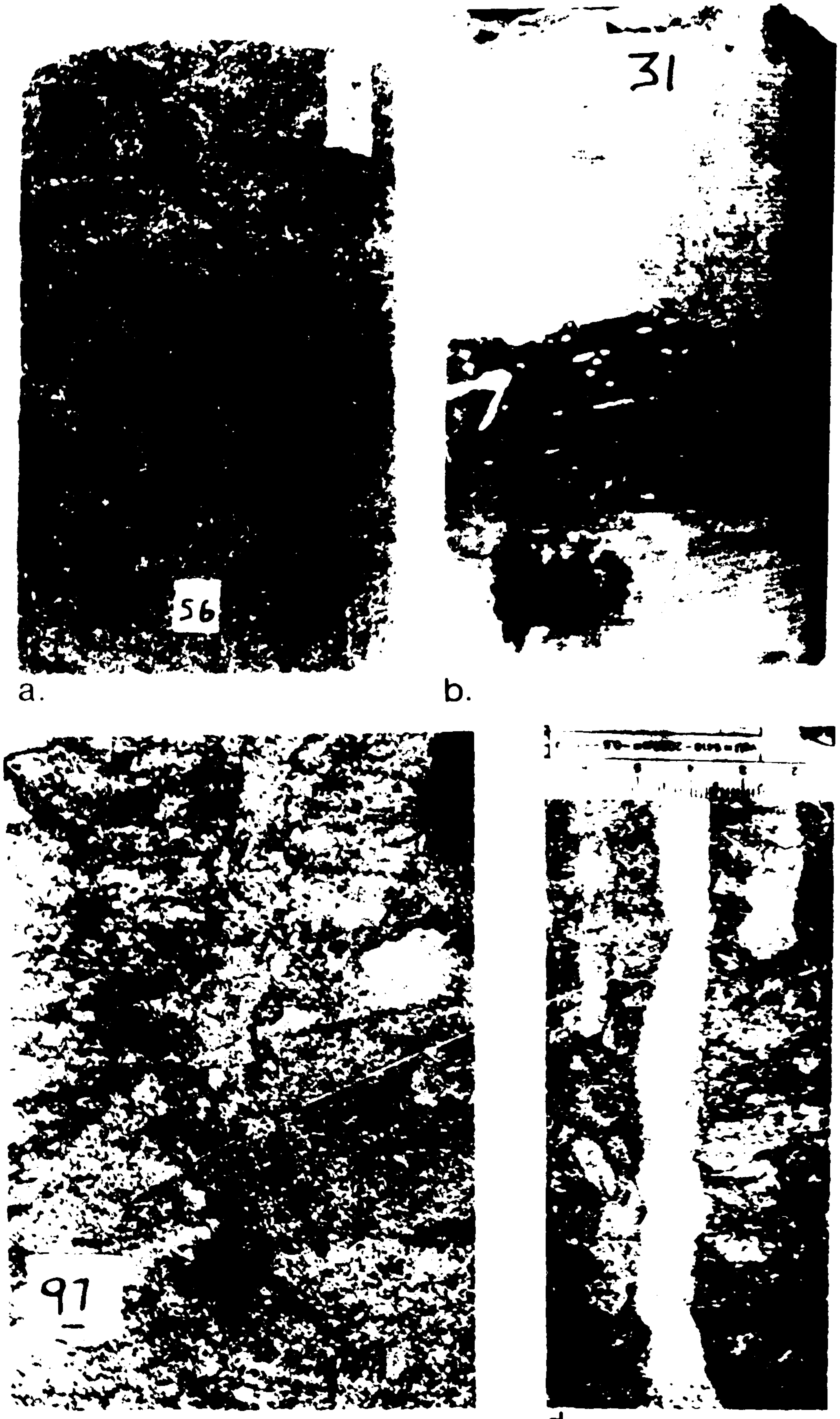
FIGURE 5.4 CORE PHOTOGRAPHS OF ICHNOFOSSILS

a) *Paleophycus* burrows (arrows) are lined tubes and infilled with sediment similar to the surrounding matrix. Core is 8 cm wide in photograph. c-32-G 94-B-8, 1039.8 m.

b) *Planolites* burrows (arrows) in the interbedded mudstone and fine grain sandstone, lithofacies 2. Core is 5 cm wide in photograph. C-35, 324.8 m

c) Probable *Rosselia* structure (arrow) crosscut by vertical burrow. Core is 6 cm wide in photograph. a-23-A 94-B-8, 919.1 m.

d) *Skolithos* burrows (arrow) in a bioturbated sandstone, sublithofacies 5a PD 8001, 534.8 m.



a. b.

d.

PLANOLITES

Planolites burrows are unlined, rarely branching, straight to meandering, clyindrical to ellipitical in cross-section. These burrows are typically infilled with a sediment different in color and texture as that of the surrounding rock. The infill is generally structureless. The burrows are more or less horizontal but sometimes are oblique to the bedding plane (Figure 5.4b). Burrows may cross or interpenetrate. The burrows are generally preserved as endichnia, epichnia, and hypichnia. Pemberton and Frey (1982) indicate that *Planolites* may be distinguished from *Palaeophycus* by the type of infill and by the absence of a lining.

In core, the burrows are from 2 to 6 mm in diameter and are oval to circular in shape. The contrasting infill of the burrow is suggested to be partially the result of active backfilling by a mobile deposit-feeder, which is processing the sediment (Pemberton and Frey, 1982). No modern analog for the *Planolites* burrow has yet been distinguished, although several forms of polychaetes have been suggested to retain similar behavioral patterns.

Planolites is considered to be a facies breaking trace fossil that occurs in the interbedded sandstone and mudstone, silty mudstone and bioturbated sandstones of the Bluesky Formation. This type of burrow can be ethologically classified as a fodinichnia structure possibly occupied by a mobile deposit-feeder such as an annelid.

ROSSELIA

Rosselia is a conical- to bulbous-shaped structure which contains cone-in-cone laminae or helicoidal swirling spreiten structures surrounding a central stem. These structures are typically oriented vertical to inclined. The laminae of the trace indicate that the sediment was processed by a tracemaker. The fill is generally finer-grained than the surrounding rock, and may have a gnessic-like pattern in longitudinal section. The diameter at the top of the funnel is approximately 4 cm and tapers downward to < 1 cm (Figure 5.4c). These are generally preserved as endichnia.

These structures are interpreted to be the burrow of deposit-feeding vermiform animals working outwards from a central shaft (Howard and Frey, 1984). These may be classified as fodinichnia. These burrows were noted only within the finer-grained sandstones of the Peace River Plains area.

SKOLITHOS

Skolithos burrows occur as unbranc' d, cylindrical to subcylindrical shafts that are straight to slightly curve ı profile. The burrows are characteristically lined and are typically vertical to steeply inclined. The walls are generally smooth and unornamented (Figure 5.4d). The infill is structureless and composed of the same material as the surrounding host. Occasionally, the burrow infill will be extremely glauconitic w ·· a distinctive green hue. The burrows can be extremely long, penetrating deeply into the sandstones to lengths greater than 2(cm. Burrows are generally preserved as endichnia.

These single entranced shafts are interpreted to be domiciles of suspension-feeding organisms. Probable producers have been suggested to be numerous forms of polychaetes (Frey and Pemberton, 1985). Thus, these burrows toponomically represent domichnion structures.

Skolithos burrows occur in several lithofacies of the Bluesky Formation, including: the interbedded sandstone and mudstone lithofacies (lithofacies 2); the small-scale cross-bedded sandstones (l·· ·acies 3) and both sublithofacies of the bioturbated, glauconitic sandstones (lithofacies 5), although more commonly in the coarser-grained units (sublithofacies 5a). These burrows occur in both the Foothills and Peace River Plains areas. The *Skolithos* burrows in the interbedded sandstone and mudstone lithofacies appear to be distinctly smaller in size than the burrows in the bioturbated sandstones of the Peace River Plains.

TEICHICHNUS

Teichichnus burrows appear as gently curved, and closely nested, vertically stacked laminae (spreiten structures). The spreiten are typically oriented subhorizontal to oblique to the bedding plane and lie in a vertical plane. The spreiten may be retrusive (displaced upward) or protusive (displaced downward). The final tunnel records the last position occupied by the burrower. The spreite appear to be truncated at their ends. The burrows average 3 cm in width and 4 cm in height. The fill in Bluesky cores are typically finer-grained than the surrounding host. These burrows are generally preserved as endichnia.

Teichichnus is formed by the upward and downward migration of a tunnelling organism as it moves back and forth in a single vertical plane searching the sediment for food. These are considered to be produced by a worm-like deposit-feeding organism. Howard and Frey (1984) indicate that these are feeding or dwelling-feeding structures of vermiform organisms. The muddy burrow laminae surrounded by cleaner coarser-grained material has been attributed to selective size sorting of particles by the tracemaker (Howard and Frey, 1984).

Teichichnus burrows have been noted in the interbedded sandstones and mudstones of the Foothills where these structures are generally restricted to the sandstone lenses (Figure 5.5a). These typically destroy the primary bedding. *Teichichnus* is also found in the bioturbated sandstones of the Peace River Plains area (lithofacies 5).

FIGURE 5.5 CORE PHOTOGRAPHS OF ICHNOFOSSILS

a) *Teichichnus* burrow in a fine grain laminated sandstone, lithofacies 2. Burrow extends to top of sandstone bed to mudstone. Core is 5 cm wide in photograph.C-35, 319.1 m.

b) Oval-shaped, compressed *Terebellina* burrow (arrow) in a bioturbated sandstone, lithofacies 5. c-32-G 94-B-8, 1042.2 m.

c) Conglomerate filled, unlined burrows incised into Gething mudstone are *Thalassinoides* burrows. Burrows are suggested to have been emplaced in a firmground. PP 7506, 90.2 m.

d) *Zoophycos* burrow (arrow) in fine grain cross-laminated sandstone bed of lithofacies 2. Core is 5 cm wide in photograph. C-35, 367.4 m.

a.

b.

c.

d.

TEREBELLINA

Terebellina burrows are single, vertically oriented, distinctly lined tubes that commonly curved or arcuate at the base. The lining is the most distinctive feature of the burrow in that it is commonly composed of calcium carbonate or tightly packed sand grains. The linings range from 1 to 3 mm in thickness. The diameters of the tubes are characteristically small, averaging 4 mm. The burrows usually taper in diameter downwards. In cross-section, the burrows are typically doughnut-shaped with the infill darker colored than the lining, or the tubes are often flattened to an oval shape. The burrows are preserved as endichnia. These burrows are similar in morphology to the ichnogenus *Schaubcylindrichnus*. However, the latter may be recognized by the multiples or bundles of tubes. In core the burrows appear white, ovoid burrows which seem to be predominately horizontal (Figure 5.5b). However, this is the result of the strong curvature at the base of the tube.

These burrows are considered to be the dwelling structures of worm-like suspension-feeders, such as a polychaete. Thus, these may be toponomically classified as a domichnia structure.

Terebellina has been noted most commonly in the bioturbated, fine-grain sandstone lithofacies (sublithofacies 5b) in the Peace River Plains area and rarely in the interbedded sands and muds lithofacies (lithofacies 2) of the Foothills area.

THALASSINOIDES

Thalassinoides structures are large, clyindrical burrows which form three-dimensional branching systems which are connected to the surface by a shaft. The branches are 'Y' or 'T' shaped and are typically enlarged at the points of bifurcation. The burrows are thinly-lined to unlined where reinforcement is not necessary. The burrows are circular to oval in cross-section and greater than 2 cm in diameter. Infill is generally different than the surrounding sediment. The infill material indicates passive (gravitational) sedimentation after evacuation by the organism. *Thalassinoides* burrows are differentiated from *Ophiomorpha* by the absence of an ornamented wall lining. These burrows are preserved as epichnia, endichnia and hypichnia. *Thalassinoides* burrows are considered to be the dwelling burrow of a shrimp or shrimp-like organism.

In Bluesky cores, these burrow networks extend downwards from the Bluesky-Gething formational contact into the mudstones of the Gething Formation. They are typically sharp-walled, unlined and infilled with chert pebbles and sand from the overlying conglomerate (Figure 5.5c). The lack of lining in these burrows indicates that they may hacve been emplaced into a semi-cohesive substrate or 'firmground'.

ZOOPHYCOS

Zoophycos is composed of a complex, sheet-like downwards spiral or whorl of spreiten structures. The diameter of sucessive whorls or volutions generally increase downwards. These whorls are either flat or curved and wind about a central cylindrical axis. The overall shape is analogous to the shell of a gastropod. The laminae contained within each whorl may be curved or arcuate, with a small ridge. The central stem is presumed to open to the sediment surface. These structures are preserved as endichnia.

In core, the burrow generally appears as a long, sediment back-filled tunnel, with no indication of successive whorls (Figure 5.5d). Generally, the tunnels are 4 cm in length and 5 mm in height. The infill is comparatively close to that of the surrounding host.

These traces are interpreted to be complex feeding patterns of vermiform organisms where each whorl represents the path of the feeding proboscis. The complex form may indicate a impoverished food supply or relatively calm to restricted waters. Thus, these traces may be classified as feeding structure or possibly a grazing-feeding structure (fodinichnia/paschichnia) possibly occupied by an annelid.

These traces have primarily found in the interbedded sands and muds lithofacies (lithofacies 2) of the Foothills area. Generally these are restricted to the sandstone lenses.

5.4.2 *ICHNOFOSSIL ASSOCIATIONS*

The designation of trace fossil suites is based strictly on the diversity, abundance and distribution of ichnogenera in Bluesky cores. The overall occurrence of trace fossils in Bluesky cores can be divided into at least six distinct, recurring ichnofossil associations (Table 5.3). These ichnofossil associations represent the repeated occurrence of an assemblage of traces in a substrate when the required ecological conditions were met (Frey and Seilacher, 1980). However, because of temporal environmental variations, individual ichnogenera may be transitory, occurring locally in associations other than those designated. Therefore, the general character and ethological trends displayed by a suite is of more importance than the presence or absence of an individual ichnotaxon (Frey and Seilacher, 1980). Establishing a specific behavioral trend of the ichnocoenose is of paramount importance when inferring some conditions of deposition. When combined with the physical sedimentary evidence an environmental reconstruction is much closer at hand. The overall assemblage of traces in Bluesky sediments point to a shallow marine environment. The following discussion will provide a description of the ichnofossil associations within the Bluesky Formation plus an analysis of their paleoenvironmental significance.

Table 5.3 : Distribution and relative abundance* of ichnofossils comprising the ichnofossil suites in Bluesky cores.

Ichnogenus	Ichnofossil Suite					
	1	2	3	4	5	6
<u>Dwelling Burrows</u>						
Diplocraterion	C					
Ophiomorpha	O					
Palaeophycus	A	A	R	C		
Skolithos	A			C		
Terebellina			A			
<u>Feeding Burrows</u>						
Asterosoma	R	O				
Chondrites		R		O		O
Cylindrichnus	O					
Gyrochorte				O		
Planolites	R	A		A		A
Rosselia		O				
Teichichnus	O	O		C		
Thalassinoides					A	
<u>Grazing Structures</u>						
Helminthopsis		A	A	A		A
Zoophycos				O		

* Relative Abundance: R = Rare O = Occasional C = Common A = Abundant

Ichnofossil Suite 1. This suite is dominated by several varieties of dwelling burrows, including *Skolithos*, *Palaeophycus*, *Diplocraterion* and *Ophiomorpha*. However, rare dwelling/feeding structures of *Asterosoma*, *Cylindrichnus*, *Teichichnus* and *Planolites* are common locally. These traces are associated with the heterolithic glauconitic, coarse-grain sandstone (sublithofacies 5a). Figure 5.6 illustrates the relative proportion of each trophic group which comprises this suite of trace fossils. The asssemblage of traces in this suite are predominately vertically-oriented with lined burrow walls. The density of burrows is very high. Producers of the dwelling burrows have been interpreted to be suspension feeders (some polychaetes and crustaceans) which gather food from the overlying wave-agitated waters. Such an assemblage of trace fossils with thick burrow wall-linings is indicative of high energy conditions and the shifting, granular nature of the substrate (Howard and Frey, 1984).

The high density of traces, however, may indicate a somewhat slow depositional rate where bioturbation is in pace with sedimentation, rather than strictly a high abundance of biota. Figure 5.7a provides an example of the occurrence of ichnofossil suite 1 in core.

Ichnofossil Suite 2. Suite 2 is commonly associated with fine-grained, glauconitic sandstones (sublithofacies 5b) and minor siltstones. As described in Chapter 4, these sandstones are generally muddier and more poorly sorted than the lithofacies in the above suite. This suite is characterized by the dwelling structure *Palaeophycus* as well as a host of feeding, grazing and combined feeding-dwelling structures, including: *Asterosoma*, *Helminthopsis*, *Planolites*, *Rosselia*, *Teichichnus*, *Terebellina* and rare *Chondrites* (Table 5.3). To the casual observer the core seems to be dominated by *Palaeophycus*; however, the rock typically has a intense bioturbate texture where myriads of indistin feeding structures have completely churned the sediments. Therefore, the larger dwelling burrows may seem overemphasized or preferentially

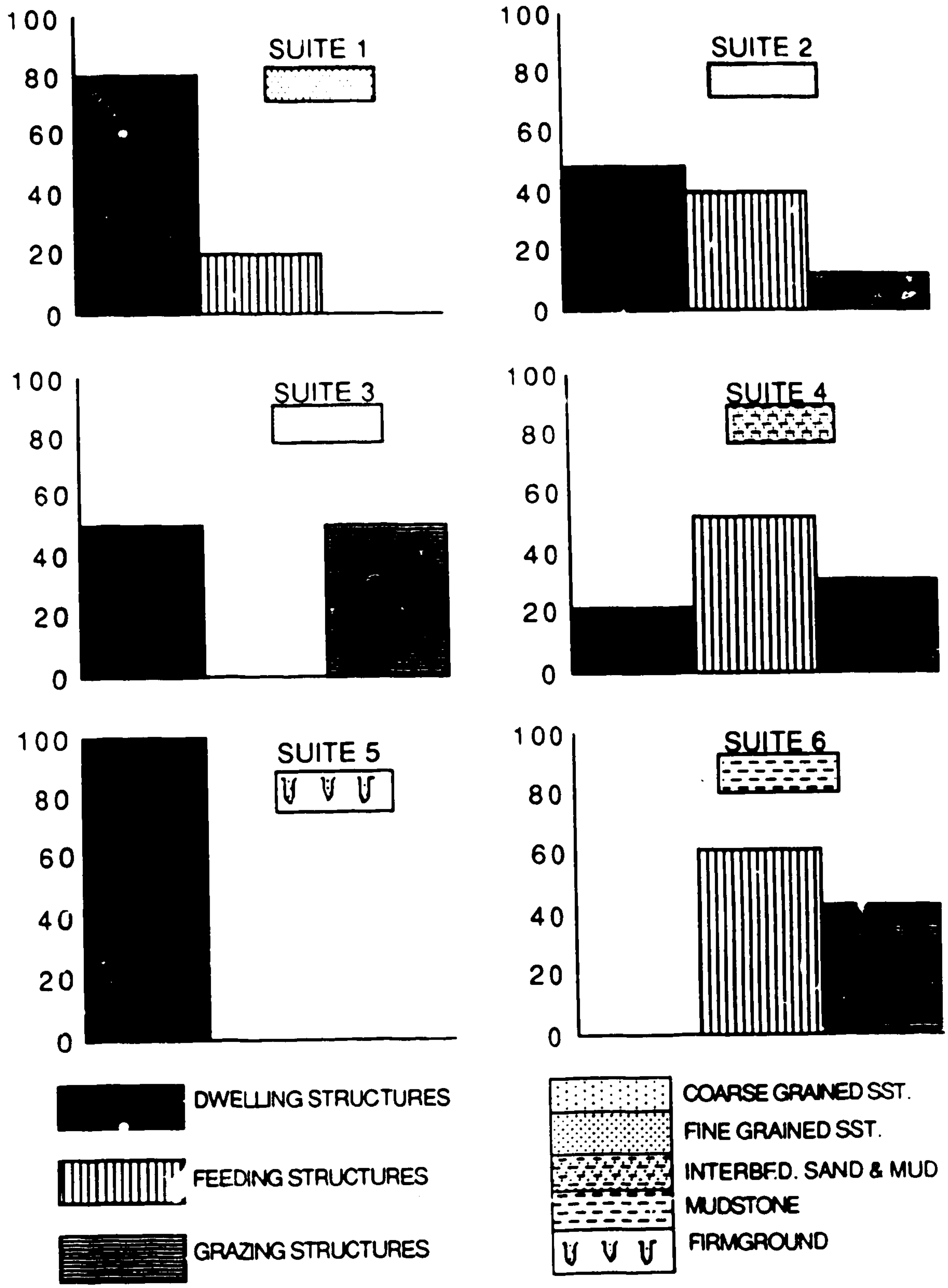


FIGURE 5.6: RELATIVE PROPORTION OF EACH TROPHIC GROUP IN ICHNOFOSSIL SUITES 1-6

FIGURE 5.7: CORE PHOTOGRAPHS OF TRACE FOSSIL SUITES

a) **Ichnofossil Suite 1**: Suite 1, in a medium-grain bioturbated sandstone (sublithofacies 5a), is characteristic of the *Skolithos* ichnofacies. Single arrows point to *Diplocraterion* burrows and double arrow points to a *Palaeophycus* burrow 6-29-81-15, 1094.7 m.

b) **Ichnofossil Suite 2**: Suite 2 contains trace fossils characteristic of the *Cruziana* ichnofacies. This suite of trace fossils occurs in a fine-grain, bioturbated sandstone Arrow points to a *Palaeophycus* burrow PD 8001, 538 m.

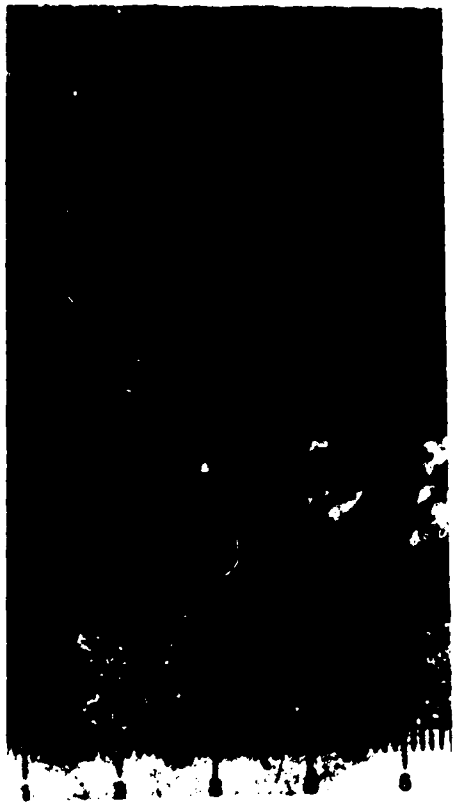
c) **Ichnofossil Suite 3**: Suite 3 contains trace fossils characteristic of the *Cruziana* ichnofacies. Compressed *Terebellina* burrows are indicated by arrow. This suite of trace fossils is considered to occur in somewhat restricted circulation conditions. PD 8001, 557.1 m.

d) **Ichnofossil Suite 4**: Suite 4 occurs in the interbedded mudstone and fine-grain sandstone lithofacies. Arrow points to a *Skolithos* burrow. C-35, 354.5 m.

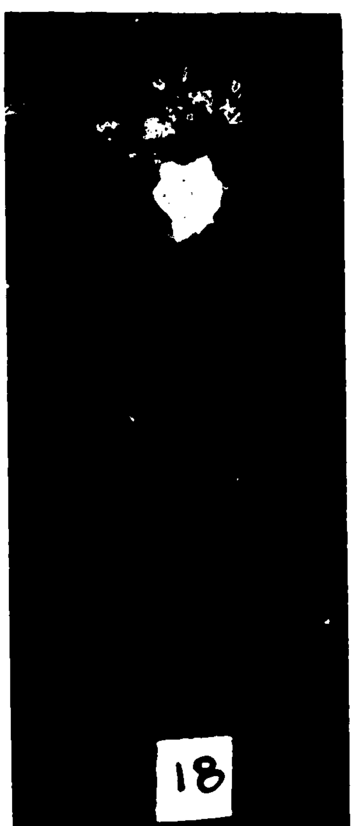
e) **Ichnofossil Suite 5**: Network of conglomerate infilled burrows (arrows) occur at the base of the Bluesky Formation. Dotted lines outline stratification in mudstone of the Gething Formation. This suite of trace fossils is characteristic of the *Glossifungites* ichnofacies. c-32-G 94-B-8, 1042 m.

f) **Ichnofossil Suite 6**: *Planolites* burrows (arrow) in silty mudstone comprises part of Suite 6. This suite of trace fossils is indicative of the *Cruziana* ichnofacies, possibly in an offshore-like setting.


a.


b.


c.



d.


e.


f.

preserved; hence, feeding structures are possibly more abundant than initially considered. Figure 5.6 illustrates the relative proportion of each trophic group that comprises suite 2. Overall, the rate of bioturbation is extremely high.

The assemblage of traces contained within this suite demonstrates a distinct shift in strategies as compared to suite 1, from mostly ve[illegible] horizontal components (Figure 5.7b). The extreme rate of bioturbation, and combined feeding-dwelling structures suggests that sediments were probably less shifting with slower sedimentation rates and with a high degree of deposited food.

Ichnofossil Suite 3. Association 3 is closely related to suite 2 which is similarly common to the fine-grained, glauconitic sandstones. Only a few trace fossils, representing two distinct types of behavior, were recognized. The grazing trace *Helminthopsis*, the dwelling burrow *Terebellina* and small *Palaeophycus* structures represent a low diversity-high abundance suite. The smaller number of recognizable traces is more likely related to obscuring of other forms by a high degree of biogenic reworking, rather than a dearth of organisms.

The dominance of grazing traces indicates that organic detritus was present on the seafloor in quantities sufficient to support the grazers, possibly in large numbers. The trace fossil *Helminthopsis* is often found in the deeper, quieter waters of the offshore (Ekdale *et al.*, 1984). *Palaeophycus* and *Terebellina* are dwelling burrows which were most likely occupied by suspension-feeders or predaceous vermiform organisms. An interesting morphological feature of the *Terebellina* burrows, which has some environmental significance, is their compressed oval shape (Figure 5.7c). The compression of the *Terebellina* burrows is somewhat analogous to the deformation of *Palaeophycus* in soupgrounds (Pemberton and Frey, 1982). Softgrounds such as these, occur in quiet but not totally restricted environments (Ekdale *et al.*, 1984). These burrow forms when emplaced in a fluidized medium become compressed due to the weight of the overlying sediment column. It has also been noted that *Terebellina* burrows found in the upper reaches of the shoreface maintain their round cylindrical shape; whe as in deeper-waters and finer grained sediment, the burrows are typically smaller and compressed (Tillman and Martinsen, 1979).

Ichnofossil Suite 4. This assemblage of trace fossils is the most intriguing of the Bluesky cores, occurring in the coarsening-upward interbedded dstone and sandstone unit (lithofacies 2). It is typically dominated by simple feeding structures of *Planolites*, *Teichichnus* and *Chondrites*, and by grazi ctures of *Helminthopsis* and rare *Zoophycos* (Table 5.3). Scattere hroughout the occurrence of this suite are associated forms of the dwelling structures *Skolithos* and *Palaeophycus*. Figure 5.6 illustrates the relative proportion of each trophic group within this suite. The *Skolithos* and *Palaeophycus* burrows are generally contained within the sand units. Commonly the traces are generally small in size, for example, the spreiten structures of *Teichichnus* are relatively narrow and the shaft length and tube diameter of *Skolithos* are relatively small when compared to species found in ichnofossil suite 1 (Figure 5.7d). The degree of bioturbation ranges from low to moderate to locally high. Numerous beds are characterized by a bioturbate texture where discrete ichnofauna are impossible to identify. In areas where *Teichichnus* burrow density is high, a bioturbate texture exhibiting spreiten structures is reconizable only. No single bed records a mutual ubiquitous occurrence of all the trace fossils oberved in this assemblage. Of the seven dominant forms only *Teichichnus* and *Planolites* occur consistently throughout the lithofacies. Traces such as *Skolithos* and *Palaeophycus* become progressively more abundant vertically upwards, whereas *Zoophycos* disappears in the upper sections of the core.

Taken collectively, this assemblage is dominated by horizontal, deposit-feeding structures that are exploiting relatively nutrient-rich, fine-grain sediments. The dwelling structures are generally restricted to the thin, sharp-based, cross-laminated sandstone lenses. The occurrence of dwelling structures may indicate the displacement of the indigenous benthic community by opportunists as a result of a physiologically stressful event. Overall from an ethologic context, this assemblage is characterized by a generally low diversity of traces, an impoverished marine assemblage, a reduction in overall burrow size, and a dominance of functionally simple feeding strategies.

Ichnofossil Suite 5. This is a monosp[illegible] assemblage consisting of a network of burrows that have been tentatively identified as *Thalassinoides* (Table 5.1). The burrows are closely-spaced dwelling structures that characteristically have sharply defined and un[illegible]ed walls. The burrows are typically passively infilled, with sand and pebbles from the immediately overlying strata, after after desertion of the burrow (Figure 5.[illegible]). Commonly these burrows are incised into the mudstones beneath the Gething-Bluesky formational contact or in sand units immediately beneath matrix-supported conglomerates (sublithofacies 1a). The mudstones locally contain small feeding structures of *Planolites*.

The tracemakers which constructed these permanent domiciles are suggested to be suspension feeders or carnivores. The lack of reinforced walls for burrows incised in mudstones indicates that the sediment may have been fairly cohesive at the time of emplacement. Sediment cohesiveness generally discounts the need for burrow linings (Ekdale *et al.*, 1984). In addition, tube walls are suggested to be utilized by an organism to isolate itself from intruding ambient pore water solutes within the sediment (Rhoads, 1975). If the sediment is dewatered, as firmgrounds generally are, the need for the organism to protect itself with wall linings would no longer be necessry. Thus, these dwelling burrows possibly represent overprinting of bioturbation into a relict substrate that was possible partially exhumed by erosive currents.

Ichnofossil Suite 6. Suite 6 is a low diversity as[illegible]mblage dominated by the fee[illegible]ing structures *Planolites* and *Ch[illegible]ndrites* and by the grazing structure *Helminthopsis*. This assemblage is consistently asso[illegible]ated with the silty mudstones (lithofacies 8) (Table 5.3, Figure 5.6)

The diversity of burrows is very low [illegible] burr[illegible]w density is high. At times it is difficult to recognize distinct burrow forms an only a [illegible] [illegible]re is discerned. This assemblage of traces exclusi[illegible] [illegible]ts the activities of a community of deposit-feeding organisms (Figure 5.7f). Intens[illegible] [illegible]on and the ethologic type indicates slow, continuous rates of deposition. The dearth [illegible]ling burrows suggests that most food particulates were deposited rather than suspended, indicating that little wave or current activity was present.

5.4.3 *GEOLOGIC SIGNIFICANCE OF TRACE FOSSIL SUITES*

Collectively, the complete asse[illegible]blage of ichnofauna present in the Bluesky cores indicate that normal marine and [illegible]arginal marine settings were responsible for the deposition of the Bluesky Formation. However, separately the different ichocoenoses reflect varying behavioral resp[illegible]ses [illegible]y the organisms to distinct subenvironments or niches. Overall, suites 1, 2, 5, and 6 are typical of normal marine conditions. Suite 3 reflects somewhat restricted conditions and suite 4 characterizes a [illegible]ackish water setting.

In general, most assemblages in the Bluesky Formation, are consistent with the energy-depth relationships exhibited by the primary lithofacies. Specifically, the behavioral characteristics of each ichnocoenose reflects the intensity of current energy associated with the deposition of each lithofacies. The predominance of dwelling traces associated with a coarse-grain sandstone in suite 1 represents the highest energy levels. This is followed by the trophically diverse suite 2 in a fine-grain sandstone reflecting a response to lower energy conditions. The deposit-feeding traces of of suite 4 and 6 in mudstones reflect the lowest energy levels.

Each of these suites is also representative of a specific ichnofacies as defined by Seilacher (1967a). These are

a) *Skolithos* Ichnofacies: Suite 1

b) *Cruziana* Ichnofacies: Suites 2, 3 ,

c) Brackish Water Ichnofacies: Suite 4

d) *Glossifungites* Ichnofacies: Suite 5

The trace fossil assemblage of Suite 1 is characteristic of the *Skolithos* ichnofacies. The ichnogenera contained within the assemblage reflect activities of a trophically-narrow suite comprised predominantly of carnivores, epifaunal and infaunal suspension-feeders combined with several deposit feeders. The behavioral patterns indicate moderate to high energy levels, with water agitation sufficient for the requirements of suspension-feeders, but weak enough to allow settling of finer organic detritus to maintain the sediment-feeders. The high degree of bioturbation indicates either slow periods of sedimentation or high faunal density. Freely shifting substrates appear to have been present as indicated by the presence of thick burrow wall linings. In addition, the long tubes of suspension-feeders may be interpreted as a protective measure shelter against unstable substrates.

The occurrence of numerous beds of parallel-laminated sandstones within the burrowed segments reflects a dynamic and episodic nature to sedimentation within this unit. Periodic additions of sediment into the deposit has produced some thick cycles of laminated- to burrowed-sequences. These sequences are especially pronounced in the more shoreward sand bodies of the Bluesky Formation in the Peace River Plains area. In the farther offshore sand bodies of the Bluesky Formation, these distinctive sequences are less pronounced. Howard (1972) noted that modern shoreface sequences also contained laminated to burrowed sediment. Howard interpreted these sequences as alternating periods of storm and fairweather sedimentation just below fairweather wave base. These sequences were also recorded in the ancient shoreface deposits of the Star Point and Blackhawk Formations of Utah (Howard and Frey, 1984).

A sedimentological study of the Gulf of Gaeta (Dorjes and Hertweck, 1975) documented the ichnological zonations found in a beach to offshore sequence. The Gulf of Gaeta is a high energy, microtidal beach dominated by storm- and wave-induced currents. Dorjes and Herweck (1975) documented abundant burrows of the organism *Capitomastus minimus* and lined dwelling tubes of the polychaetes *Diopatra neapolitana*, *Onuphis eremita* and *Ouemia fusiformis* in the lower shoreface at water depths from 2-5 m. Lithologically, this zone contains interbeds of bioturbated and ripple- [illegible] allel-laminated sand.

Even though this setting may be dissimilar in depositional setting to that of the Bluesky Formation, it does provide some insight as to the environment that contains the traces representative of the *Skolithos* ichnofacies. The high faunal density and the predominance of dwelling tubes with reinforced walls in Suite 1 is similar to that of the Gulf of Gaeta. Thus, it would seem possible that these trace fossils were emplaced in a high energy setting with freely shifting substrates, such as that found in a prograding shoreface. However, the association of deposit-feeders in Suite 1, such as *Asterosoma*, in a clean sand indicates more of an offshore sand origin.

Association 2 is a trophically fairly uniform suite of trace fossils. Traces such as *Helminthopsis*, *Planolites*, *Teichichnus*, *Asterosoma* and *Rosselia* were made by deposit feeders indicating that the organic content of the sediment was relatively high, and hence, energy levels were generally low. However, the occurrence of many *Palaeophycus* burrows suggests an environment where water agitation was sufficient to support the requirements of these tracemakers.

Overall, the shift in feeding strategies from suspension- to deposit-feeding structures suggests a distinct decrease in energy levels as compared to Suite 1. Collectively, the trace assemblage of Suite 2 is characteristic of the *Cruziana* ichnofacies. This is supported by the well-preserved and intense level of bioturbation, the exhibited behavioral patterns and lithologically, by fine-grained sandstones, and increased mud content.

A close analogy to this setting is found in the upper offshore zone of the Georgia coastline where modern sediments are intensely bioturbated (Dorjes and Hertweck, 1975; Howard and Reineck, 1972, 1981). Traces recorded from the recent upper offshore zone are mostly polychaete burrows and tubes of crustaceans and vagile carnivores (Dorjes and Hertweck, 1975). The sediments are characterized by abundant burrows of the polychaetes *Glycera*, *Capitomastus*, and the stomatopod crustacean *Squilla*. Water depths for this zone range from 5 to 10 m and the sediment is composed of silty sands (Howard and Reineck, 1981). The level of bioturbation and density of burrowing fauna in the offshore are considered to be the highest in the neashore profile (Dorjes and Hertweck, 1975, Howard and Reineck, 1981). Generally, the preservation of stratification within this zone is minor (Howard and Reineck, 1981).

The diversity and density of ichnofossils in Suite 2 occurring in a fine-grain sandstone resembles that of the upper offshore of the Georgia coastline. The extreme amount of bioturbation, the dearth of physical structures and the wealth of feeding structures indicates an overall low energy setting with stable substrates. This trace fossil suite appears to be depositionally adjacent to, but bathymetrically deeper than Ichnofossil Suite 1. Suite 2 possibly represents a slightly more distal position in an offshore deposit than suggested for Suite 1.

Trace Fossil Suite 3 is characterized by a mixture of grazing structures (*Helminthopsis*) and dwelling structures (*Terebellina* and *Palaeophycus*) in a fine-grained, structureless sandstone. This trophically-diverse ichnocoenose reflects the activity of carnivores, suspension-feeders and deposit-feeders. The high degree of bioturbation and paucity of escape structures indicates slow, continuous rates of deposition. The diversity of feeding behaviors in this assemblage indicates somewhat intermediate environmental conditions, where water agitation was high enough for the requirements of suspensions-feeders, but low enough to allow settling of the fine sands and organics essential for the deposit-feeders.

This suite of traces is characteristic of the *Cruziana* ichnofacies. This ichnofacies is typical to lagoons, offshore zones and tidal flats (Ekdale *et al.*, 1984). However it differs somewhat from the above trace fossil suite (Suite 2), which also is characteristic of the *Cruziana* ichnofacies, by the type of ichnotaxa present and the nature of the associated sediment. Trace Fossil Suite 3, as mentioned above, occurs in a much cleaner, homogeneous sand. In many respects, it resembles the sand that is typical of washover deposits of barrier islands (c.f. Reinson, 1984), deposited mainly from suspension rather than traction. This may explain, in part, the structureless nature of the sediment. The compressed nature of the *Terebellina* burrows is also very significant. Generally this form of deformation has been attributed to restricted circulation conditions (S.G. Pemberton, 1986, per. comm.). A notable occurrence of this association is found in the well PD 8001, in Township 81, Range 20 W6 of the study area, at the base of a thick sandstone accumulation. This would imply that this ichnofossil suite occurs in a somewhat sheltered area.

where the prevailing currents were possibly being damped. In short, this suite probably reflects somewhat restricted circulation conditions in an offshore setting.

Trace Fossil Suite 4 is considered to represent an impoverished assemblage of traces that is associated with the interbedded sands and muds of lithofacies 2. This suite is dominated by feeding structures *(Planolites, Teichichnus,* and *Helminthopsis)* as well as associated forms of dwelling structures *(Palaeophycus* and *Skolithos)*. The combination of feeding and minor dwelling structures indicates that some elements of both the *Skolithos* and *Cruziana* ichnofacies are present. In addition, the overall reduction in ichnotaxonomic diversity, the simplicity of the feeding structures and generally reduced size of the traces suggests a somewhat physiologically-stressful environment (Wightman *et al.*, 1987; Benyon *et al.*, 1988). Variations in salinity is consi[illegible]d to be a major factor influencing the distribution of benthic marine faunas (Rhoads, 1975; Wightman *et al.*, 1987). Salinities deviating fr[illegible] normal marine levels are generally found in nearshore environments where large inputs of freshwater o[illegible] exchange with fully marine waters result in brackish conditions (Dorjes and Howard, 1975; Fursich and Werner, 1984). Environments such as this are rarely on a basin-wide scale, but more confined to settings such as estuaries, tidal flats, and delta platforms which suffer short-term fluctuations.

Recognition of salinity influenced settings is largely based on the contained fauna (Fursich and Werner, 1984). Apart from the reduced diversity levels, salinity reduction is inferred from the mixture of feeding strategies. In situations such as these, mud-loving organisms that produce feeding structures would be juxtaposed with dwelling structures constructed in the sand units (Ekdale *et al.*, 1984). Organisms common to these environments would be trophic generalists, employing nonspecialized feeding strategies, and where infaunal habitats are more abundant than epifa[illegible] [illegible]hould also be stre[illegible] that the relative abundance of traces found within brack[illegible] water [illegible]posits will be reduced (Dorjes and Hertweck, 1975). To examine the animal-sediment relationship in a brackish setting, a marginally-marine setting in the Norton Sound is used to possibly better illustrate the environmental significance of the biologic and physical trends displayed by this assemblage.

The Norton Sound is a broad, micro-tidal re-entrant of the Bering Sea, characterized by low rates of tectonic subsidence and extremely shallow water depths (<25m) (Howard and Nelson, 1982). The Yukon Delta, on the southern margin of the Sound, is the major source of sediment that enters the Sound, mostly during the ice-free months of May through October. The delta of the Yukon River is relatively young and consists of pro-delta, delta-front and delta plain components (Dupre, 1982). Although the Yukon River transports large quantities of silt and sand into the delta region, much of it is resuspended during storm events and transported away by the Alaskan Coastal Water Mass, thus, bypassing the Norton Sound and Bering Sea area (Howard and Nelson, 1982). What sediment is stored in the delta area is reworked by a variety of processes including waves, wind and tidally-induced currents, strong oceanic currents, as well as ice movement (Dupre, 1982). Thus, the depositional setting is said to be river- and storm-dominated (Dupre, 1982).

Seasonality and spring freshwater run-off is very much apparent in the Yukon River and, as such, the Norton Sound displays rapid salinity fluctuations. During the summer months, excessive run-off of freshwater creates a salinity zonation within the Sound, the degree of salinity increases from the delta platform seaward towards the northern end of the sound where the salinity is almost normal (Howard and Nelson, 1982). Proximal-distal trends are also demonstrated in the sedimentary structures. Physical Structures are best developed on and adjacent to the Yukon Delta, including parallel-, and ripple-laminated beds (Howard and Nelson, 1982). Dominance of physical structures near the Yukon Delta are the result of: 1) very shallow water wave and current energy, 2) increased rates of deposition, and 3) reduced rates of [illegible]

Nelson, 1982). They also stated that bioturbation patterns also reflect the salinity gradient with a high density, normal marine assemblage at the north end and a low-density, restricted assemblage at the delta platform (Figure 5.8).

The effects of storm-surge depositional processes that occur yearly from mo[illegible] severe storms are also prominent within the Sound (Nelson, 1982a). As discussed in Ch[illegible] on shelf processes, storm surges result in graded sand layers interbedded with biotur[illegible] muds, up to 100 km from the Yukon Delta shoreline (Nelson, 1982a). Prox[illegible]al-distal tr[illegible] these [illegible] layers were noted, including: 1) coarser, thicker and more vertically comple[illegible] sequences of laminations (parallel, trough, and cross) occur in the proximal zones, and 2) distally, the sand layers change to thin silt beds with flat and ripple-drift lamination (Nelson, 1982a).

The modern Yukon Delta in some respects serves as a biologic model for this Bluesky trace fossil suite as well as the sediment type and distribution for lithofacies 2 (interbedded sandstone and mudstone). The Norton Sound contains a shelf prodelta sequence accumulating in a somewhat restricted embayment, in which sediments from the Yukon delta are redistributed by storm-events. [illegible] Norton Sound displays a basinward increase in bioturb[illegible] and salini[illegible] a basinward decrease in grain size and sedimentary structures.

These same trends are exhibited in Association 4 of the Bluesky cores. This association, which was subject to progressive reductions in salinity vertically upwards, reflects laterally and vertically the regressive nature of this sequence. Since this stratigraphic package is overlain by Chamberlain delta sediments, it seems likely that Trace Fossil Suite 4 reflects a prodelta delta front setting. This in itself is important because it signals a change in basin depositional sty[illegible] from predominately marine in the northern half of the study area to marginal marine followed by continental setting in the southern half of the study area.

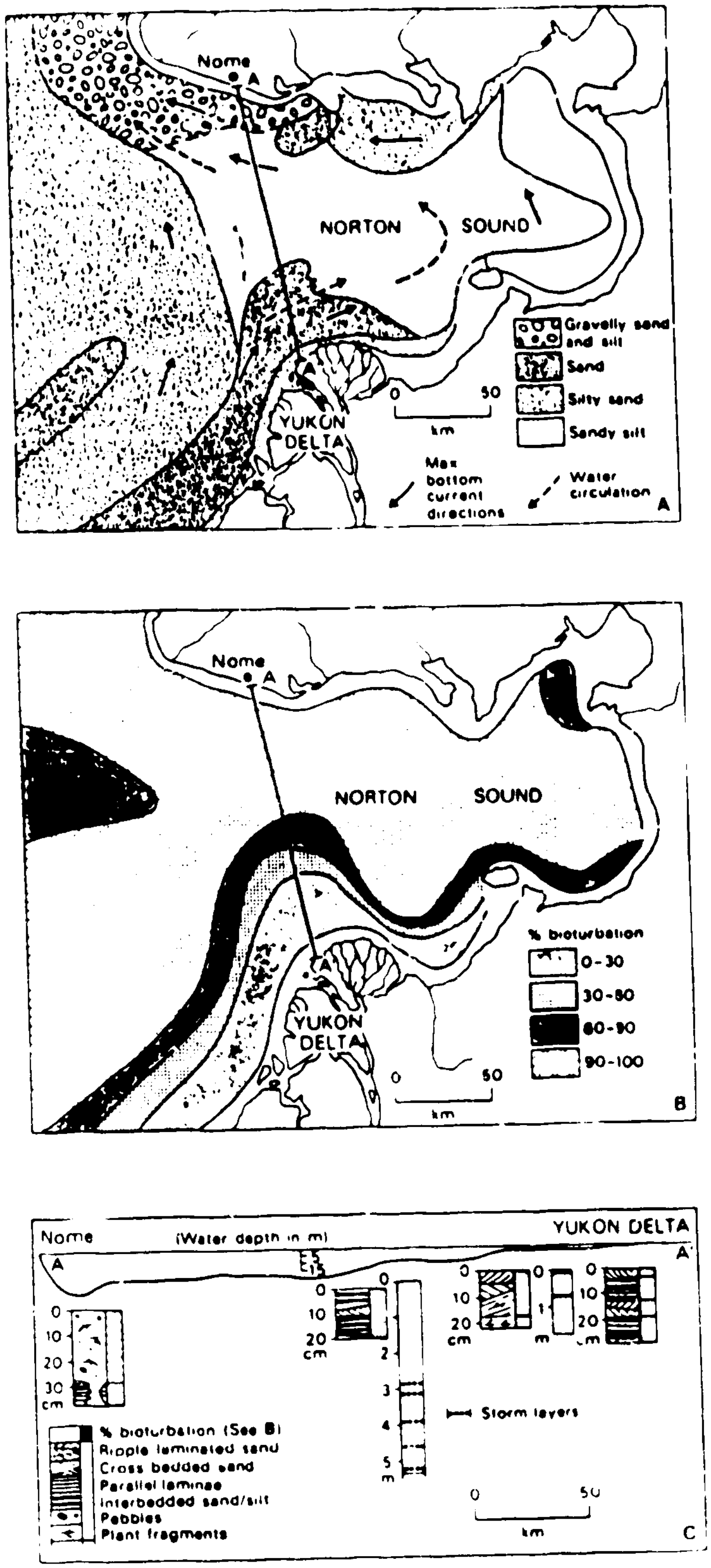


Figure 5.8 Sedimentation and bioturbation in the delta-influenced shelf of Norton Sound region of the N.E. Bering Sea. (A) Generalized sediment distribution (B) Distribution of bioturbation (C) Physical sedimentary structures and storm layers in a transect along A-A' (modified from Howard and Nelson, 1982).

Trace Fossil Suite 5, albeit monospecific, is characteristic of the *Glossifungites* ichnofacies. Substrates of the *Glossifungites* ichnofacies are typically firm, but unlithified, compacted muds, or silts which tend to develop in protected, low-energy settings, or high-energy settings where the sediment offers some resistance to erosion (Pemberton and Frey, 1985). Trace fossils characteristic of this ichnofacies consist of unlined dwelling structures of suspension-feeding or carnivores, such as bivalves, crustaceans, and polychaetes (Pemberton and Frey, 1985).

Although this suite is typically found in other ichnofacies, it is the mode of occurrence and preservation that indicates the *Glossifungites* ichnofacies. These unlined, vertical burrows typically occur at an erosive contact between the Gething and the Bluesky Formations. The contact is generally disconformable and consists of mud deposits overlain by matrix-supported conglomerates of sublithofacies 1a. This discontinuity surface represents a break in deposition and a dramatic shift in depositional regimes (Bromley, 1975). The Gething mudstones locally contain intrastratal feeders such as the *Planolites* burrows, which are overprinted by the *Thalassinoides* burrows. This association indicates that the Gething mudstones were possibly semi-cohesive firmground and that they retained relict features indicative of prior conditions. This situation is somewhat analogous to the preomission- and omission-suites described by Bromley (1975). The burrows contained within the Gething muds may be termed the preomission assemblage and the *Thalassinoides* burrows constitute the omission assemblage.

The fill of the vertical burrows is very significant since it is composed of the same pebble and sand mixture as the overlying matrix-supported conglomerate rather than Gething Formation sediments. This suggests that the burrows remained open in the sea floor through the initial stages of the processes which emplaced the conglomerate. Both the *Thalassinoides* burrows and the occurrence of conglomerate suggest a shift in environmental parameters with increased current energies.

One notable example of modern sediments, which may aid in the interpretation of this suite and associated sediments, is the Georgia Coast (Pemberton and Frey, 1985). At many locales along the Georgia coast, active erosion storm turbulence and the late Holocene transgression has exposed relict salt marsh deposits which have been subsequently overprinted by organisms under the new depositional regime (Pemberton and Frey, 1985). Thus, this sedimentary sequence records a relict soft-substrate assemblage of traces as well as a modern suite of opportunists in a high-energy firm substrate suite which is characteristic of the *Glossifungites* ichnofacies (Pemberton and Frey, 1985).

Suite 5 of the Bluesky cores possibly exhibit similar hydrographic characteristics as the Georgia setting. Unlithified, but dewatered, shallow buried muds of the Gething Formation were possibly exhumed or reworked by the transgressing Boreal Sea. The high-energy nature of the transgressive currents allowed opportunistic organisms to occupy the exposed firm-substrate briefly before the eplcement the conglomerates. This aspect in itself has important implications for the origin of the matrix-supported conglomerate (sublithofacies 1a) in that the burrows must have been emplaced under saline conditions.

Trace Fossil Suite 6, which is characterized by a low-diversity assemblage of traces *(Planolites* and *Helminthopsis)*, is typically associated with fine-grain deposits. In this case, the traces record markings, organic harvesting or castings of ingested sediments by intrastratal feeders. Burrows are all horizontally oriented and occur in relatively high abundance. The sediment itself records very few, if any, current structures.

From the biologic and lithologic evidence, it can be concluded that this was a relatively low-energy environment with little indication of wave or current influence. The sediments were probably rich in organic matter in order to support these active deposit-feeding organisms. The paucity of suspension-feeders is probably due to the low-level of water turbulence, hence little suspended food, or the overlying water column was too turbid and thus, inhospitable to filter-feeding (Ekdale *et al.*, 1984).

Sedimentation rates were also probably slow and continuous. This is assumed by the absence of sand layers from pulses of sedimentation or storm-events and the extensiveness of bioturbation.

Overall this suite is indicative of the middle to lower offshore conditions and the *Cruziana* ichnofacies. However, the trace fossils and its associated sediment type are typically found interbedded between deposits of bioturbated, glauconitic sandstones. This stratigraphic association suggests that there were temporal changes within the depositional setting rather than major events of transgression and significant deepening of the water column. These conditions in the nearshore zone must be produced by restrictions in circulation which allows deposition of suspended fines. Situations such as these can occur in the troughs of large bedforms or in tidal slack water periods. However, considering the thickness of these units (commonly >10 cm), a tidally-influenced genesis is unlikely.

6.0 PALEOGEOGRAPHY AND DEPOSITIONAL ENVIRONMENTS

The Bluesky Formation in the Peace River Plains and the Foothills areas of northeastern British Columbia records two distinctly different depositional settings of marine and marginally marine sedimentation. The lithofacies succession occurring in the Peace River Plains region is interpreted as shelf bar complex deposited in relatively shallow, open marine waters as a result of the rapid southward transgression of the Boreal Moosebar sea. On the basis of its constituent lithofacies, ichnofaunal content and stratigraphic position, it is considered to have been deposited in an offshore, inner-neritic setting. The orientation and pronounced lenticular geometry of the sandstone bodies indicates some form of bar buildup subparallel to a paleoshoreline. The lithofacies succession from the Sukunka to Wapiti Rivers, on the other hand, consists of nearshore progradational prodelta/delta front sediments deposited in a brackish water embayment prior to the deposition of the Chamberlain delta. These deposits formed at or near the same time as the shelf bar deposits. The paleogeographic setting of the Bluesky Formation (Figures 6.1 and 6.2), including the Chamberlain delta, illustrates the position of the shelf bars as well as the progradational limit of the delta front sediment. This interpretation of depositional environments coincides, with some modifications, to prior interpretations made by several authors such as Smith *et al.*, (1984) and Jackson (1984).

Despite inherent local variations, the general vertical trend through the deposits in the Peace River Plains area is one of coarsening-upward sand sequence, accompanied by a shift in the type of bioturbation and increased development of stratification. A typical vertical section through the shelf bars show a grading upwards from muddy bioturbated sandstones into burrowed and cross-laminated sandstones, and capped by non-burrowed, cross-stratified sandstones (Figure 6.3). Overall, the diversity and intensity of bioturbation indicates a marine environment with open circulation, normal levels of salinity, and stability, in which bioturbation kept pace with

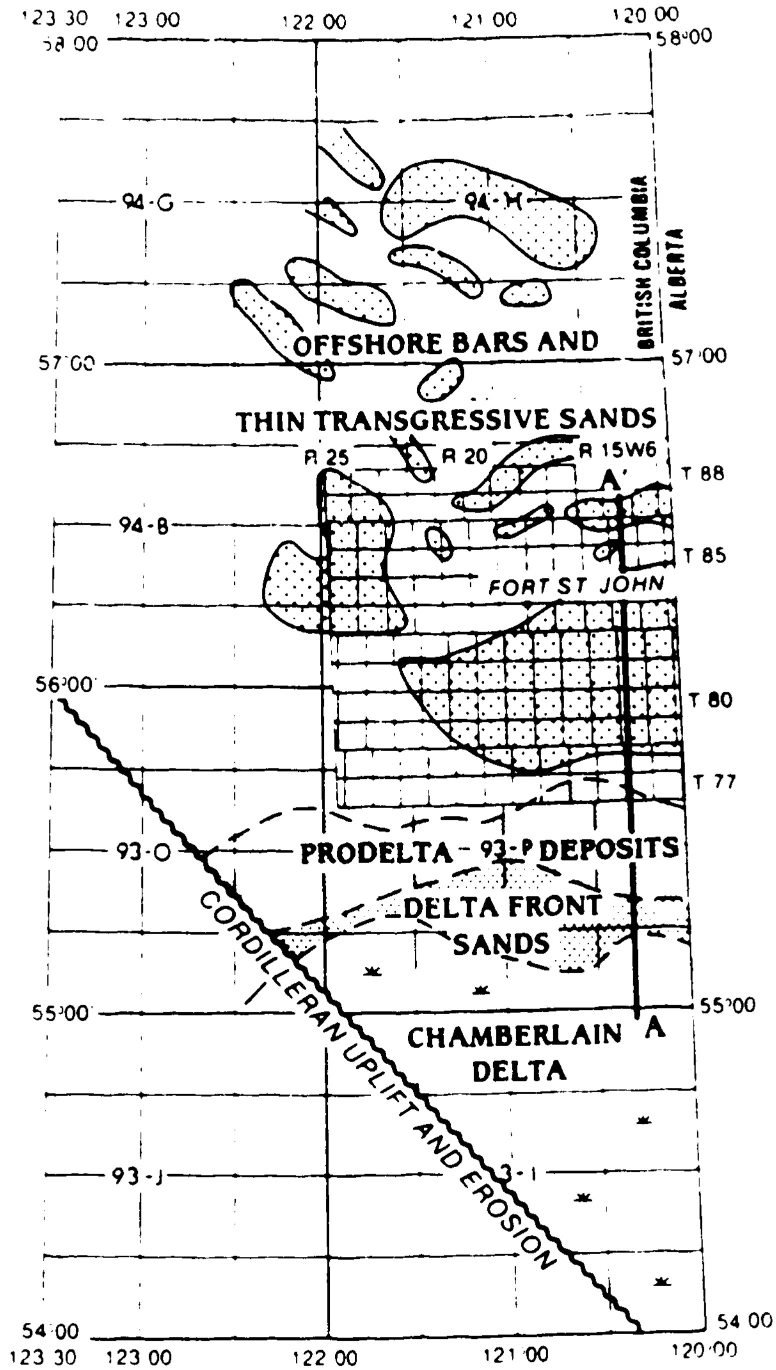


FIGURE 6.1:

PALEOGEOGRAPHY OF BLUESKY AND CHAMBERLAIN SEDIMENTS. BLUESKY SEDIMENTS CONSIST OF A COMPLEX ARRAY OF OFFSHORE BAR, THIN TRANSGRESSIVE SANDS AND CONGLOMERATES IN A SHALLOW MARINE SETTING. PRODELTA AND DELTA FRONT DEPOSITS PROGRADED FROM THE SOUTH IN A BRACKISH WATER SETTING NEAR THE END OF BLUESKY SEDIMENTATION. THE DELTAIC DEPOSITS OF THE CHAMBERLAIN FORMATION BUILT OUT OVER THE PRODELTA PLATFORM OF THE BLUESKY sedimentation. Bar sandstones probably accumulated initially in water sufficiently deep to be only minimally affected by normal wave action. As relief developed within the bar sands, increasing energy levels brought about a shift in the behavior of organisms from deposit-feeding to suspension-feeding strategies. Continued growth in relief of the bar allowed bioturbation processes to be overtaken by the physical processes of sedimentation. This coarsening-upward pattern appears to be truncated abruptly at the top by a return to fine-grained deposits containing varied ichnofauna, but dominated, for the most part, by deposit-feeding forms. This trend in depositional pattern is strongly indicative of regression and shoaling of the shelf followed by a sudden transgression and deepening of water depth or, possibly, a reduction or shift in the supply of sediment. If water depths continue to increase with transgression, accompanied by a sudden shift in sedimentation, the shoal would become a relic sandbody which would be forced into hydrodynamic equilibrium under the new energy conditions. This regressive succession overlain by deeper marine sediments, suggests that they formed as part of an overall transgressive regime.

6.1 DEPOSITIONAL SETTINGS

6.1.1 PEACE RIVER PLAINS AREA

The initial and most widespread deposit of the Bluesky Formation is the transgressive conglomerate lag that occurs immediately above the carbonaceous sediments of the Gething Formation (Figure 6.4) These lags, comprised of the poorly sorted matrix-supported pebble conglomerates (sublithofacies 1a), developed as the boreal seaway transgressed across the older deposits, reworking Gething delta deposits. Alternatively, these conglomerates may have been emplaced by fluvial processes associated with a lowstand in sea level. However, these sediments are widespread deposits rather than the localized linear features expected with fluvial deposition.

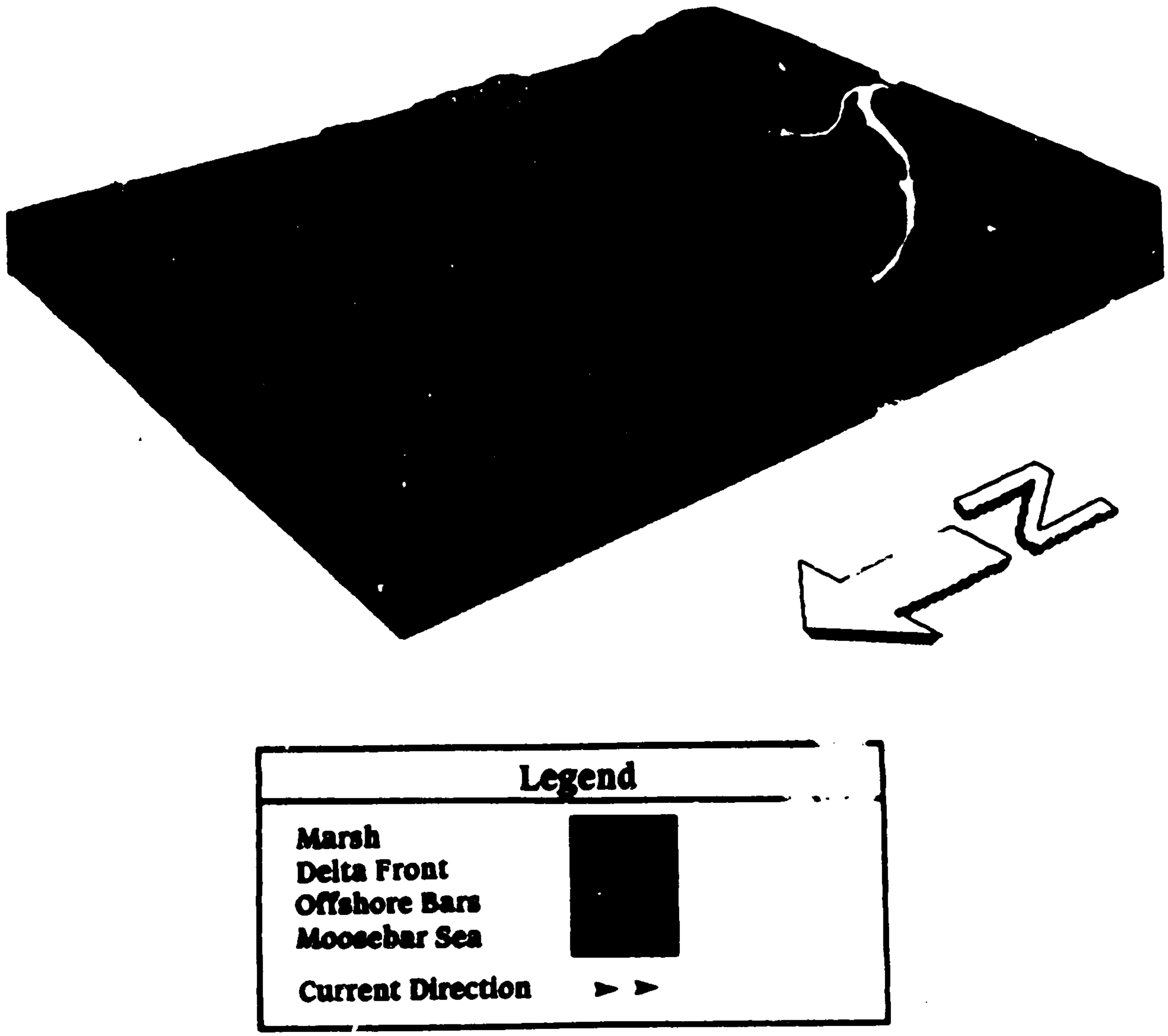


Figure 6.2 : Block diagram of inferred Bluesky paleogeography

In addition, the contact at the base of the conglomerates is frequently bioturbated and *Thalassinoides* burrows incised into underlying mudstones have been recorded. The morphology of these burrows, with sharp walls, conglomerate backfill and no linings, indicates they may be representative of a firmground suite of trace fossils (*Glossifungites* ichnofacies). This firmground condition indicates a possible depositional hiatus at the end of Gething time and erosion of a previously buried substrate. The subsequent burrowing of this substrate by benthic fauna indicates that marine or marginally marine conditions were present prior to the emplacement of the conglomerate. The conglomerate infill of the burrows indicates that the burrows remained open, due to the cohesiveness of the substrate, during the emplacement of the conglomerate. This contact, then, represents a distinct shift in sea level and possibly a sequence boundary (Haq *et al.*, 1987), separating deposits that are not genetically related.

Sedimentological studies of offshore bar complexes, such as the Hygiene Sandstone in the Denver Basin (Porter, 1976) and the Shannon Sandstone of Wyoming (Tillman and Martinsen, 1984), have provided commonly-used terminology which can be used here, with little modification, to describe the components of the shelf bar complexes. The lithofacies and their respective depositional settings are described in order of successive decrease in energy levels.

The low- to high-angle cross-stratified medium-grain sandstone (lithofacies 7) is interpreted to represent the central bar complex, the consistently highest energy zone of the bar. The occurrence of trough- and planar cross-stratification represents reworking of sands by moderately powered asymmetrical oscillatory currents. Because of the relative coarseness of the sediment and the extensive development of stratification, this part of the shelf bar would have been dominated by bedload tractional transport processes. The dearth of trace fossils is probably the result of the physically demanding wave processes inherent in this setting. No evidence for emergent

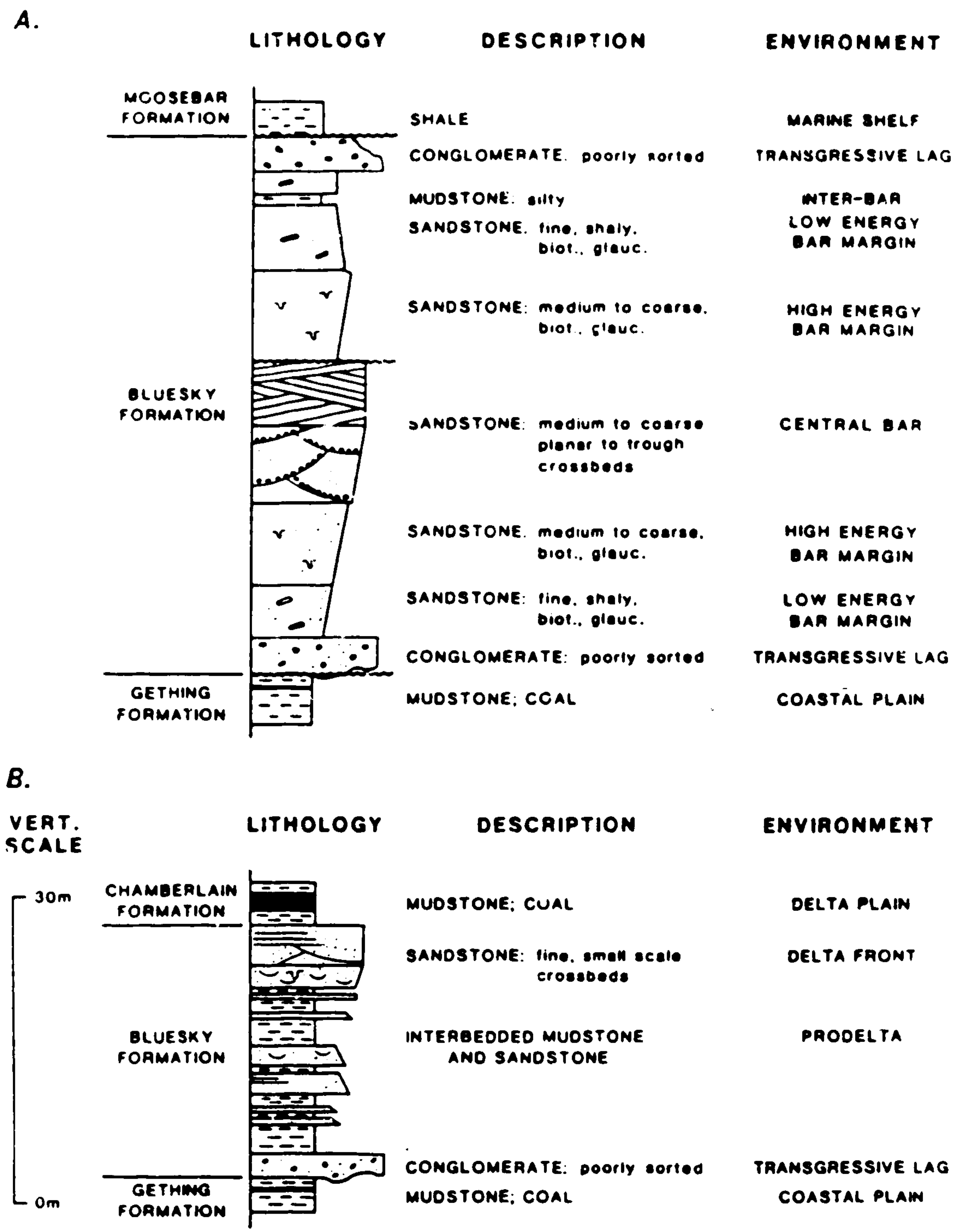


Figure 6.3: Composite vertical lithofacies successions and interpreted depositional environments for the Bluesky Formation in the A) Peace River Plains subsurface and B) Foothills area from the Sukunka to Wapiti Rivers.

or foreshore conditions have been identified. The central bar complex had aggraded or prograded over bar margin deposits.

Sublithofacies 5a and 5b, bioturbated glauconitic sandstones, are interpreted as representing proximal- and distal-bar margin environments, respectively. The proximal bar margin deposits, sublithofacies 5a, is predominantly less muddy, coarser grained and contains a suite of trace fossils representative of the *Skolithos* ichnofacies. The distal bar margin deposits are more argillaceous and support a more diverse and abundant suite of trace fossils, representative of the *Cruziana* ichnofacies. The scarcity of suspension-feeding structures in the distal bar margin deposits suggests relatively low energy levels making suspension-feeding structures inefficient. Upward in the succession into sublithofacies 5a, suspension-feeding structures become more abundant, presumably because of higher energy levels. The intensity of burrowing in both environments reflects stable conditions under which the resident infauna completely rework the substrate. Horizontally-stratified to massive sandstone deposits, which occur in laminated to burrowed sets (Figure 4.6c), are interpreted to represent deposition from storm currents.

Lithofacies 8 is considered to represent the interbar environment. The occurrence of silty mudstones and an indigenous suite of ichnofossils characteristic of the *Cruziana* ichnofacies are interpreted to represent fairly low energy conditions and would be expected to occur in areas of the shelf lateral to the shelf-bar buildups.

The succession described above is somewhat similar to that of a prograding shoreface. However, the physical and biogenic evidence suggests deposition or buildup of sandstone in the offshore. In contrast to classic models of shoreline regression, no foreshore or beach deposits were clearly defined. Indeed, the overall ichnofossil evidence point to a stable succession deposited in water depths up to 30 m.

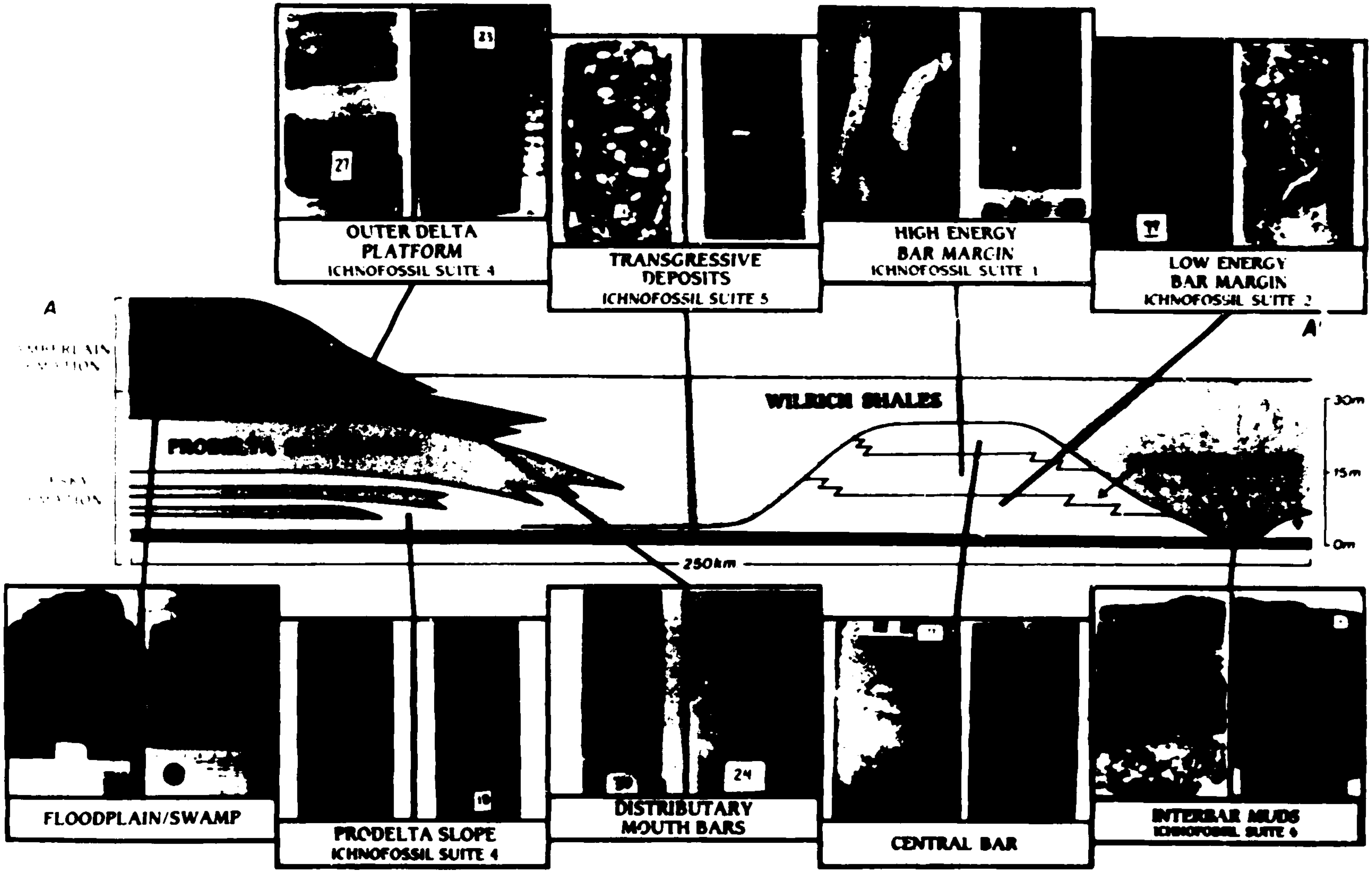


Figure 6.4: Generalized depositional model for the Bluesky and Chamberlain Formations. Core photographs illustrate the character of the deposits found in the shelf bar, delta front and delta sediments. Line of cross-section along A-A' is indicated on Figure 6.1. Refer to Chapter 5 for description of accompanying ichnofossil suites.

6.1.2 ROCKY MOUNTAIN FOOTHILLS AREA

The facies succession in this area records a prograding deltaic succession which is divisible into two main components or depositional settings. The basal section of the column consists of mud-dominated prodelta shelf deposits which are overlain by sand-dominated delta front deposits. This succession is overlain by the terrestrial delta plain deposits of the Chamberlain Formation.

The prodelta environment is transitional to the shelf, entirely subaqueous and comprises the lower reaches of the deltaic setting (Coleman and Gagliano, 1965). Lithofacies 2, the interbedded mudstones and fine-grain sandstones, is considered to represent the prodelta environment. These deposits are characterized by thinly bedded bioturbated mudstone and cross-laminated sandstone beds. The sedimentary structures within the lithofacies reflect a rhythmic or cyclical style of sedimentation, such as seasonal flood or storm events. These cyclical events are considered to deposit sediment rapidly, predominantly from suspension, tractional and gravity processes. Soft sediment deformation in the form of oversteepened bedding, microfaulting and convolute bedding indicates substrate instability. The coarsening- and thickening-upward trend of the sandstone beds indicates a shallowing of water depth as the shoreline of the Chamberlain delta advanced. This upward coarsening succession also records the gradual introduction of a brackish water suite of trace fossils (Suite 4).

The basal portion of the prodelta setting is characterized by a low-diversity *Cruziana* ichnofacies assemblage *(Planolites* and *Helminthopsis)* which indicates significant water turbidity and fluctuating suspended load sedimentation. The upward change to a higher diversity brackish water ichnofacies indicates a shift in sedimentation to higher energies and fresher water conditions as river input becomes more predominant with the advancing shoreline.

The change of depositional setting from the mud-dominated prodelta to the sand-dominated delta front is commonly transitional. The delta front deposit is the seaward subaqueous portion of the delta where relatively coarse-grain sediments are influenced by both wave- and distributary-processes (Coleman and Gagliano, 1965). Lithofacies 3, the small-scale cross-stratified medium-grain sandstone, is interpreted as a delta front sheet sand-mouthbar complex that built seaward over the interbedded mudstones and sandstones of the prodelta environment. Thick accumulations of laminated sandstones are interpreted as the deposits of distributary mouth bars, whereas the thinner sandstone beds probably represent the laterally spreading sheet sands which develop from sediment dispersal by longshore currents. Trough cross-stratification in the sandstones and the presence of brackish trace fossils indicate the proximity of the strand line. The time-equivalent strandplain lay to the south of the delta front deposits, indicating that the delta plain prograded northwards.

As the deltaic succession continued to prograde, nonmarine to marginally marine Chamberlain Formation sediments built over the delta front deposits. These sediments consist of coastal swamp or marsh, shoreface/beach, distributary channel, overbank floodplain and interdistributary bay deposits. Along the leading edge of the delta, where longshore currents were operative, sediments furnished at river mouths may have been redistributed, forming beach sandstones in places. Thick seams of coal (>5 m) are found interbedded between the floodplain and distributary channel deposits. Following deposition of the Chamberlain Formation, the region was subjected to a major boreal transgression of the Moosebar Sea depositing deep water shales over the delta plain deposits. This transgression, unlike that which initiated Bluesky deposition, was sufficient to flood all areas of northeastern British Columbia and west-central Alberta (Taylor and Walker, 1984). This transgression formed a quiet and deep water marine shelf setting.

6.2 SAND DISPERSAL MECHANISM

Although the above information summarizes the general setting for deposition of the deltaic and offshore sand sediments, it does not account for the origin of the sand on a distant shelf. Shelf sand sedimentation has received considerable attention in recent years. Offshore bar models have been proposed for many of the Cretaceous marine sandstones, including the Duffy Mountain Sandstone in northwestern Colorado (Boyles and Scott, 1982), the Hygiene Sandstone of the Denver Basin (Porter, 1976), the Sussex Sandstone of the Powder River Basin (Berg, 1975) and the Shannon Sandstone of the Hartzog Draw Field in Wyoming (Spearing, 1976; Tillman and Martinsen, 1984, 1987). Most of these offshore bar models are morphologically similar, consisting of cross-laminated and bioturbated sandstones built-up from a muddy seafloor. The only fundamental difference is the type of shelf process, such as oceanic-, tidal- or storm/wave-dominated processes, which influences the type of deposit. The Shannon Sandstone, in the Powder River Basin, has been re-interpreted by Gaynor and Swift (1988) as a'shelf ridge sand' deposit. The distribution of lithofacies in this model is suggested to be specifically related to the velocity distribution of the regional flow field across the ridge (Gaynor and Swift, 1988).

Offshore bar models have also been described for the Cretaceous section in Alberta, including: the Viking Formation in central Alberta (Reinson *et al.*, 1983; Farshori and McKay, 1986), the Wabiskaw 'C' sand in northeastern Alberta (Ranger *et al.*, 1988) and the Cardium Formation in the Carrot Creek Field (Swagor *et al.*, 1976). Most of the above-mentioned sandstone bodies, except for the Wabiskaw, are typically encased in mudstones and occur far from any recognizable paleoshoreline. The processes which are responsible for the transport of coarse clastic material into a marine shelf setting has been the focus of considerable recent debate. Indeed, the processes responsible for the deposition of sand contained in shelf deposits of the Western Interior Seaway may prove to be related to factors other than transport mechanisms.

The movement of sediment from the nearshore to the shelf has been attributed to several dynamic processes which break through the nearshore depositional systems. Sediment transport onto a shelf has been suggested to occur by river mouth bypassing, shoreface bypassing (Brenner, 1980) or storm/turbidity flows (Swift and Rice, 1984). River mouth bypassing occurs during a flood stage where the flow exiting a delta is of sufficient strength to carry sediment past mouth bars into the offshore (Brenner, 1980). Shoreface bypassing occurs when the sea transgresses a shoreline and reworks the sand into shelf sand bodies. Storm flows occur when storm-surge return flows carry sediment from the beach area into the shelf (Aigner and Reineck, 1982; Swift and Rice, 1984).

Recent sedimentological studies of modern shelves have suggested that sand bodies found on the shelf are relic sediments deposited originally at low stands of sea level and later stranded by transgression (Swift *et al.*, 1973). Studies of the northeast Atlantic shelf, however, also suggest that not only are relic sediments being reworked, but that new sediment is being introduced to the shelf (Swift *et al.*, 1979).

Sedimentologic studies of the Cardium Formation (Bergman and Walker, 1986, 1988) and the Viking Foramtion (Beaumont, 1984; Downing and Walker, 1988) have suggested that the coarse-grain material was introduced as shoreline deposits during a low stand of sea level. During the lowstand, sand and granules were supplied to the shoreline by rivers and distributed along the shoreface. The subsequent rise of sea level encased the sandstone in mudstone (Downing and Walker, 1988). For the Cardium Formation, a lowering of sea level apparently created a subaerial erosion surface that appears to constitute a sequence boundary. In the Bluesky deposits, although the contact of the Bluesky and Gething Formations represents a disconformity, there is no evidence of subaerial exposure, foreshore or beach conditions. Instead, it seems reasonable that the disconformable surface is most readily explained as a marine erosion surface formed by wave and current action when transgression of the Gething delta plain reduced sediment supply to the shelf. Unlike the Cardium erosion surface, which records a relative drop in sea level, the disconformity at the base of the Bluesky Formation records a rise in sea level, marking a change from delta plain deposition to marine deposition

It is difficult to precisely determine what exactly is the mode of sand emplacement onto the shelf in Bluesky time. In the southern half of the study area, thick delta front sandstones have been observed, and erosion of these may have contributed to the supply of sediment for bar sedimentation in the shelf areas. Similar settings have been intrepreted for the Muddy Sandstone of the Bell Creek Field in Montana (McGregor and Biggs, 1968) and the Woodbine Sandstone of the Kurten Field in Texas (Turner and Conger, 1984). Both of these studies describe the occurrence of progradational deltaic sediments coincidental with shelf sand bar sedimentation under transgressive conditions.

In summary, it appears that there are three alternate mechanisms for the origin of sediment on the shelf in Bluesky time. Sand may have been derived from the delta system in the south or from reworking and concentrating of underlying sediments into discrete sand bodies or, lastly, from deposition originally as shoreline sandstones as part of a lowstand wedge. The evidence gathered to date does not provide definitive proof for lowstand deposition; however, this does not discount its validity with further work. If indeed the Bluesky deposits were originally formed as part of a lowstand wedge, then the subsequent transgression may have reworked the sand deposits removing any trace of subaerial exposure.

6.3 CONCLUSIONS

1. The Bluesky Formation can be traced into the Peace River Foothills where it is overlain by the deltaic 'Chamberlain Formation'.

2. The Bluesky is comprised of coarsening-upward offshore bar complexes in the north half of the study area and prodelta/delta front deposits in the south half.

3. A) Offshore bar complexes are divided into:

    1) Central bar deposits consisting of cross-bedded sands;

    2) Bar margin deposits consisting of bioturbated shaly sands;

    3) Interbar deposits consisting of silty, bioturbated mudstones;

   B) Deltaic succesion can be divided into:

    1) Prodelta deposits consisting of interbedded sands and muds;

    2) Delta front deposits consisting of small-scale cross-bedded sands;

4. The base of the Bluesky Formation is marked by a transgressive lag deposit which contains a suite of firmground trace fossils (*Glossifungites* ichnofacies) at its basal erosive contact. This surface possibly represents a sequence boundary.

5. Trace fossil suites in the Bluesky cores also support the interpretation of two distinct environments for the Bluesky. Offshore bar complexes contain trace fossils characteristic of the *Skolithos* and *Cruziana* ichnofacies, whereas the prodelta/delta front deposits contain traces characteristic of a Brackish Water ichnofacies.

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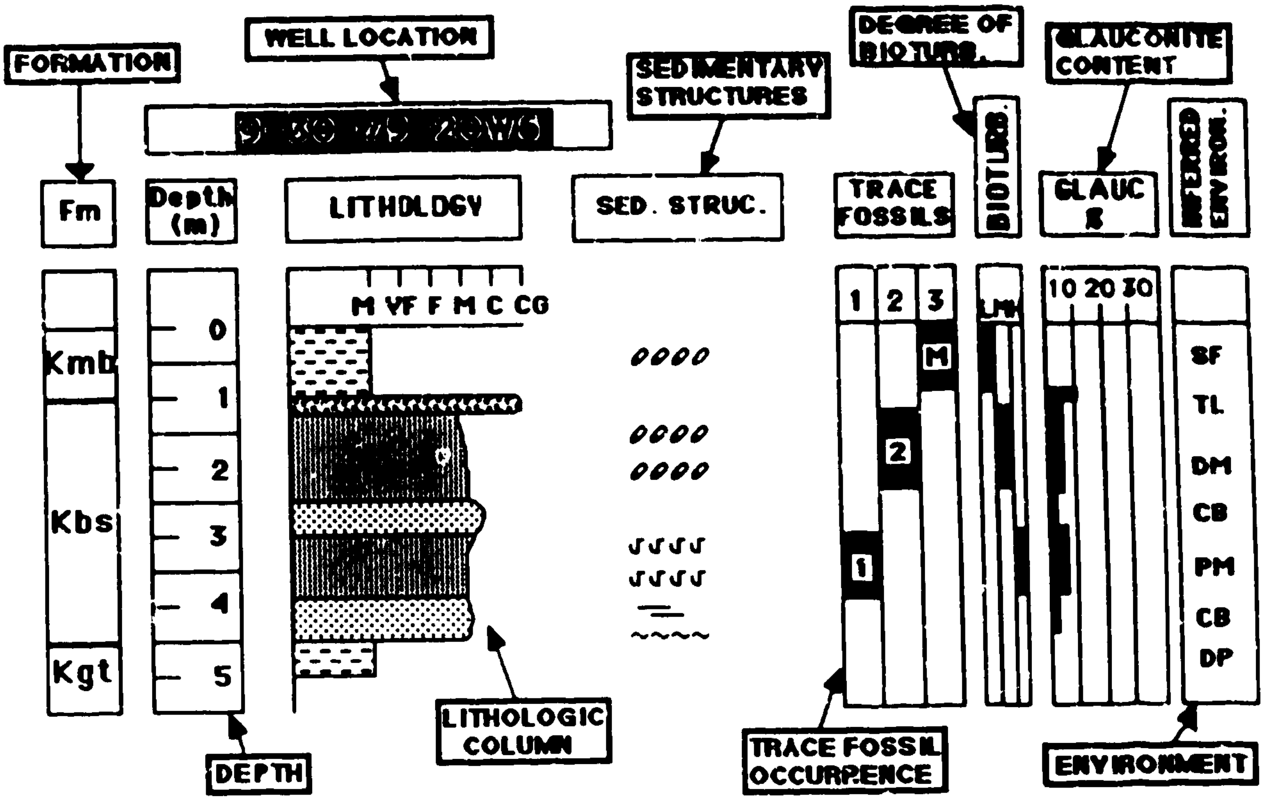
APPENDIX ONE:

CORE AND OUTCROP DESCRIPTIONS



FORMATION

Kmb – Moosebar Fm
Kch – Chamberlain Fm
Kbs – Bluesky Fm
Kgt – Gething Fm

TRACE FOSSIL OCCURRENCE

Col 1 – Suites 1 and 5
Col 2 – Suites 2 and 3
Col 3 – Suites 4 and 6 and Moosebar Suite (M)

DEGREE OF BIOTURB.

L – Low
M – Moderate
H – High

LITHOLOGY CODE

- Mudstone
- Ironstone
- Bentonite
- Coal
- Siltstone
- Interbedded Sst/Mdst
- Conglomerate w Mud-supported matrix
- Conglomerate w Sand-supported matrix
- Bioturbated Shaly Sandstone
- Non-bioturbated, Clean Sandstone

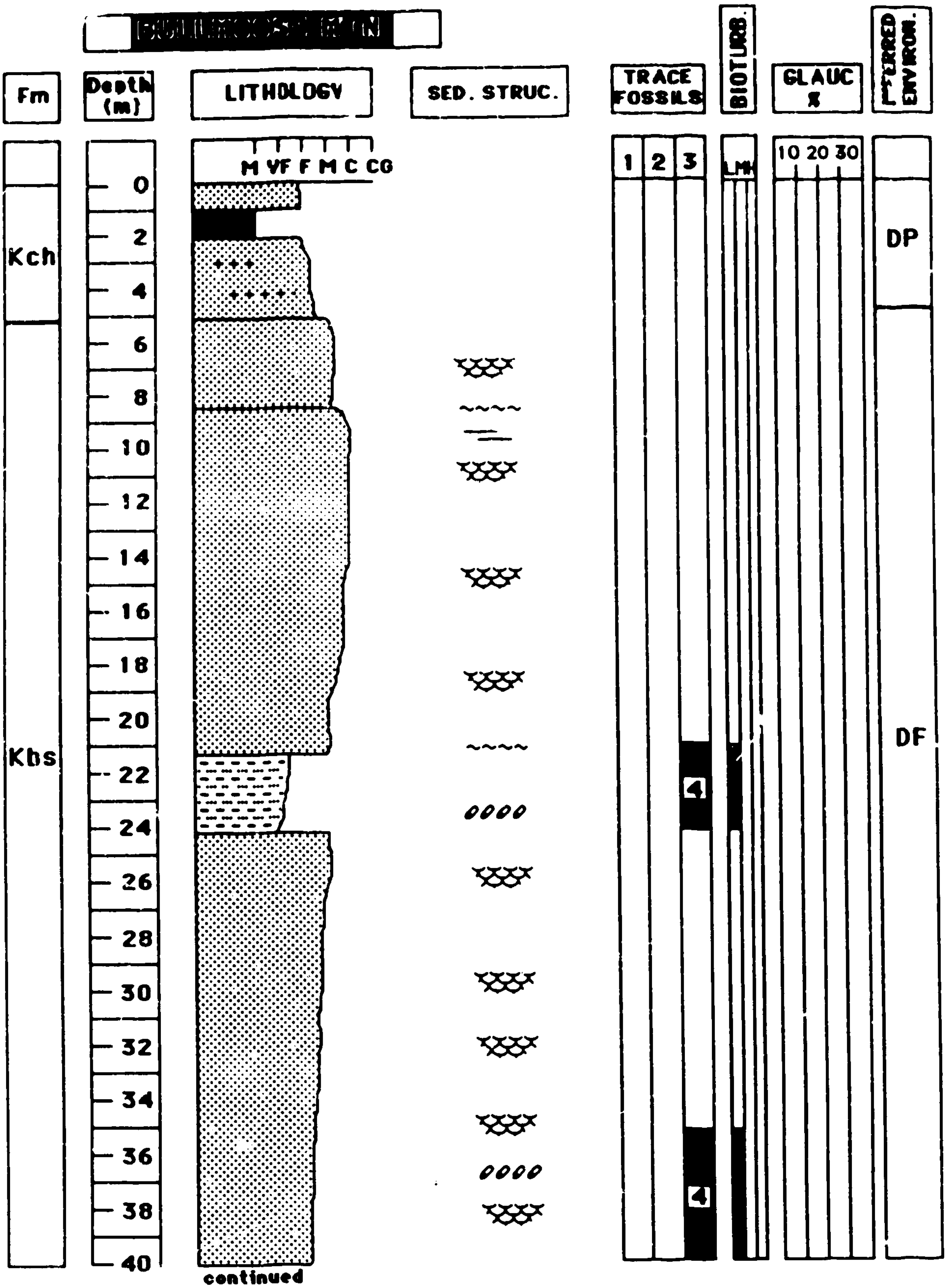
ENVIRONMENT

TL – Transgressive Lag
CB – Central Bar
PM – Proximal Bar Margin
DM – Distal Bar Margin
IB – Interbar
PD – Predelta
DF – Delta Front
DP – Delta Plain
SF – Marine Shelf

SEDIMENTARY STRUCTURES

- Dwelling Structures (Vertical Burrows)
- Feeding/Grazing Struc. (Horizontal Burrows)
- Small-scale Stratification
- Trough Cross-Stratification
- Eroded/Irregular Contact
- Burrowed Contact
- Low Angle/Planar Cross-Stratification

Legend for core description data sheets in Appendix 1.



Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
0
2
4
6
8
10
12
14
16
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32
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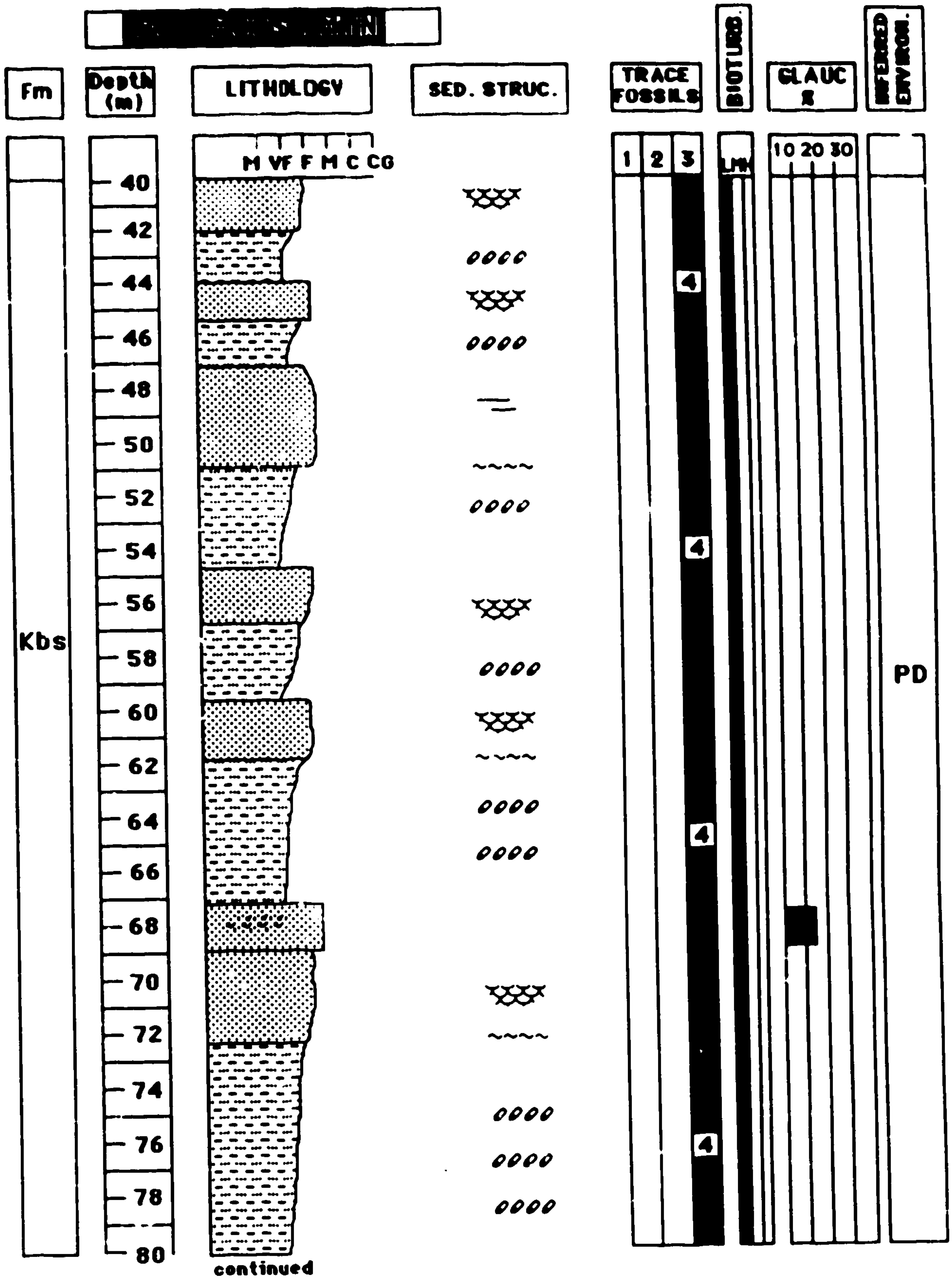
36
38
40
Kch
Kbs
DP
DF
4
4
continued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Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
40
42
44
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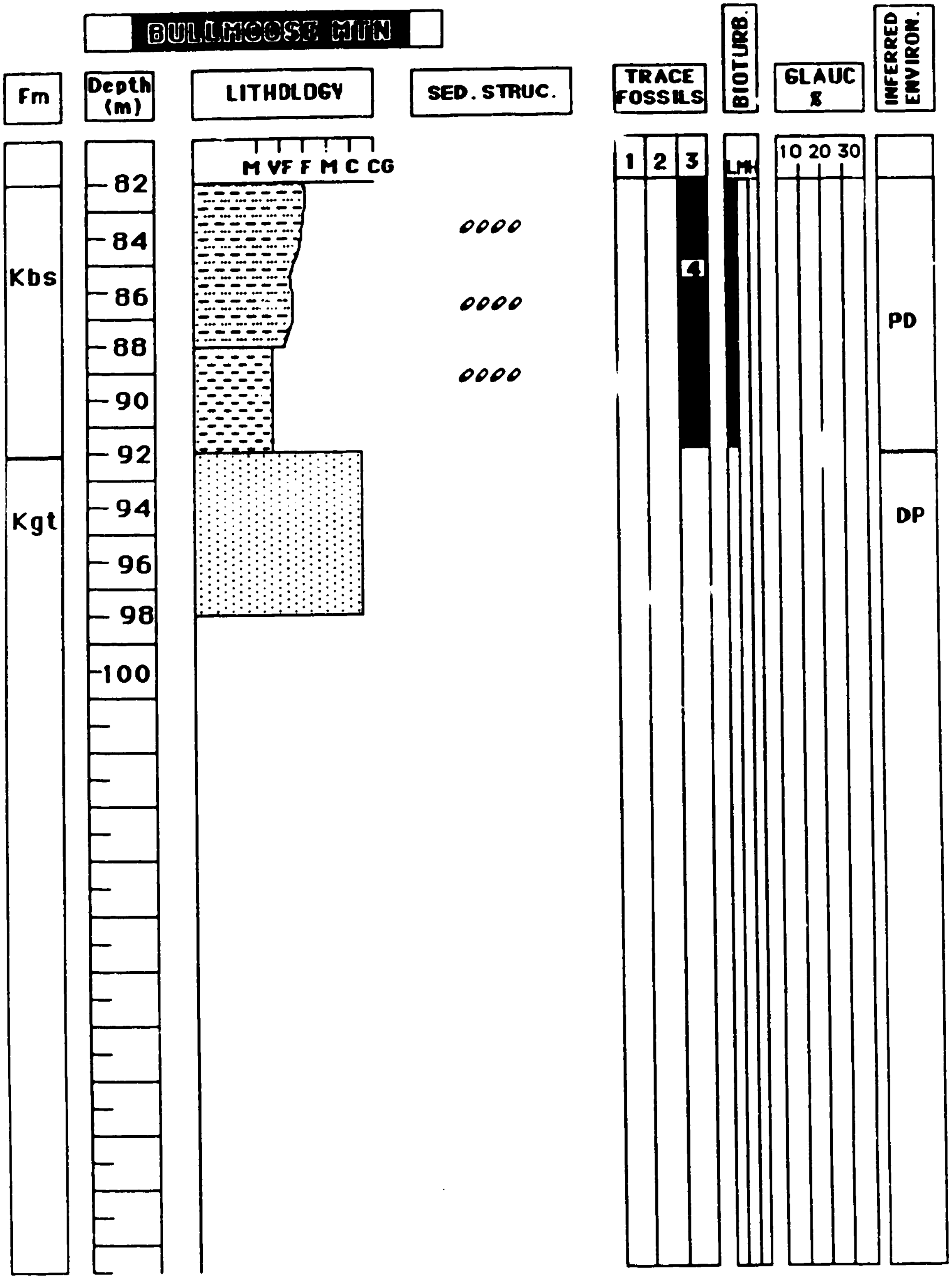
PD
continued


BULLMOOSE MTN
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
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Kgt
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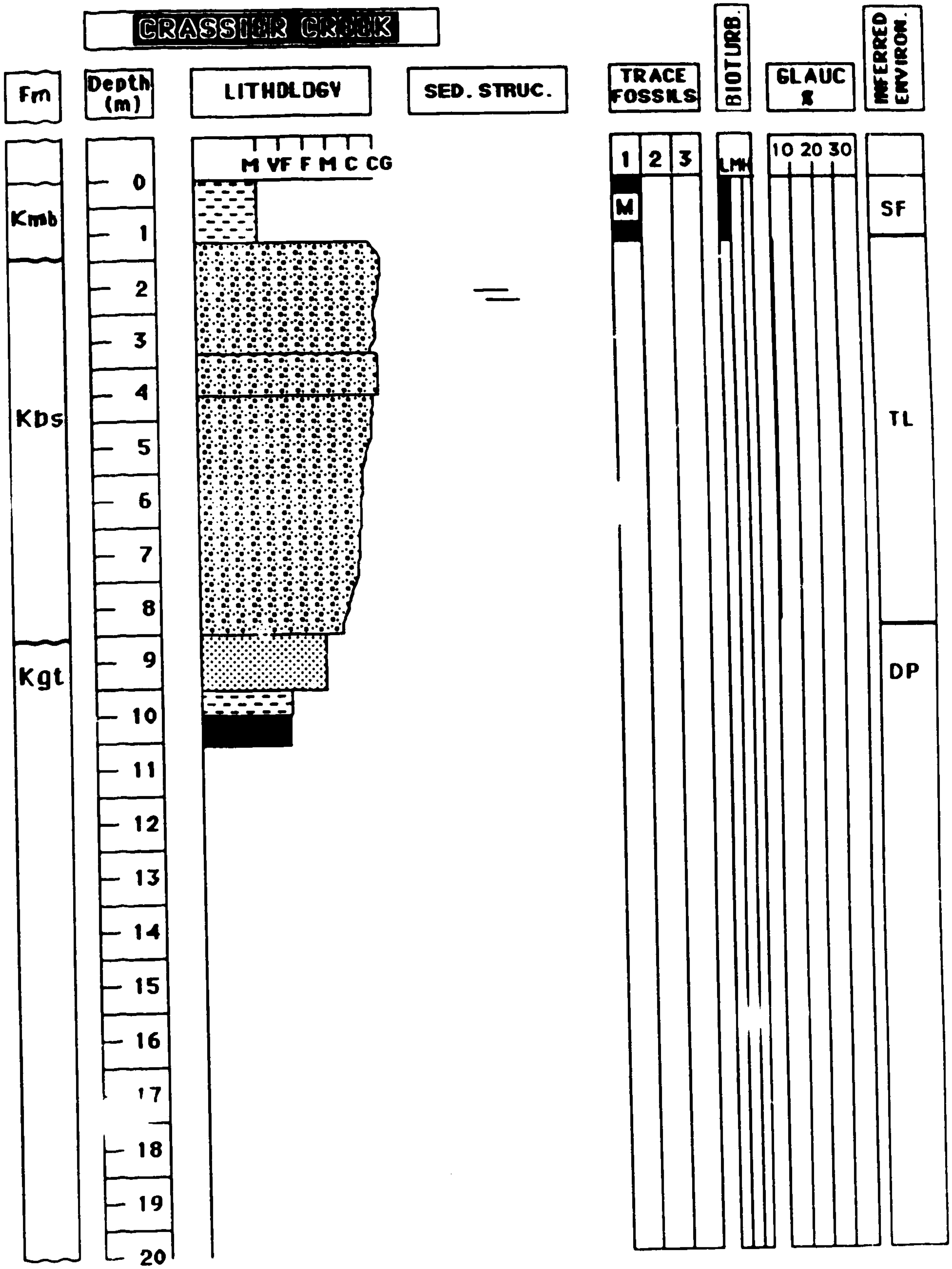
PD
DP


CRASSIER CREEK
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SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
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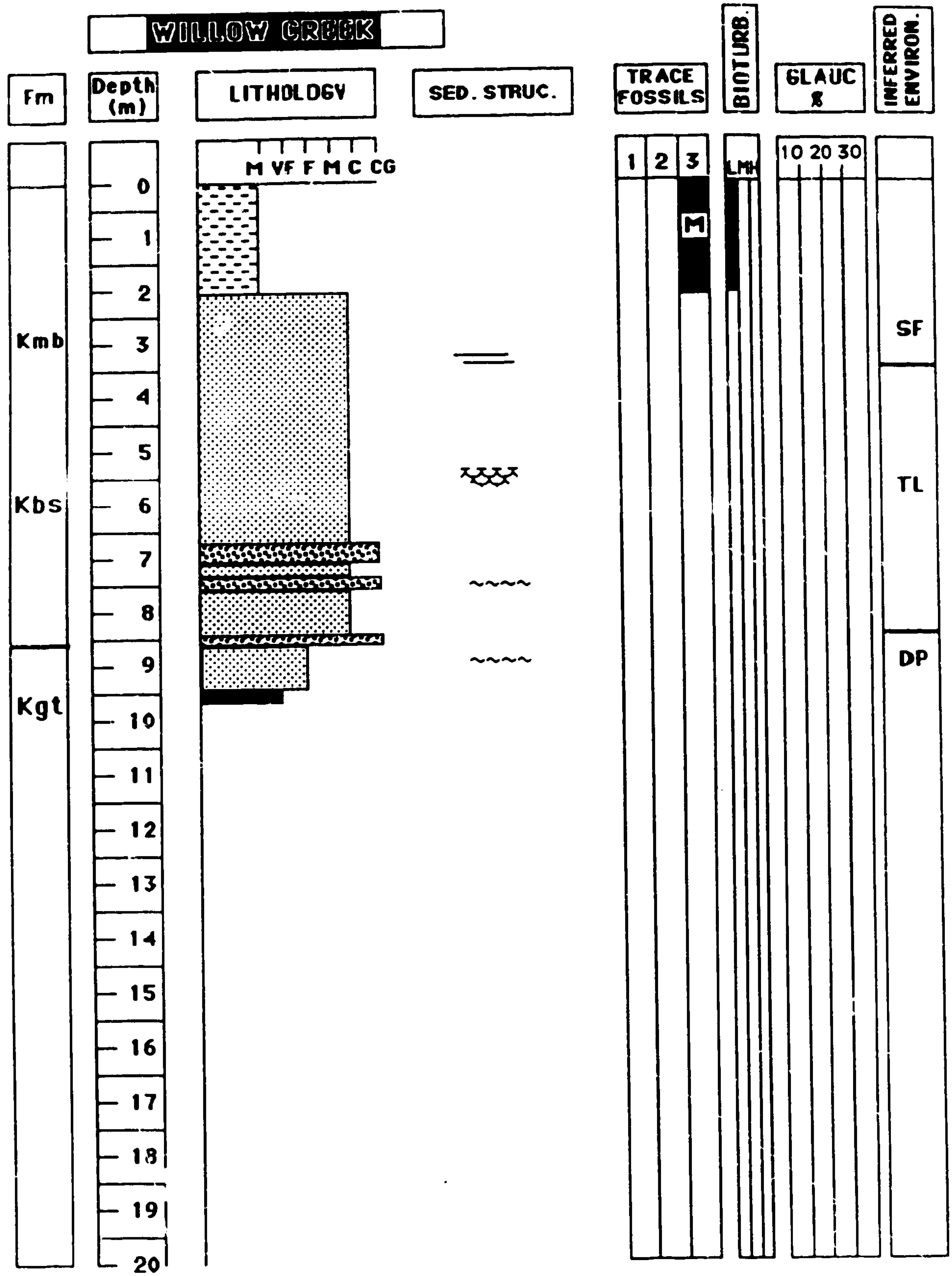
Kmb
Kbs
Kgt
M
SF
TL
DP



WILLOW CREEK
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Depth (m)
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SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
M
L M H
10 20 30
Kmb
Kbs
Kgt
0
1
2
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SF
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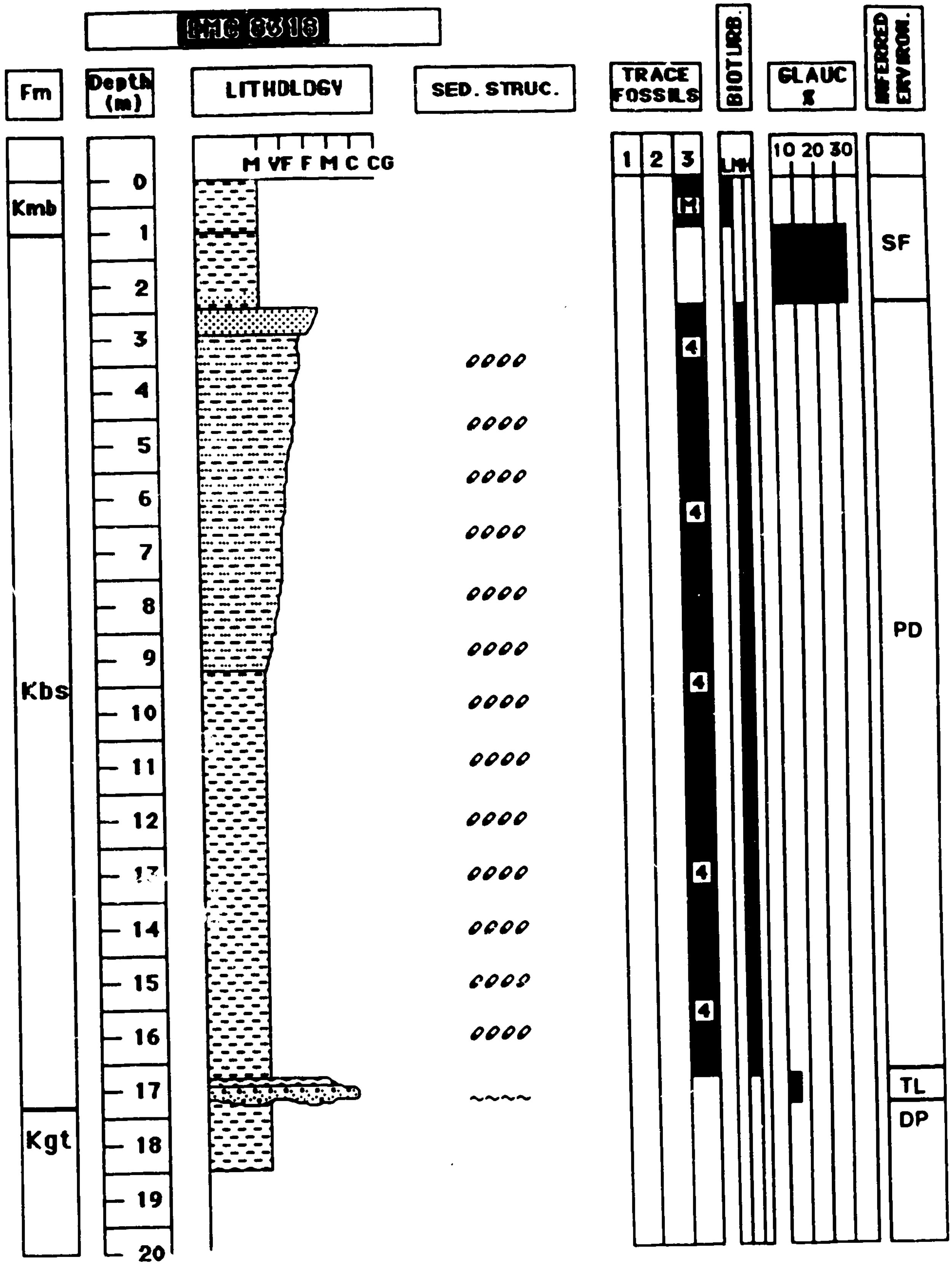

BMC 6318
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Depth (m)
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SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
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1 2 3
LMH
10 20 30
Kmb
Kbs
Kgt
0
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2
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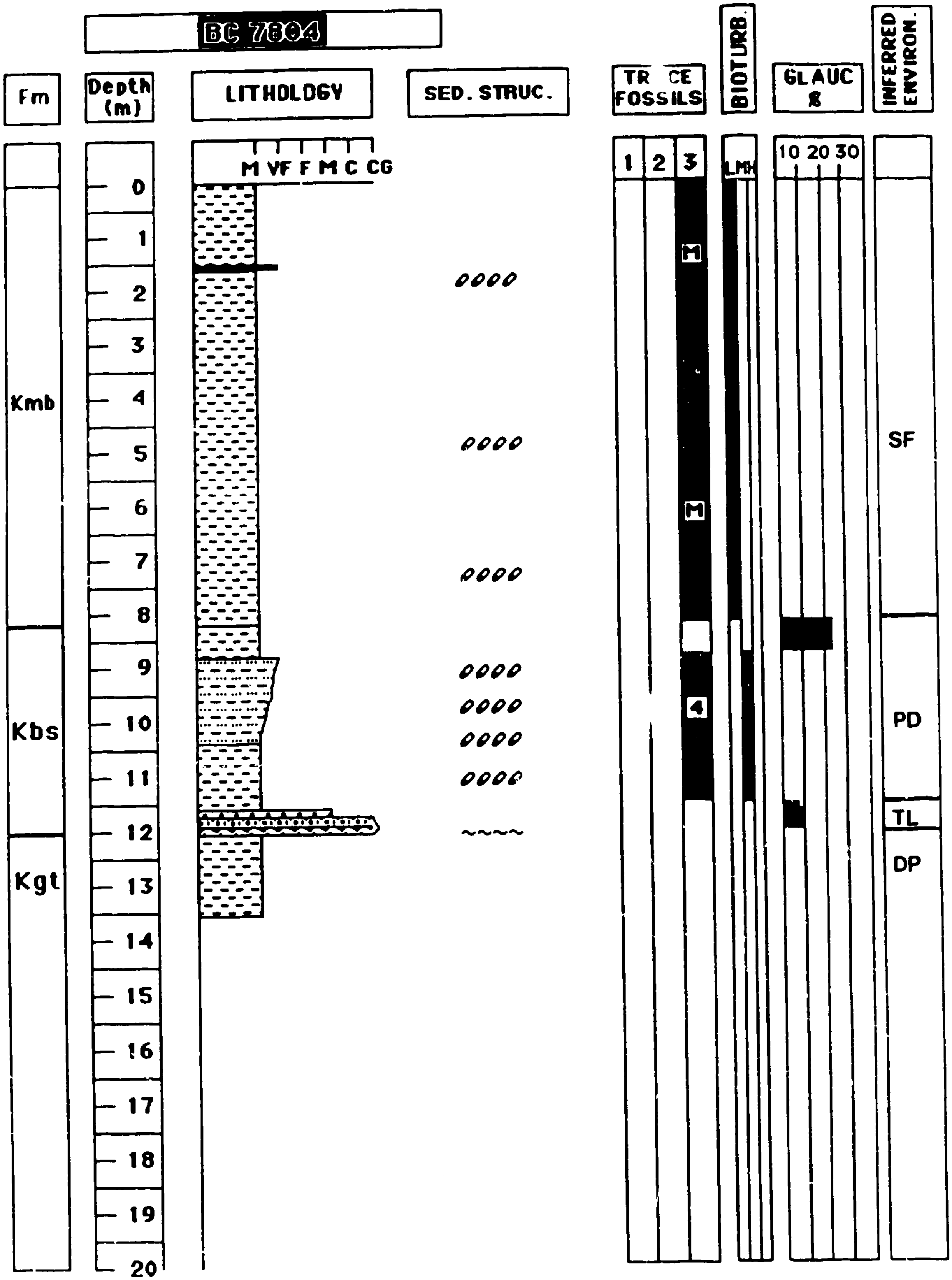
SF
PD
TL
DP



BC 7804
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SED. STRUC.
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GLAUC %
INFERRED ENVIRON.
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1 2 3
L M H
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Kmb
Kbs
Kgt
M
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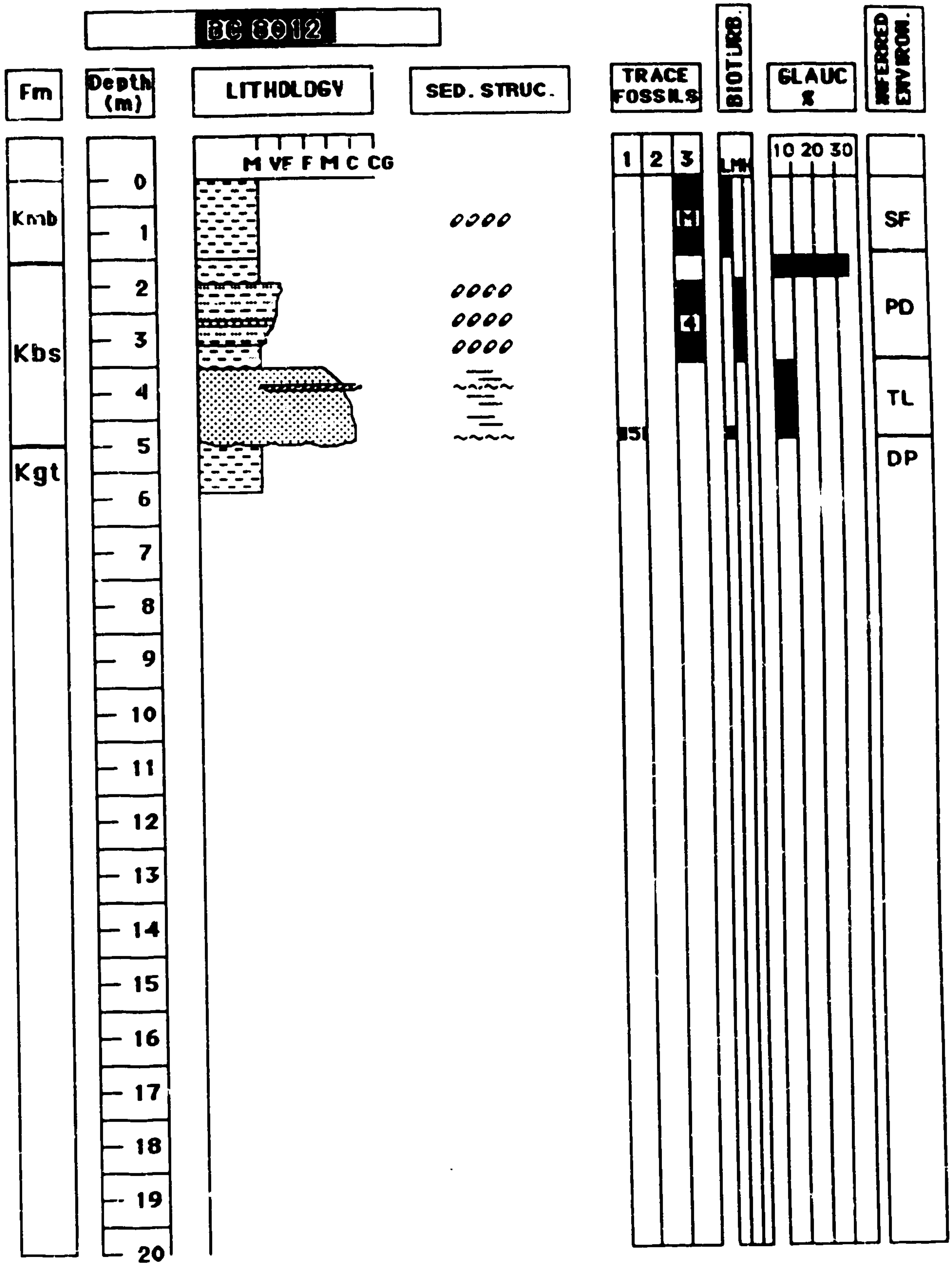
4
SF
PD
TL
DP



BC 8012
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TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
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1 2 3
L M H
10 20 30
Kmb
Kbs
Kgt
0
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2
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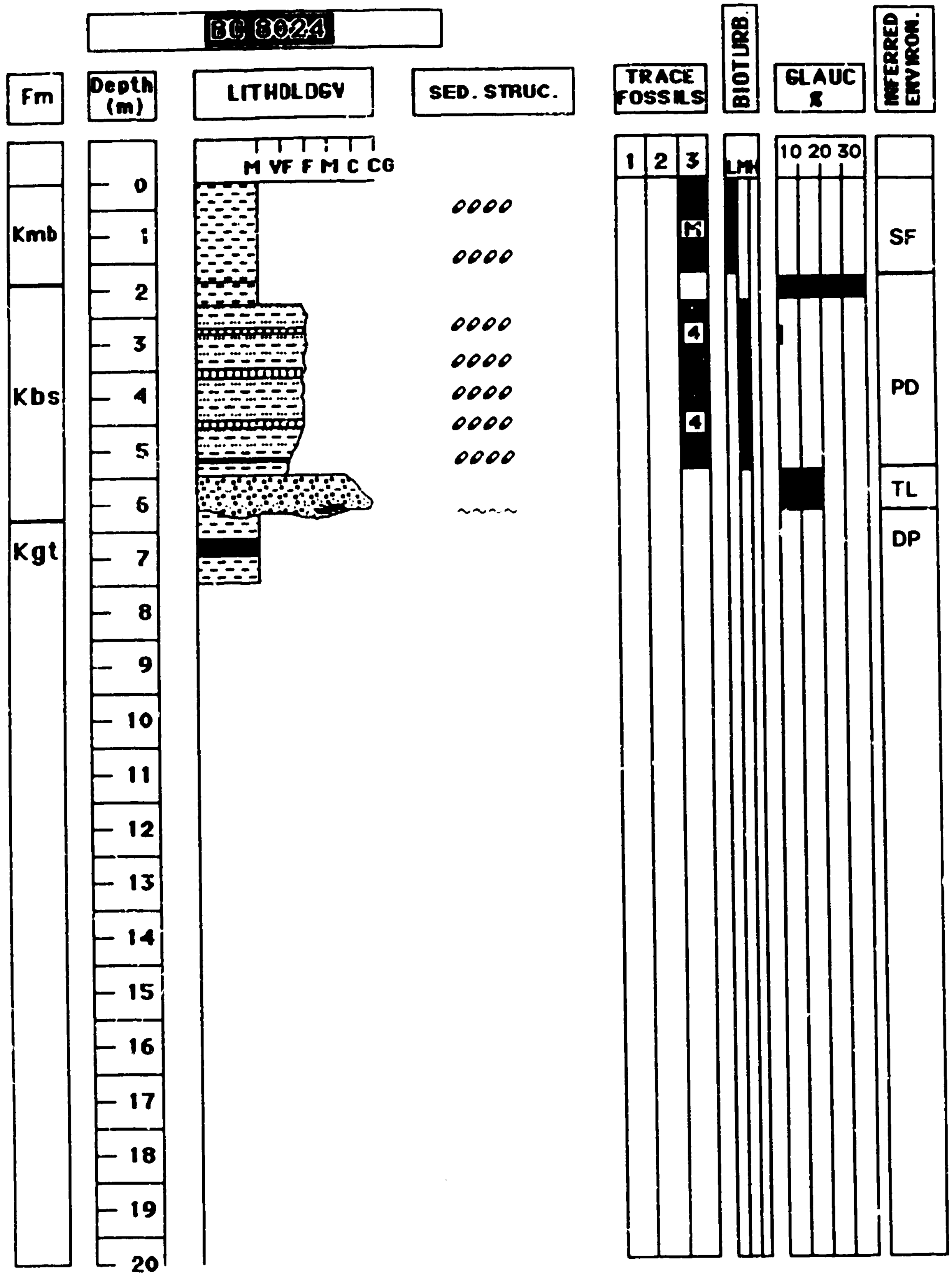

BC 8024
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GLAUC %
INFERRED ENVIRON.
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1 2 3
L M H
10 20 30
Kmb
Kbs
Kgt
0
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2
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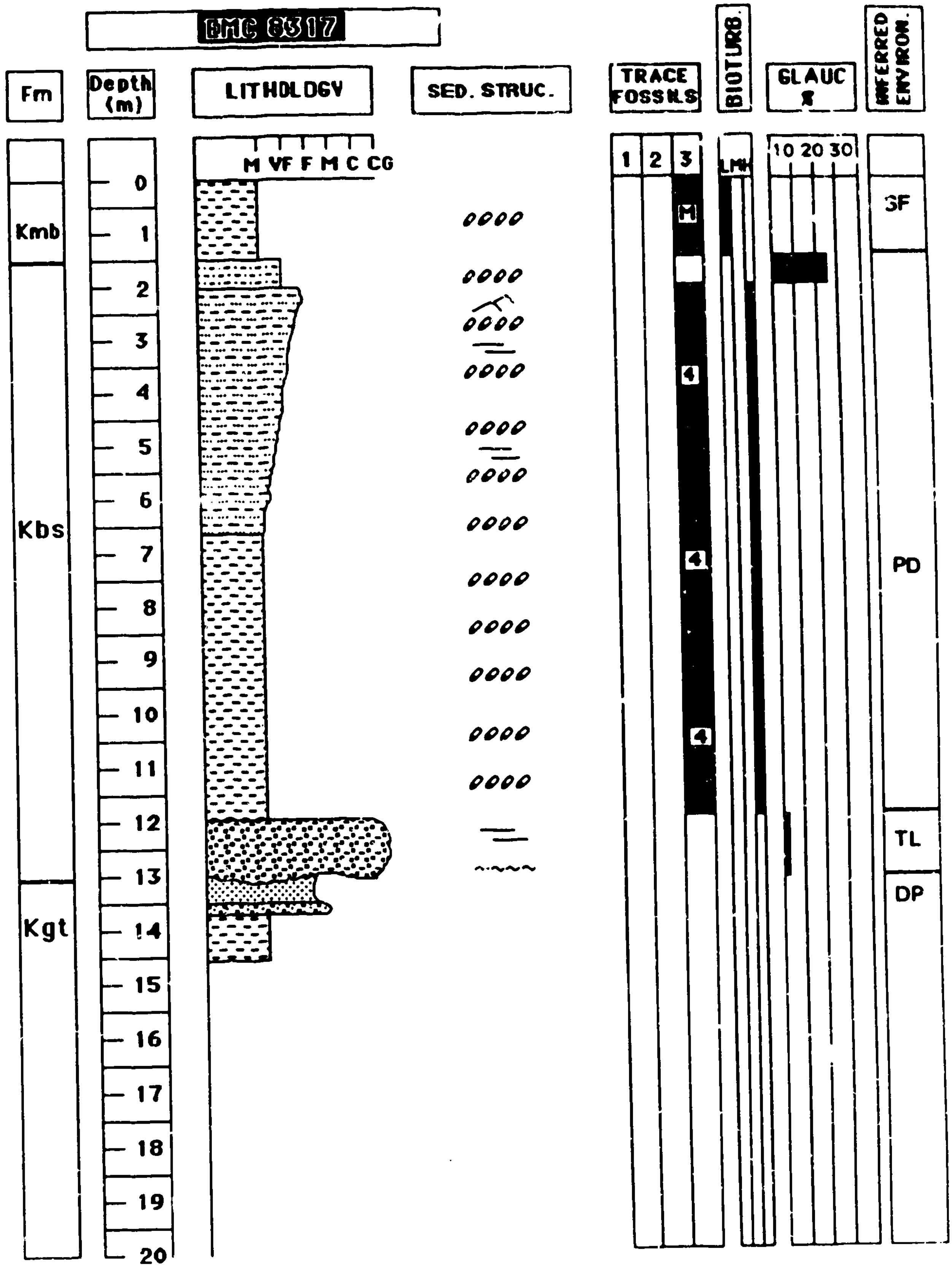
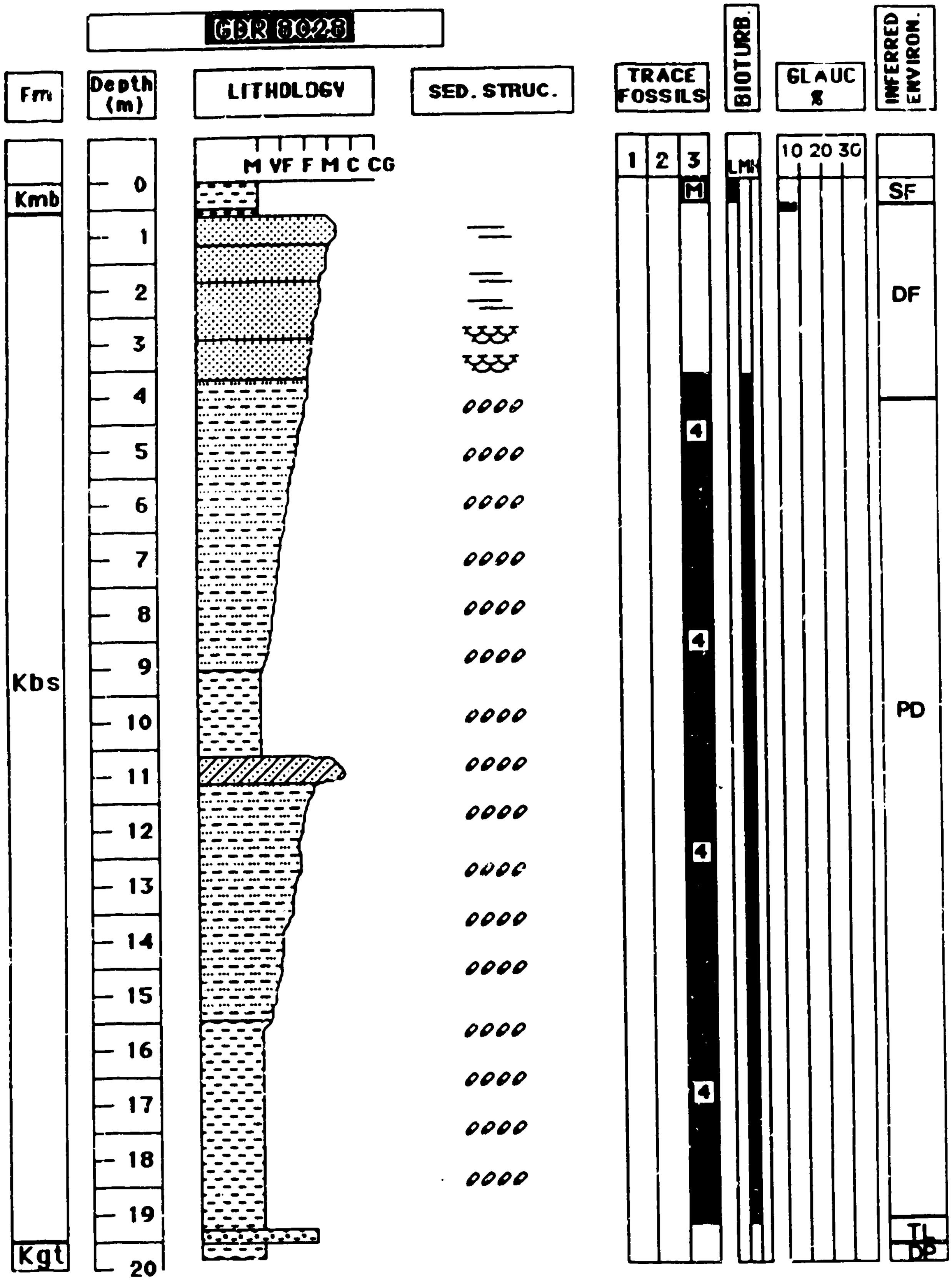
SF
PD
TL
DP



BMC 8317
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
Kmb
Kbs
Kgt
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
M
4
4
4
SF
PD
TL
DP



GDR 8028
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
Kmb
Kbs
Kgt
0
1
2
3
4
5
6
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8
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M
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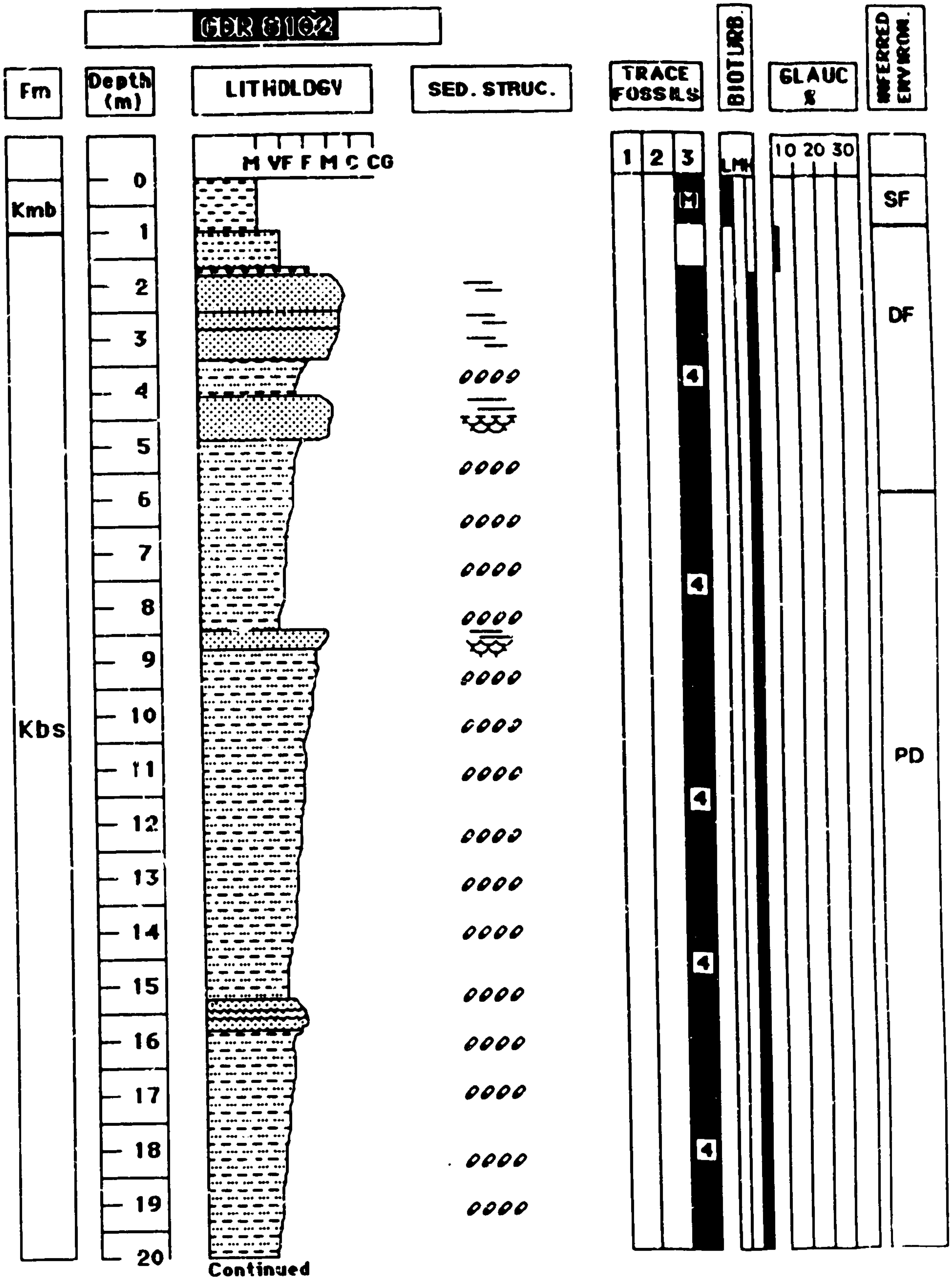
4
4
4
SF
DF
PD
TL
DP



GDR 8102
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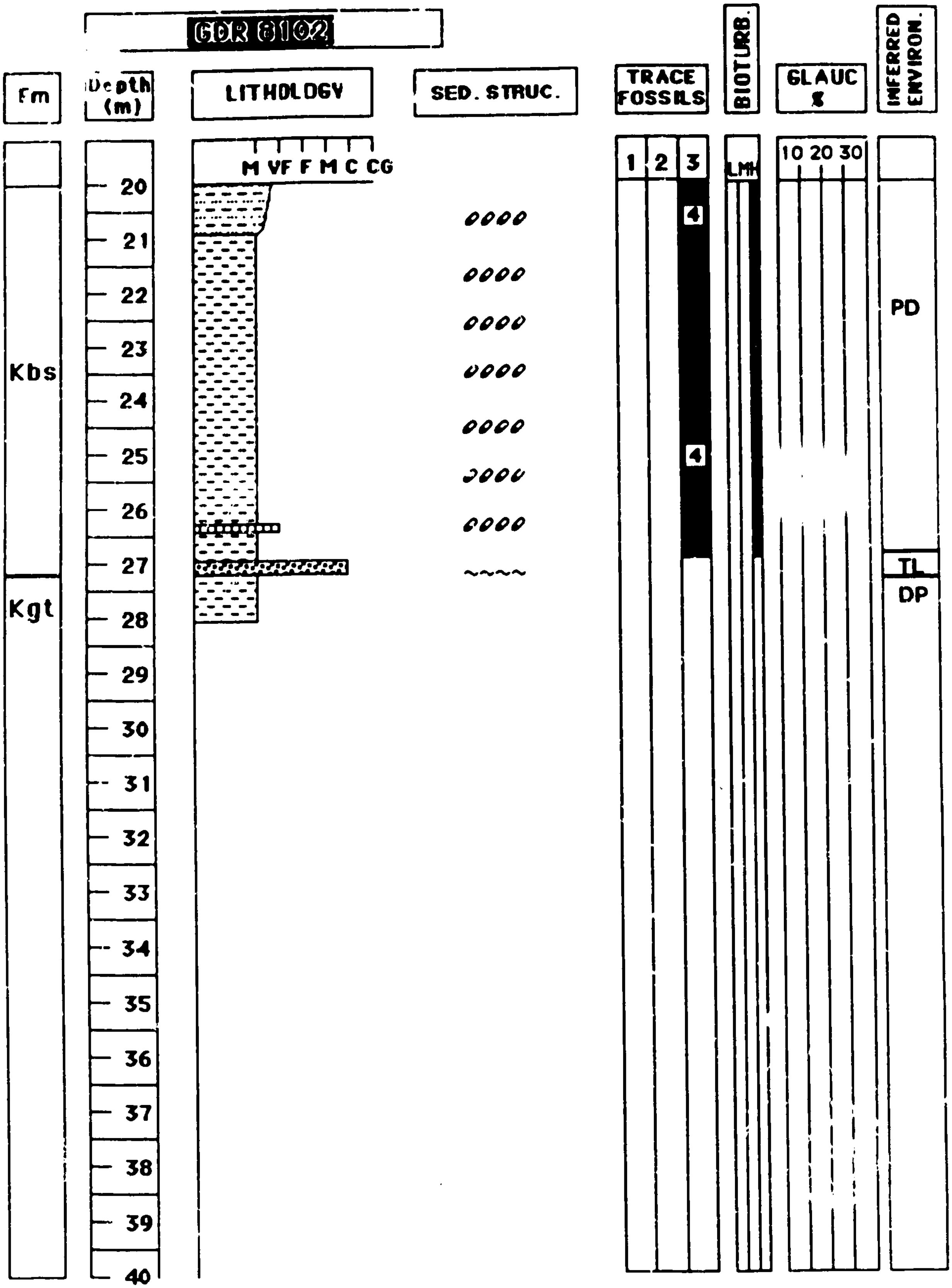
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
Kmb
Kbs
SF
DF
PD
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
Continued



GDR 8102
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
20
21
22
23
24
25
26
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Kbs
Kgt
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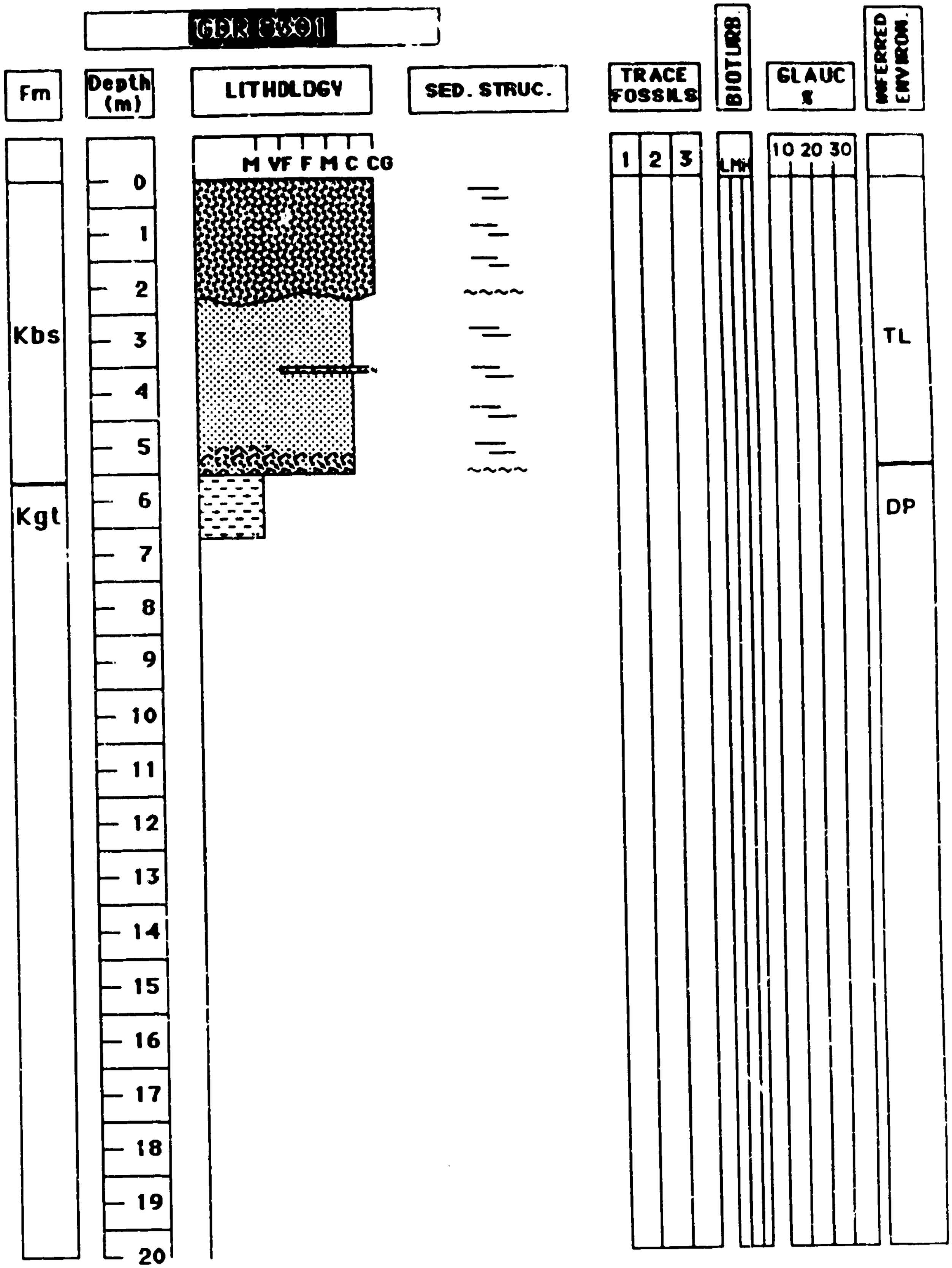
4
PD
TL
DP



Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
Kbs
Kgt
TL
DP
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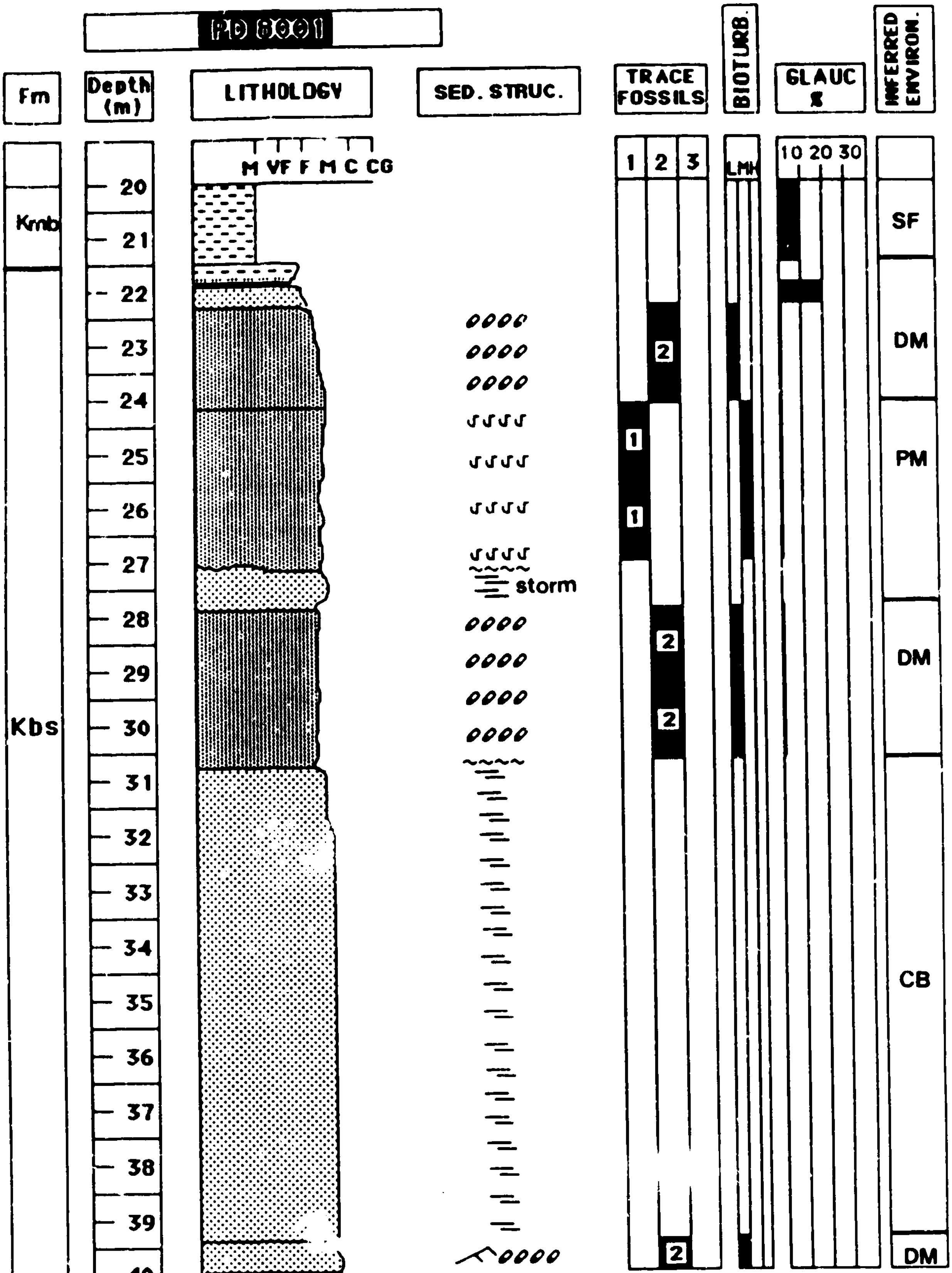
PD 8001
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
Kmb
Kbs
20
21
22
23
24
25
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27
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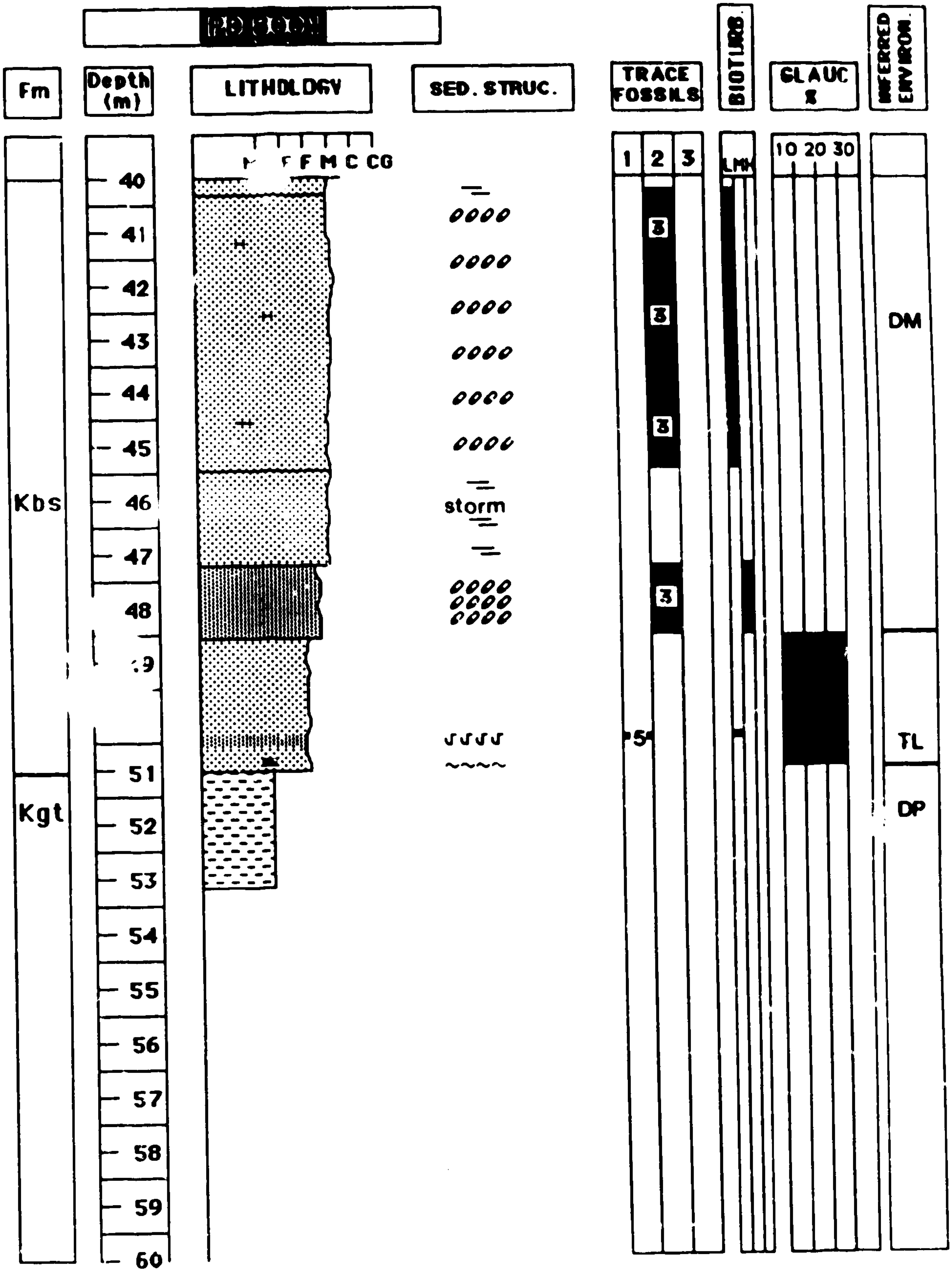
storm
SF
DM
PM
DM
CB
DM

Continued



Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
F M C CG
1 2 3
L M H
10 20 30
40
41
42
43
44
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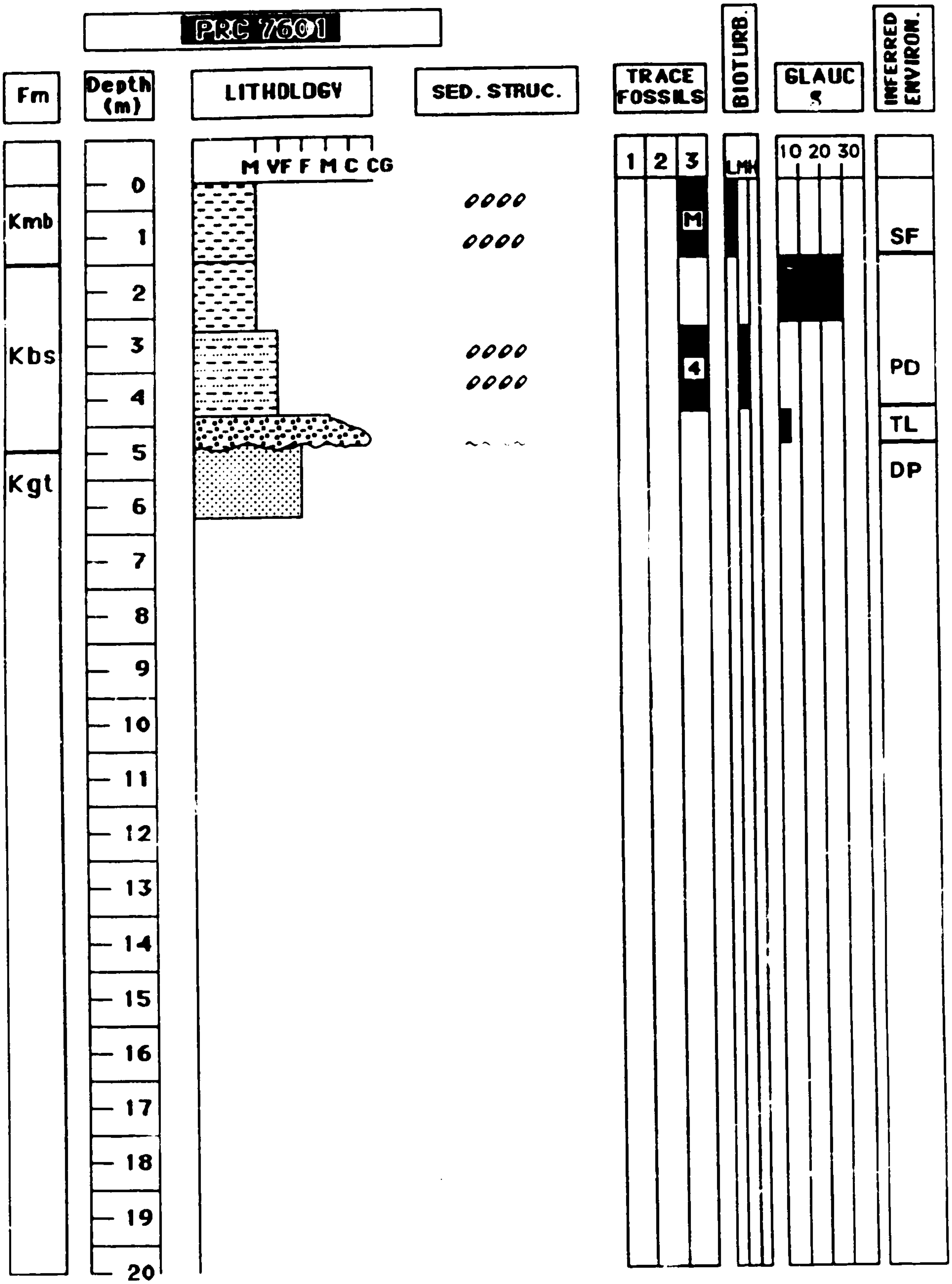
60
storm
Kbs
Kgt
DM
TL
DP


PRC 7601
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Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
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INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
Kmb
Kbs
Kgt
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2
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M
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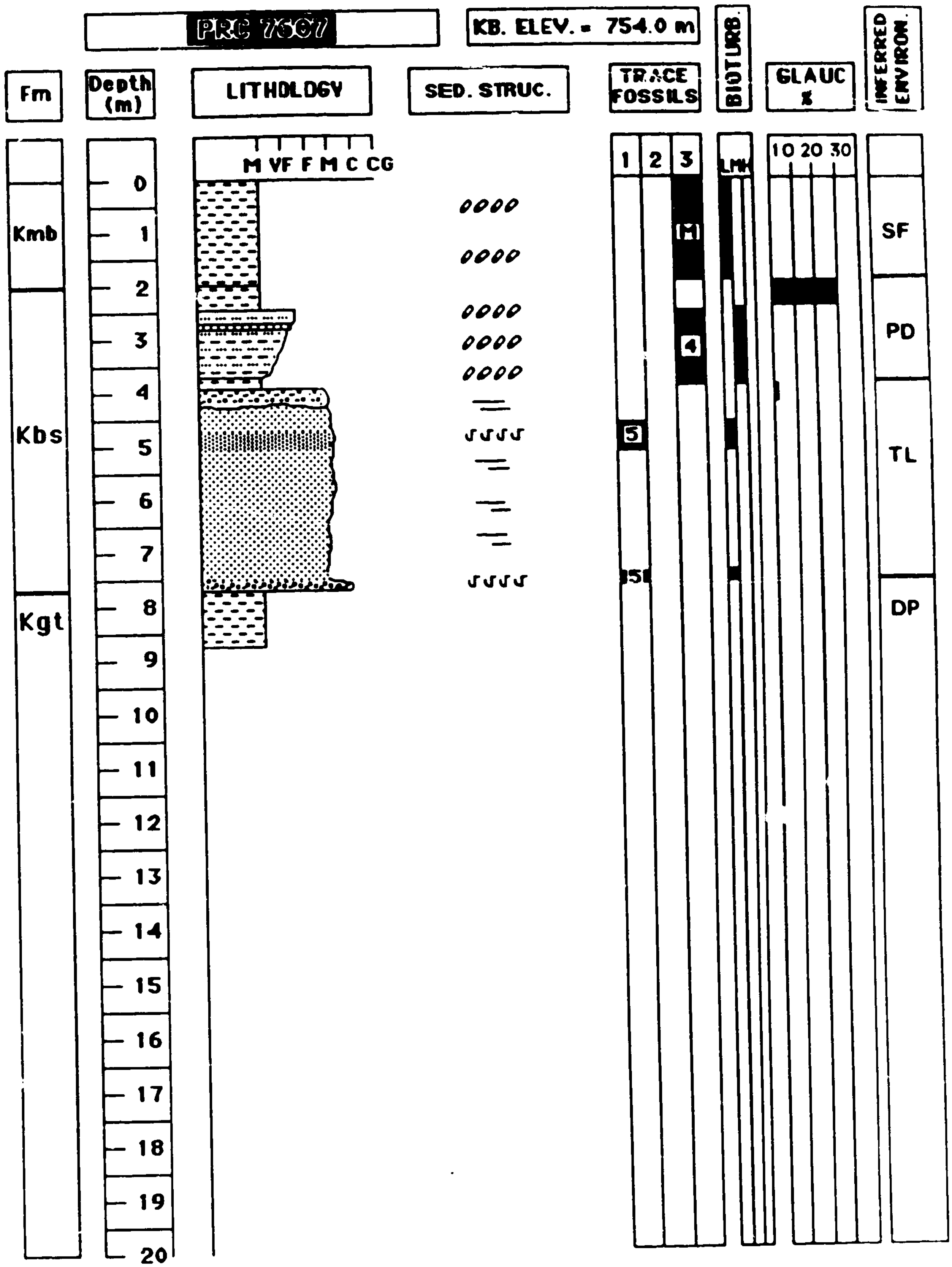
SF
PD
TL
DP


PRC 7607
KB. ELEV. = 754.0 m
Fm
Depth (m)
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TRACE FOSSILS
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GLAUC %
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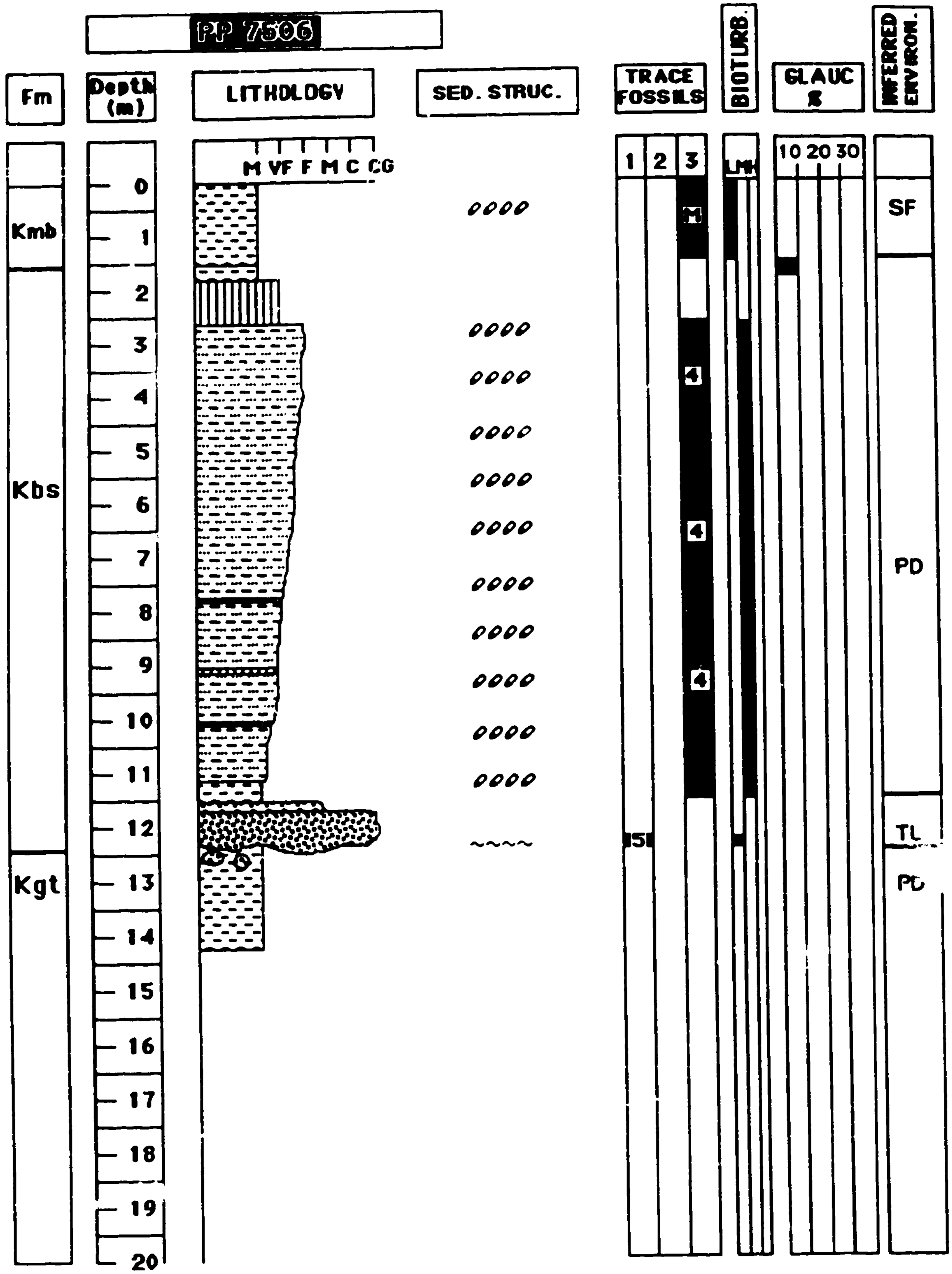
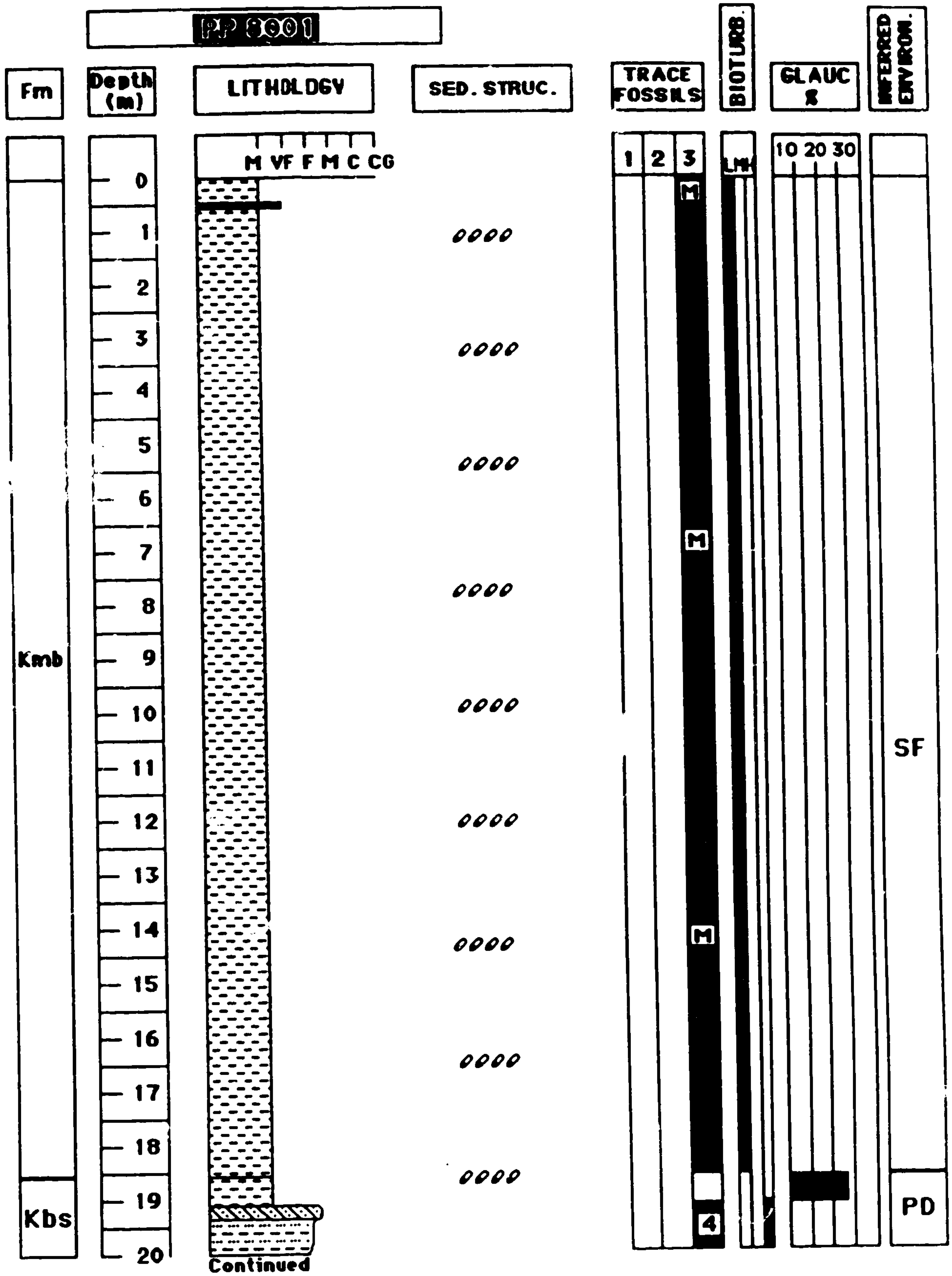
1 2 3
L M H
10 20 30
Kmb
Kbs
Kgt
SF
PD
TL
DP



PP 7506
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
Kmb
Kbs
Kgt
SF
PD
TL
PD


PP 8001
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
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1 2 3
LMH
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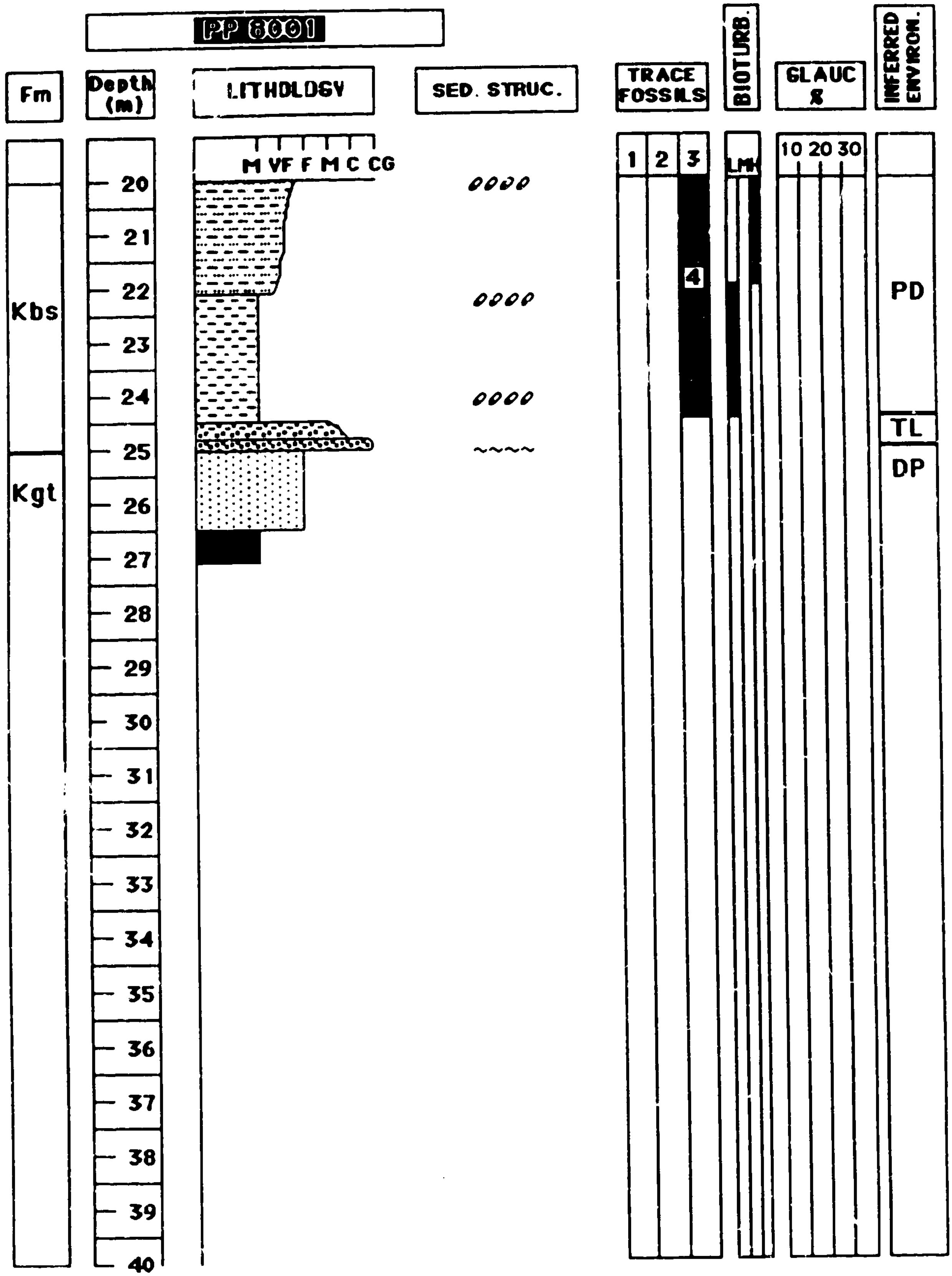
Kmb
Kbs
M
M
M
4
SF
PD
Continued



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Kgt
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PD
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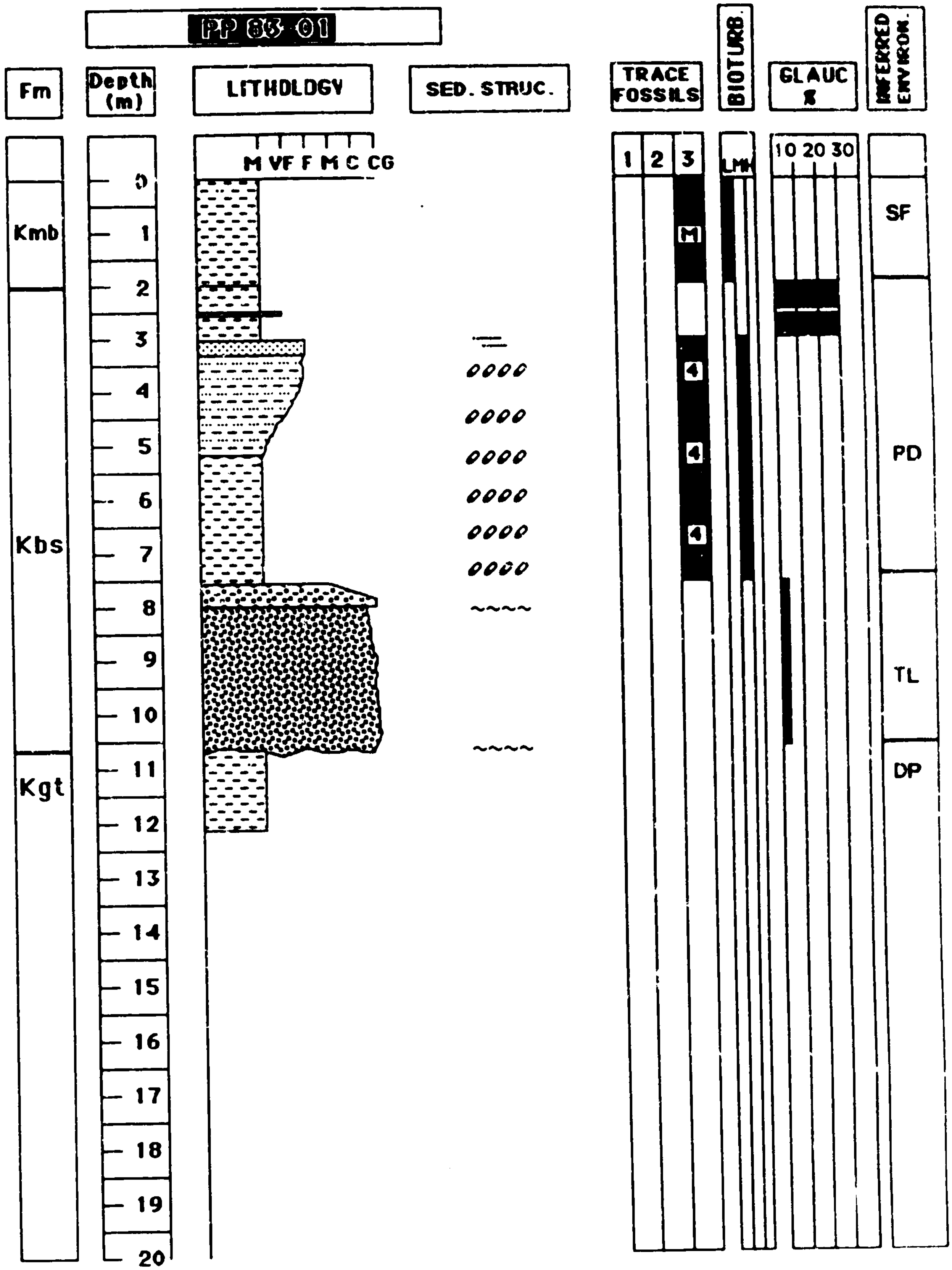

PP 83-01
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1 2 3
L M H
10 20 30
Kmb
Kbs
Kgt
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1
2
3
4
5
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M
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SF
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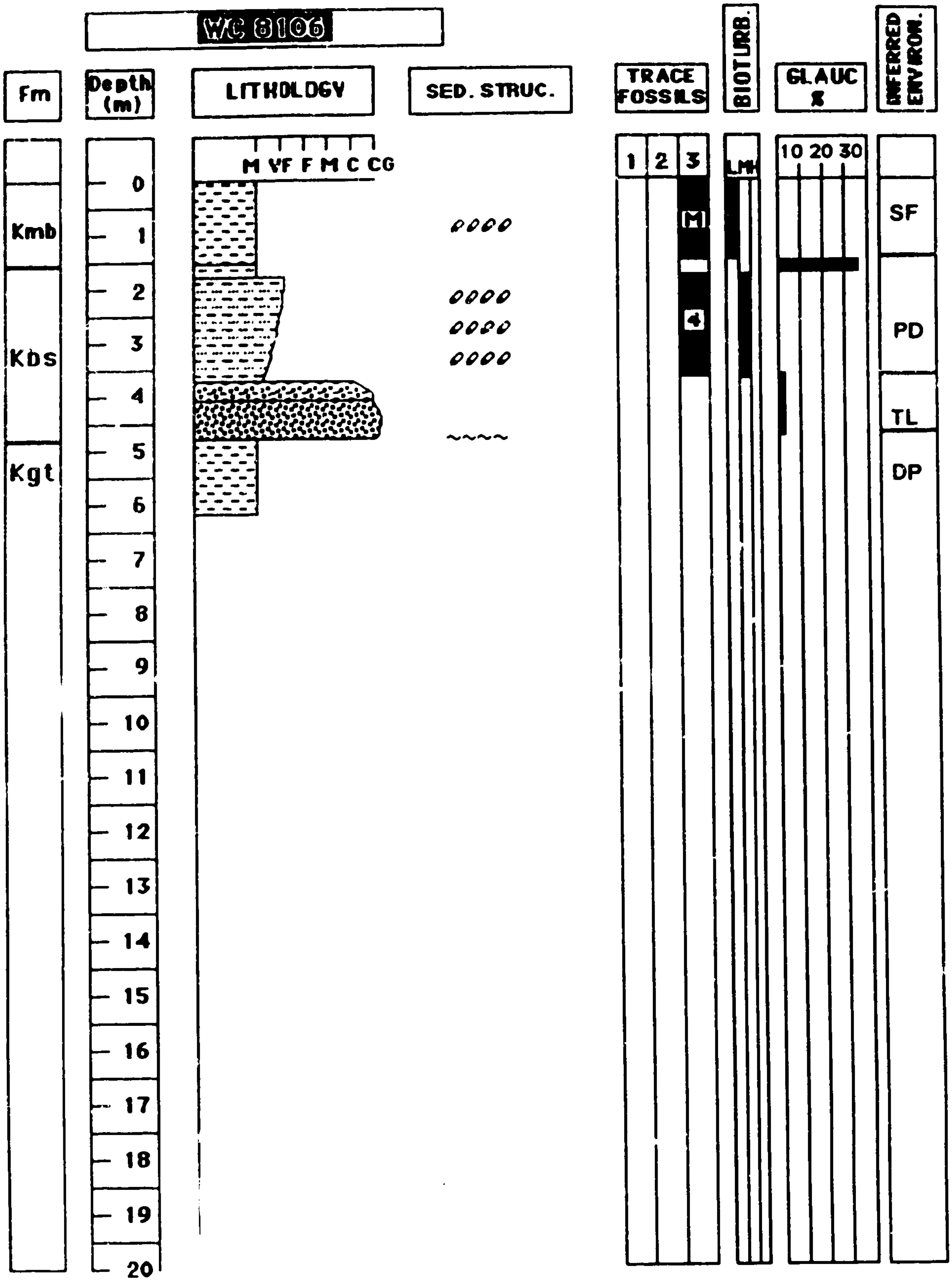
DP



WC 8106
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TRACE FOSSILS
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INFERRED ENVIRON.
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1 2 3
L M H
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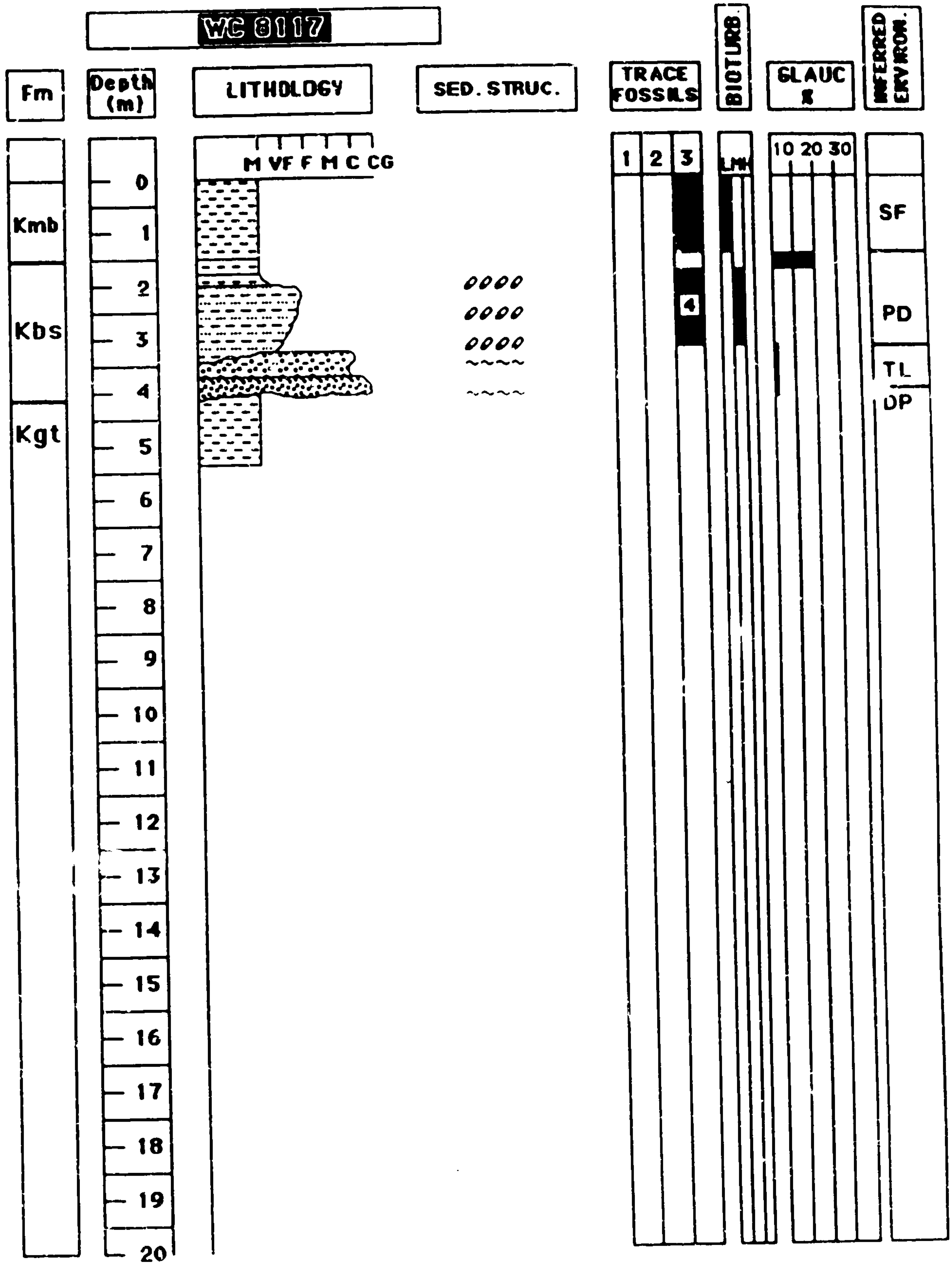
Kmb
Kbs
Kgt
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
M
4
SF
PD
TL
DP



WC 8117
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4
LMH
10 20 30
Kmb
Kbs
Kgt
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SF
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TL
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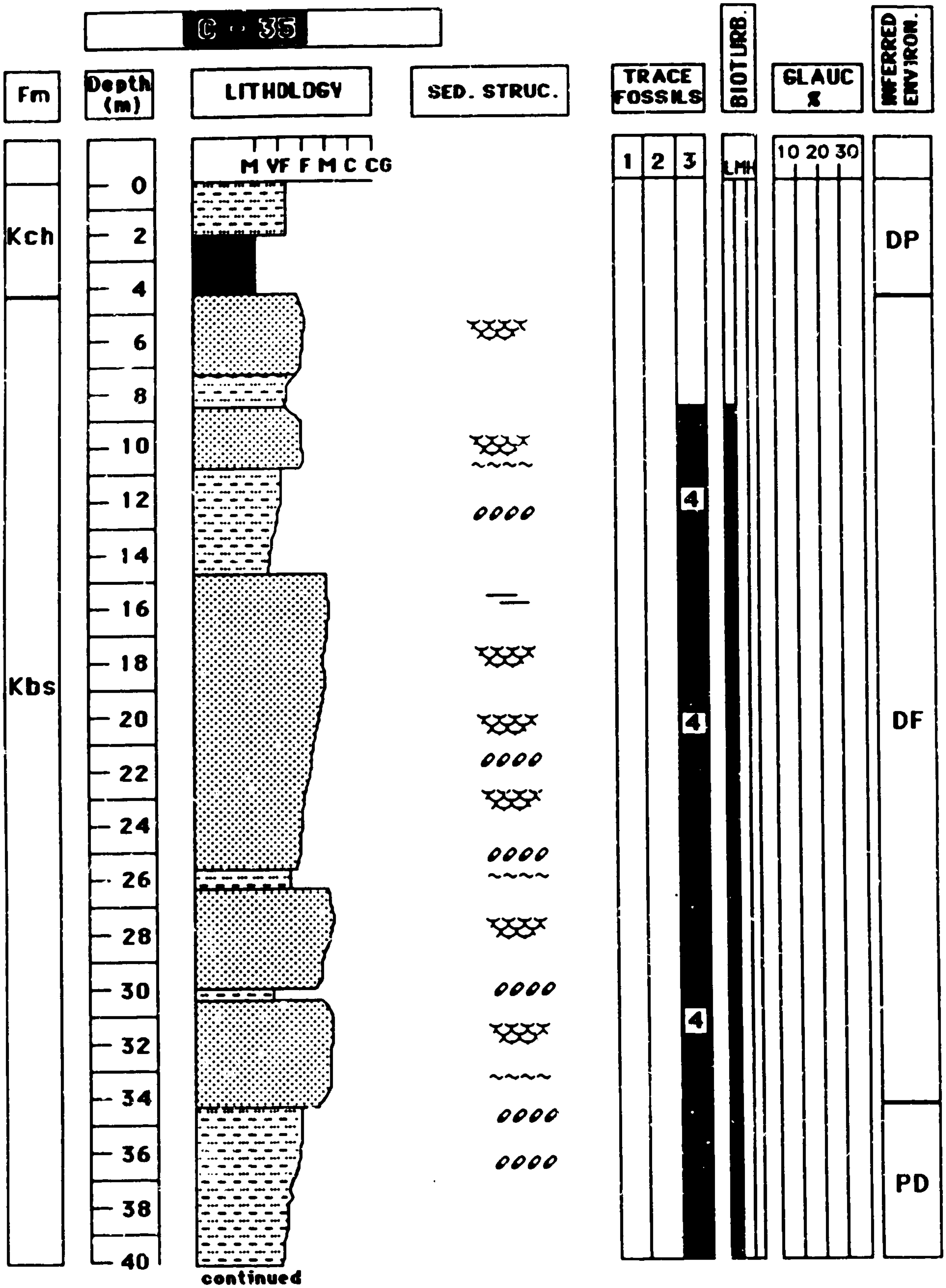

C - 35
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
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1 2 3
L M H
10 20 30
Kch
Kbs
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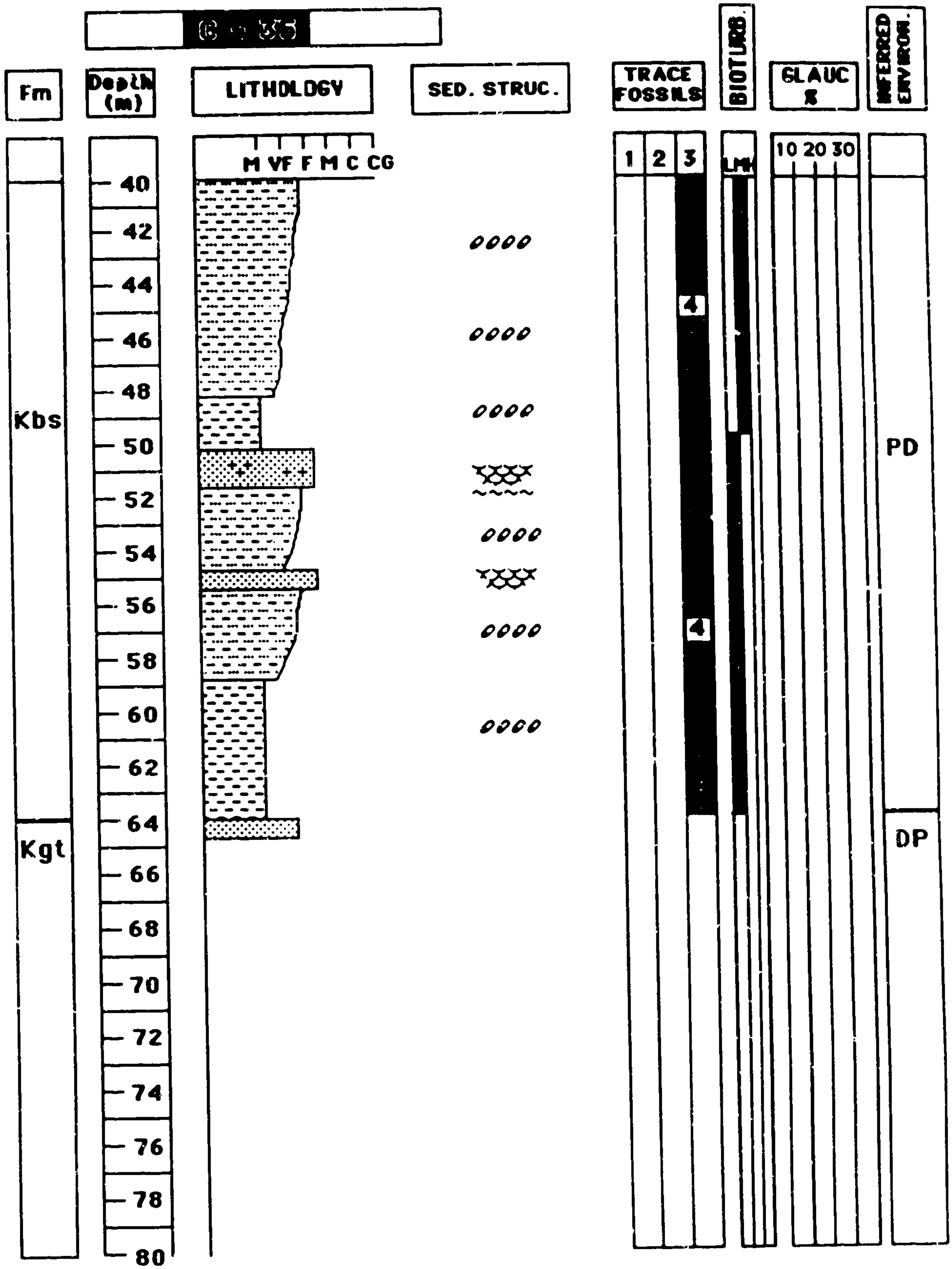
DF
PD
continued



C - 35
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Depth (m)
LITHOLOGY
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TRACE FOSSILS
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INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
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42
44
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48
50
52
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66
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70
72
74
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78
80
Kbs
Kgt
4
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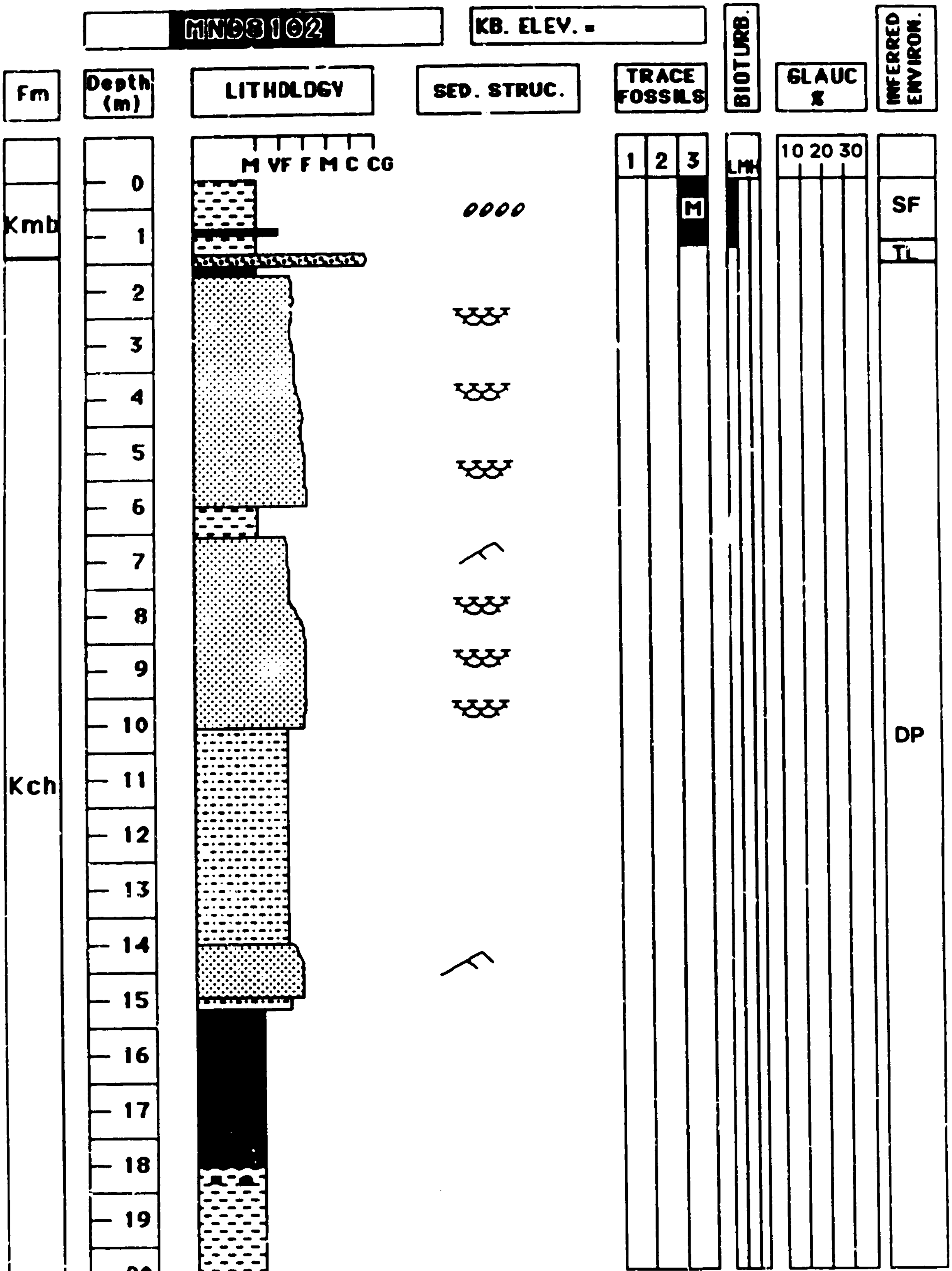
PD
DP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1 2 3
LMH
10 20 30
Kmb
Kch
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M
SF
TL
DP

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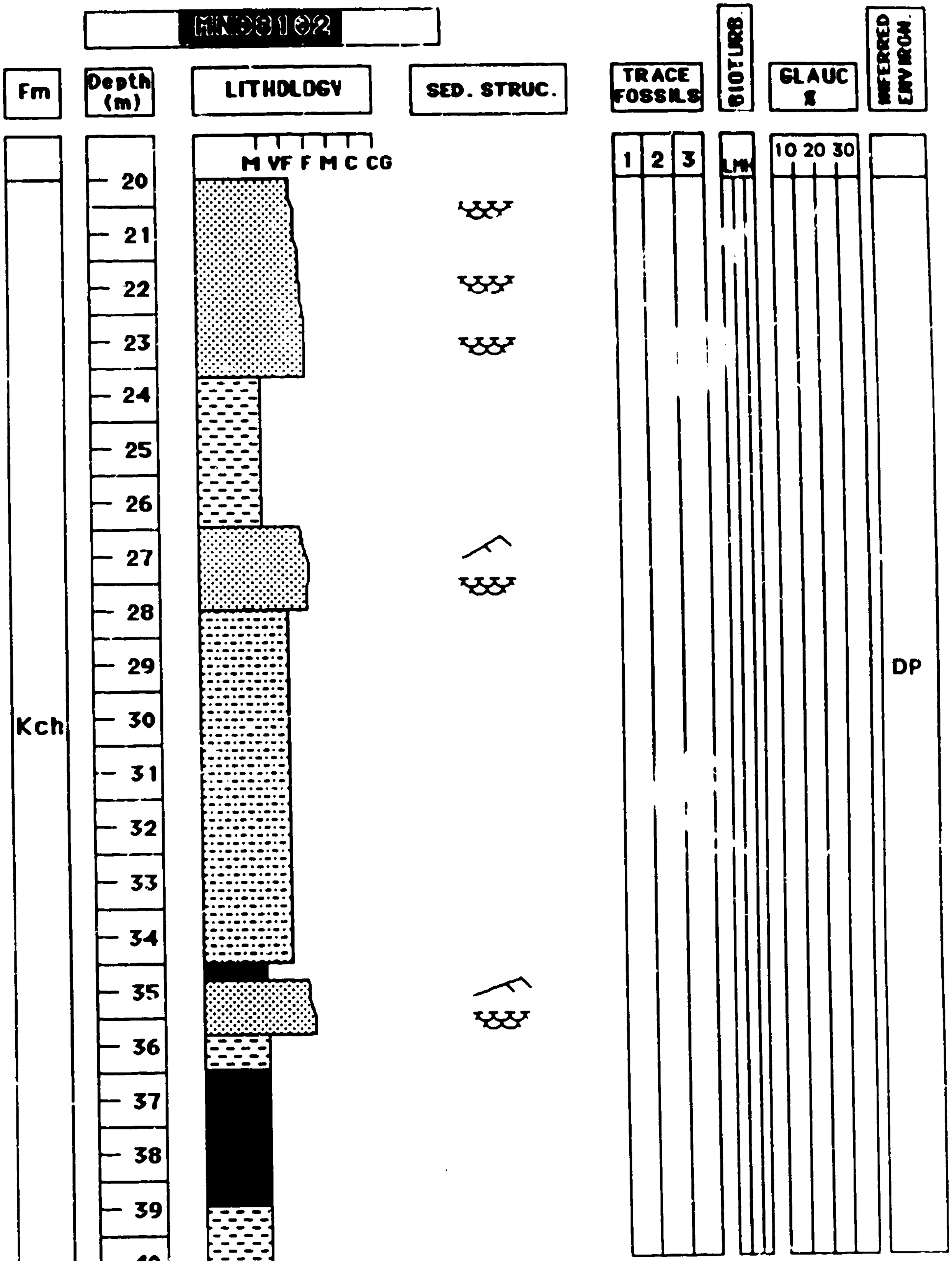
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
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1 2 3
L M H
10 20 30
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33
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35
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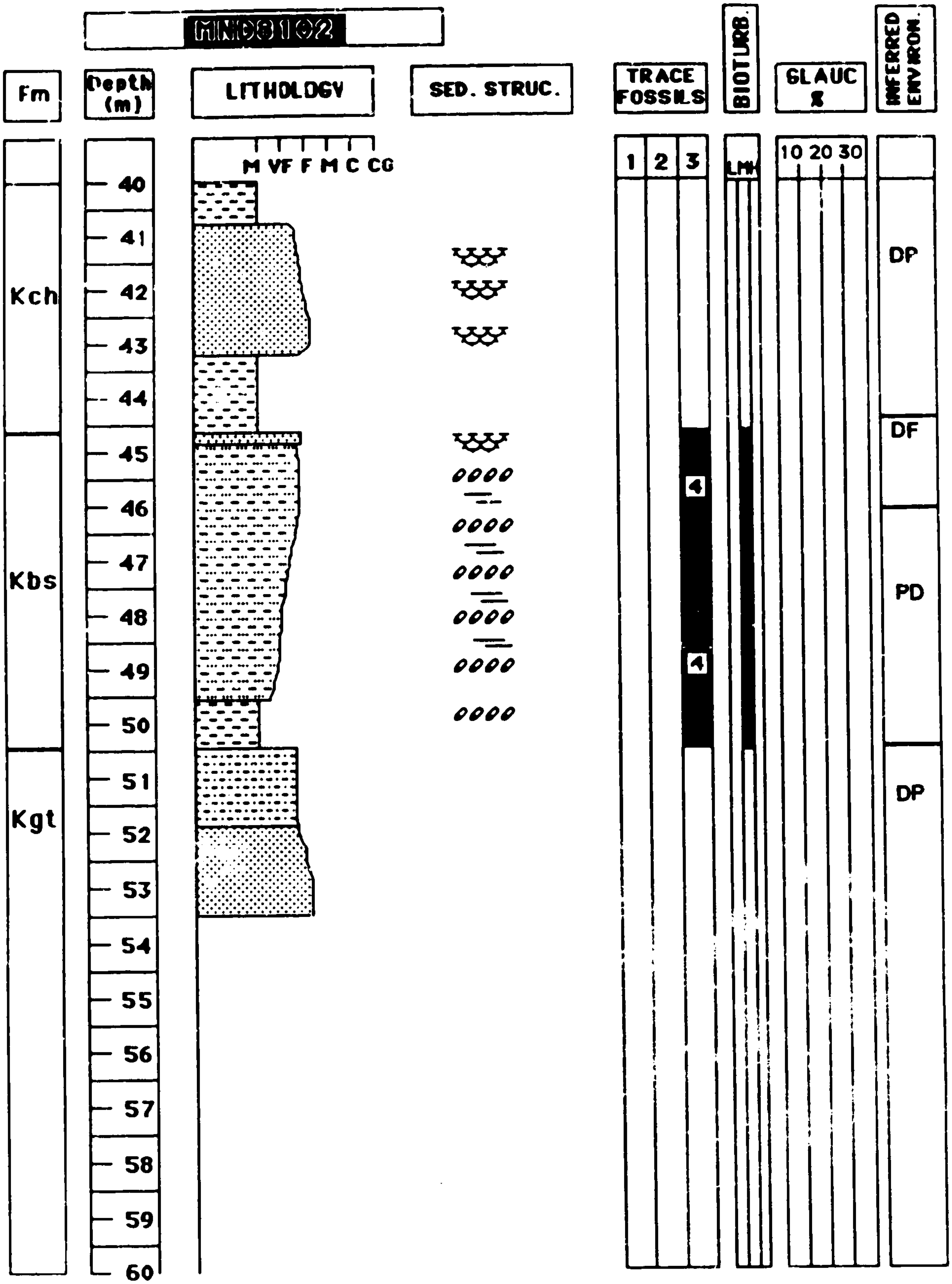
37
38
39
40
Kch
DP
(Continued)


MND8102
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
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GLAUC %
INFERRED ENVIRON.
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1 2 3
LMH
10 20 30
Kch
Kbs
Kgt
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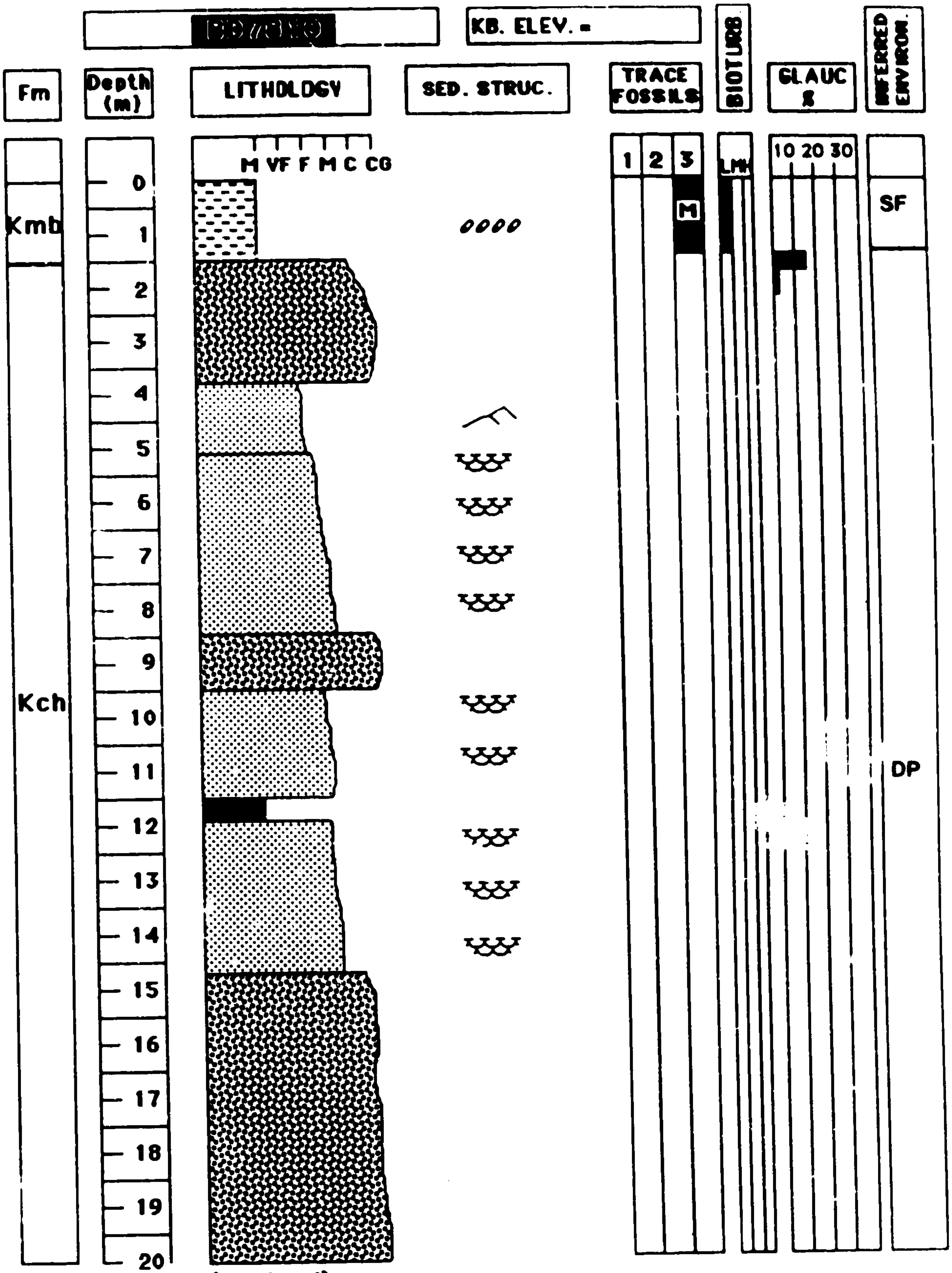
DP
DF
PD
DP



KB. ELEV. =
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB
GLAUC %
INFERRED ENVIRON.
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1 2 3
LMH
10 20 30
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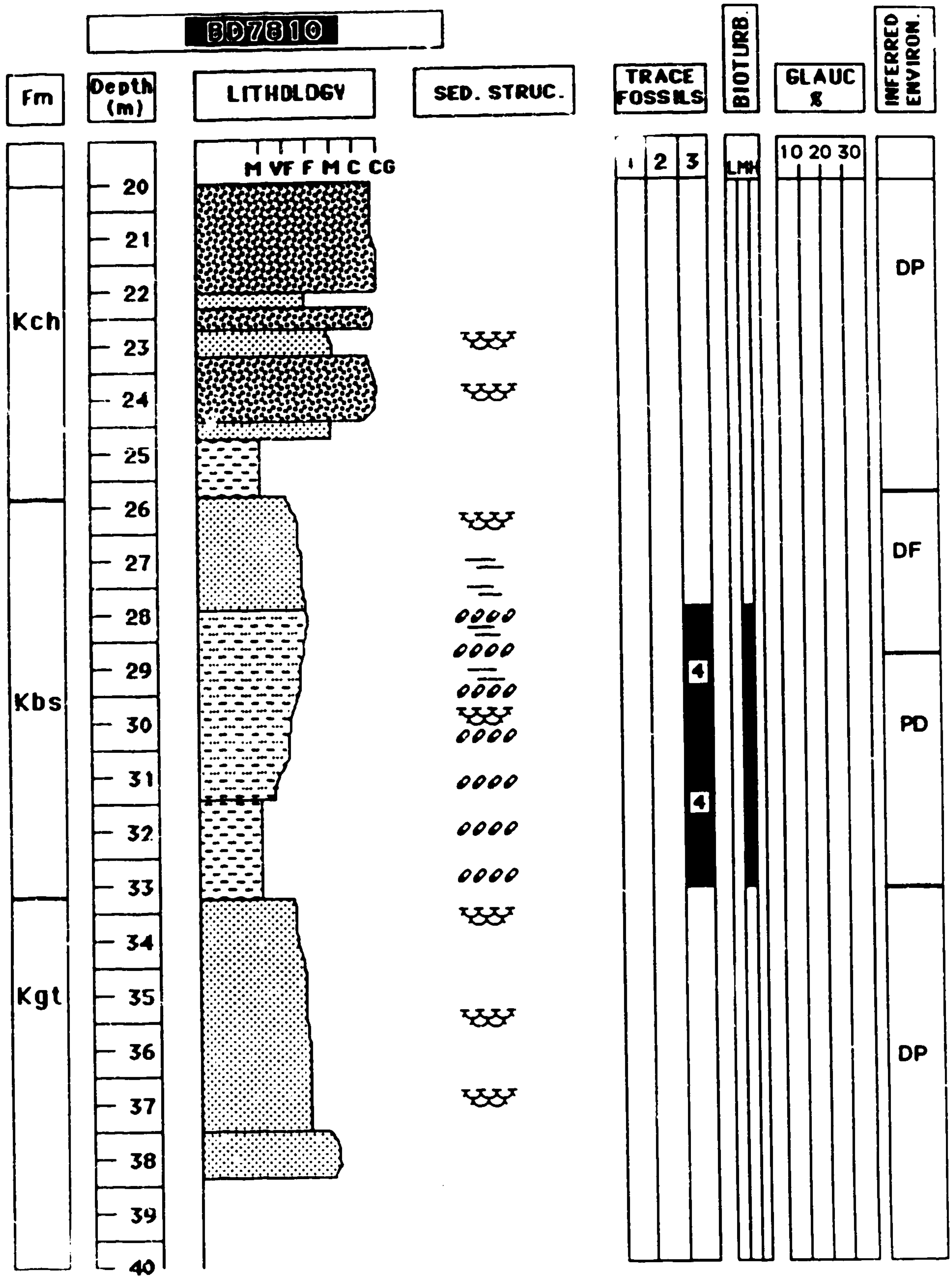
Kmb
Kch
M
SF
DP

(Continued)



BD7810
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
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21
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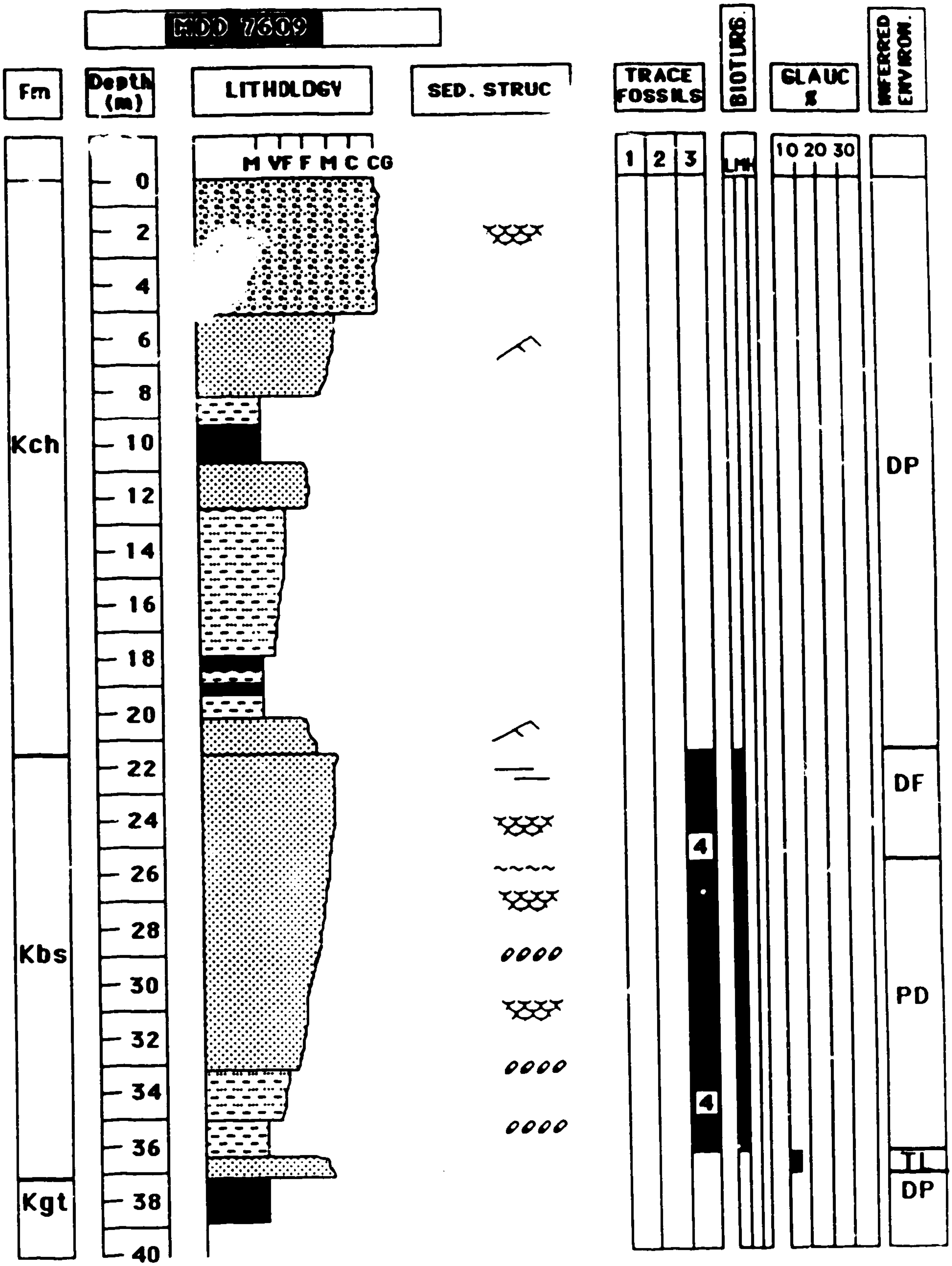
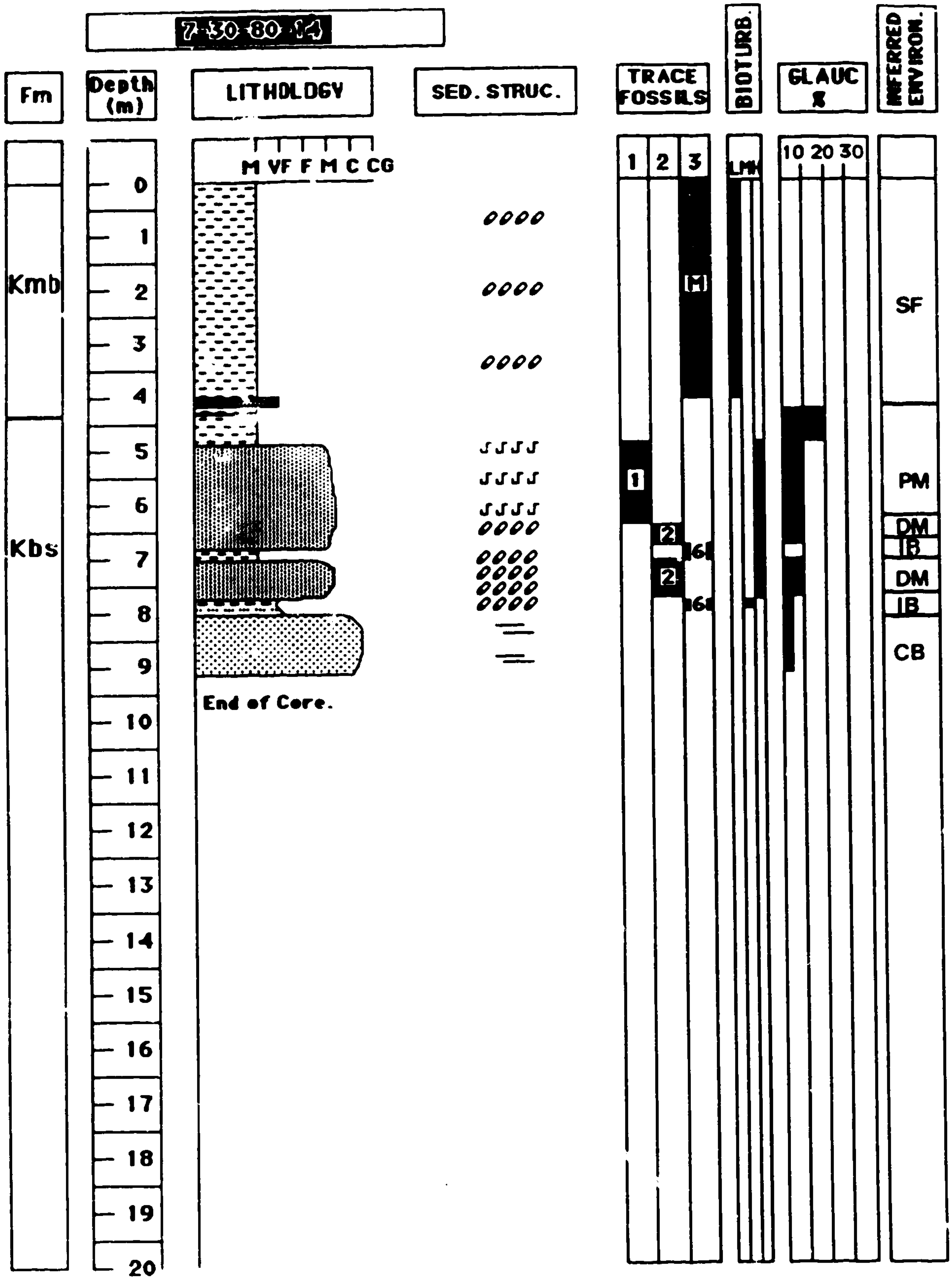
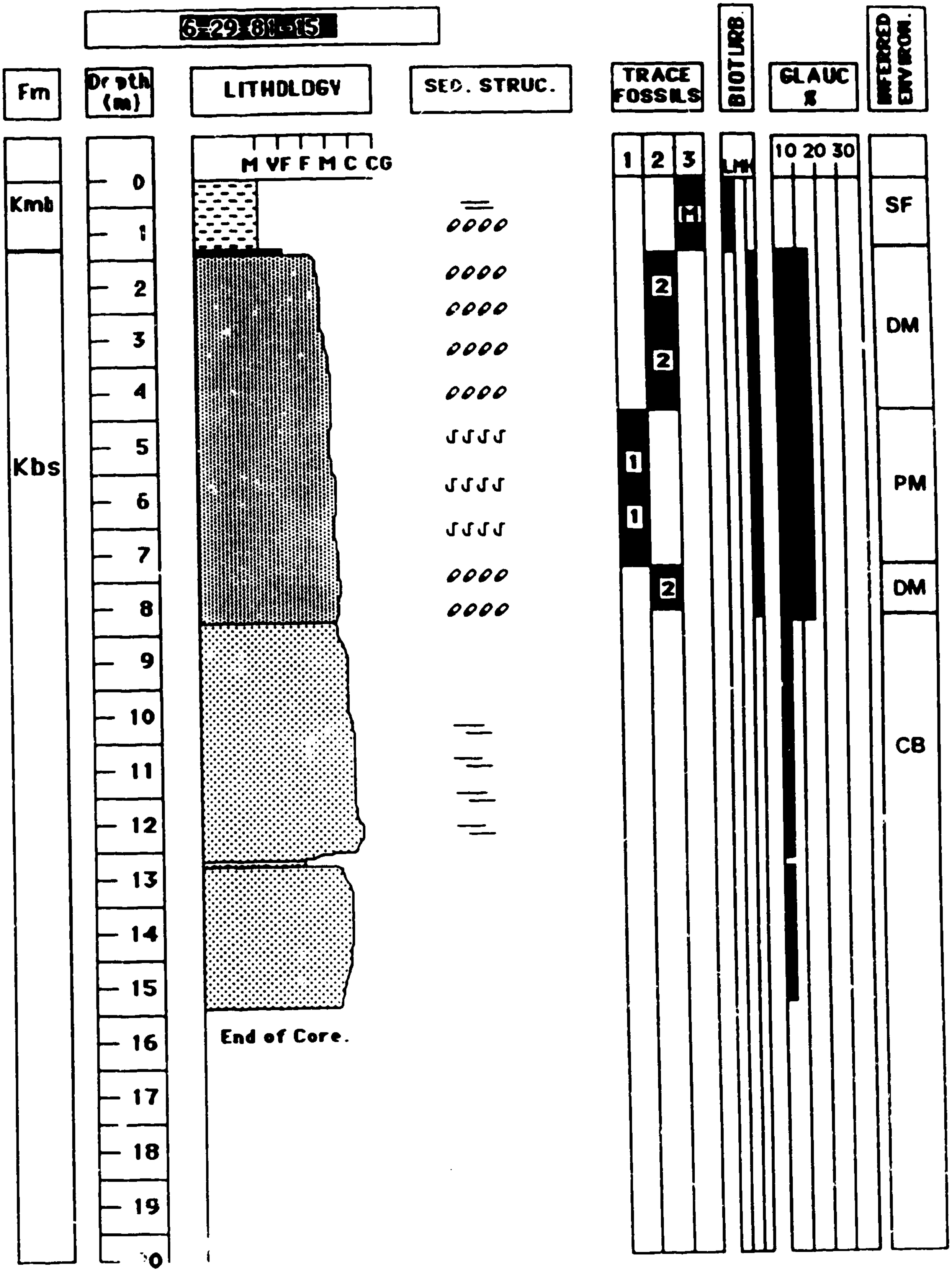
36
37
38
39
40
Kch
Kbs
Kgt
4
4
DP
DF
PD
DP


MDD 7609
Fm
Depth (m)
LITHOLOGY
SED. STRUC
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
0
2
4
6
8
10
12
14
16
18
20
22
24
26
28
30
32
34
36
38
40
Kch
Kbs
Kgt
1 2 3
4
4
LMH
10 20 30
DP
DF
PD
TL
DP


7-30-80-14
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
Kmb
Kbs
End of Core.
H
1
2
2
16
16
SF
PM
DM
IB
DM
IB
CB


6-29-81-15
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
Kmb
Kbs
0
1
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3
4
5
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7
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10
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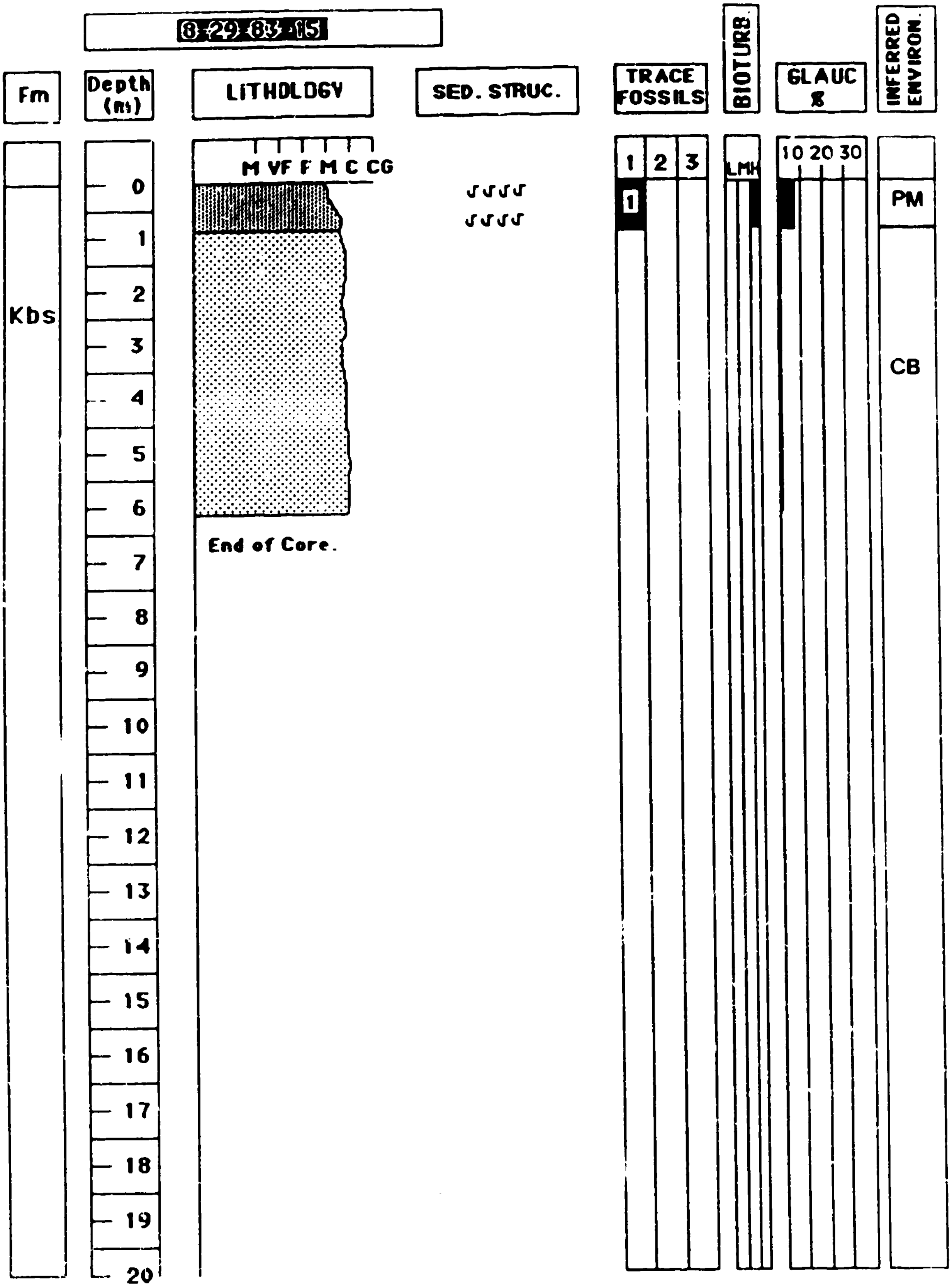
12
13
14
15
16
17
18
19
20
End of Core.
SF
DM
PM
DM
CB


8-29-83-15
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
Kbs
0
1
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3
4
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14
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16
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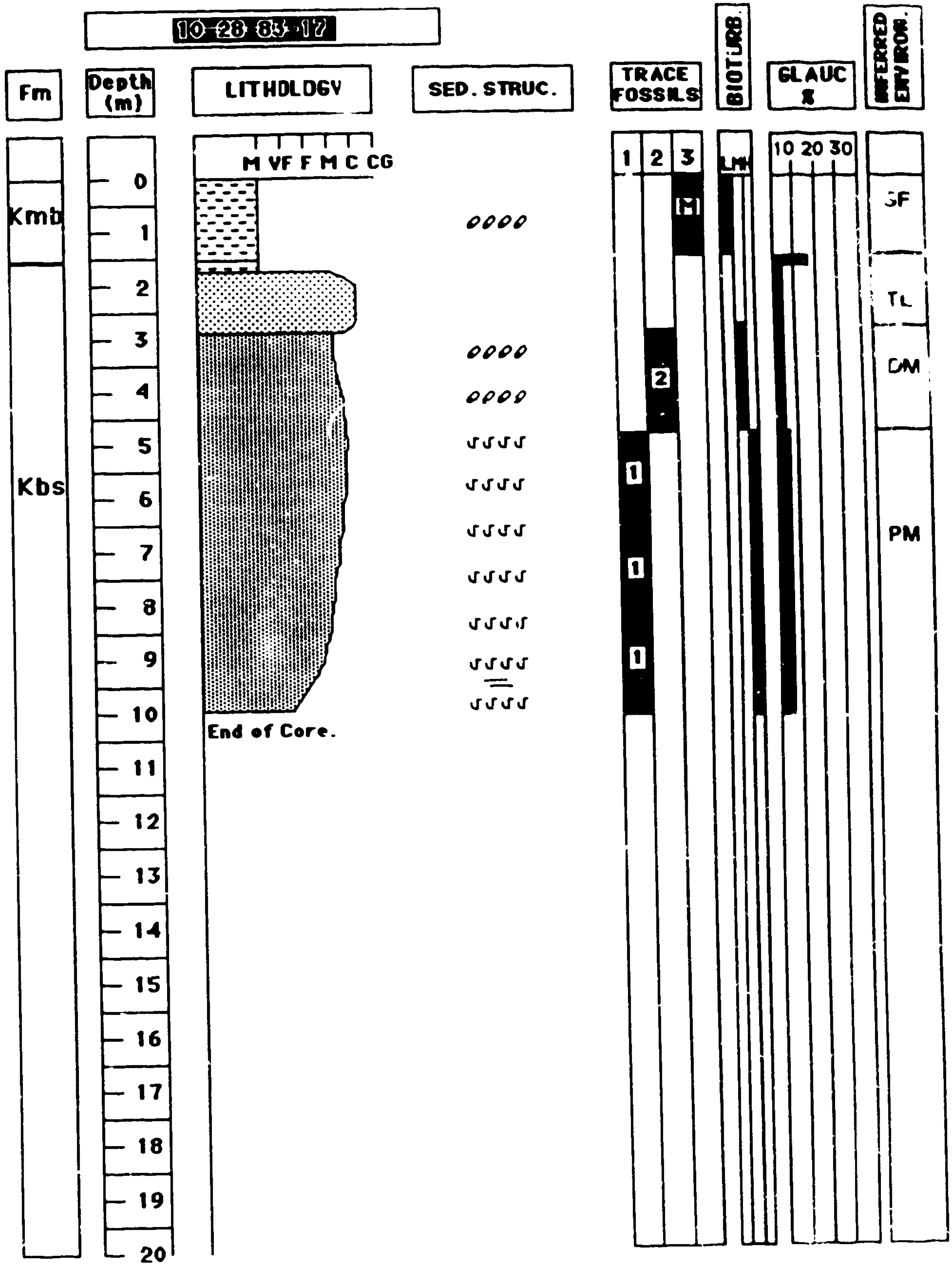
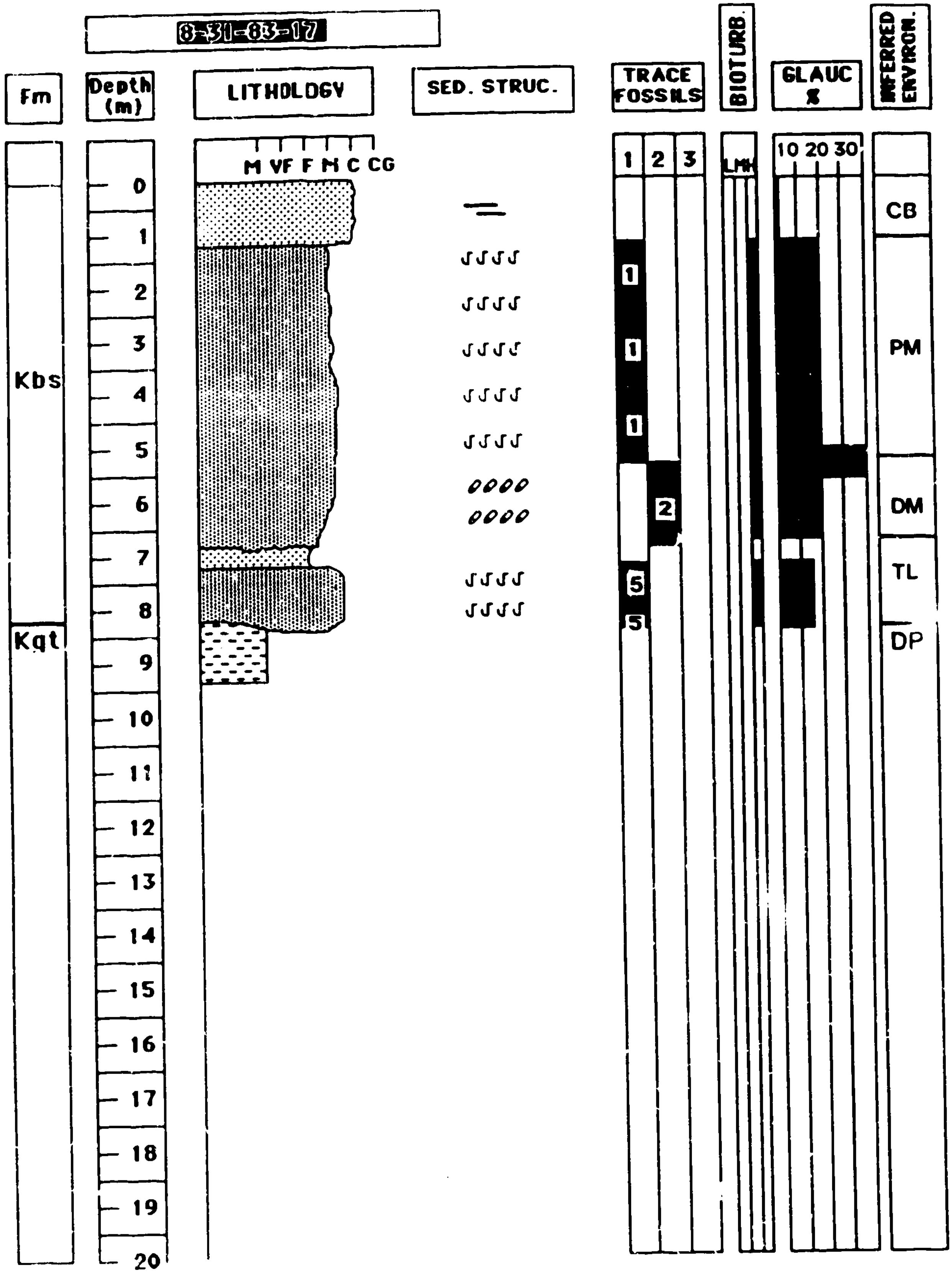
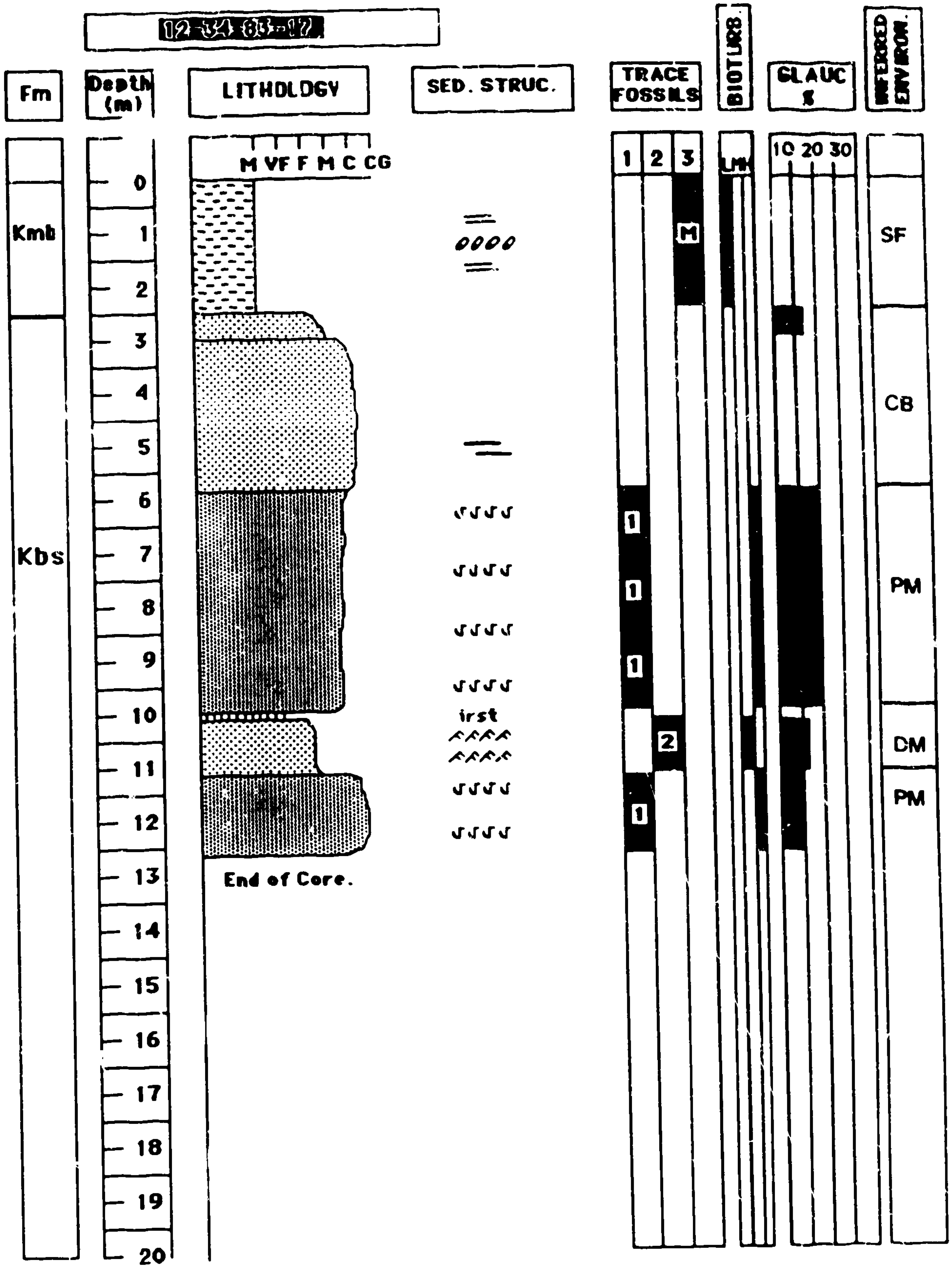
18
19
20
End of Core.
1
PM
CB



10-28-83-17
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
Kmb
Kbs
End of Core.
M
2
1
1
1
SF
TL
DM
PM



8-31-83-17
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
Kbs
Kqt
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
CB
PM
DM
TL
DP


12-34-83-17
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
0
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4
5
6
7
8
9
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15
16
17
18
19
20
Kmb
Kbs
irst
M
1
1
1
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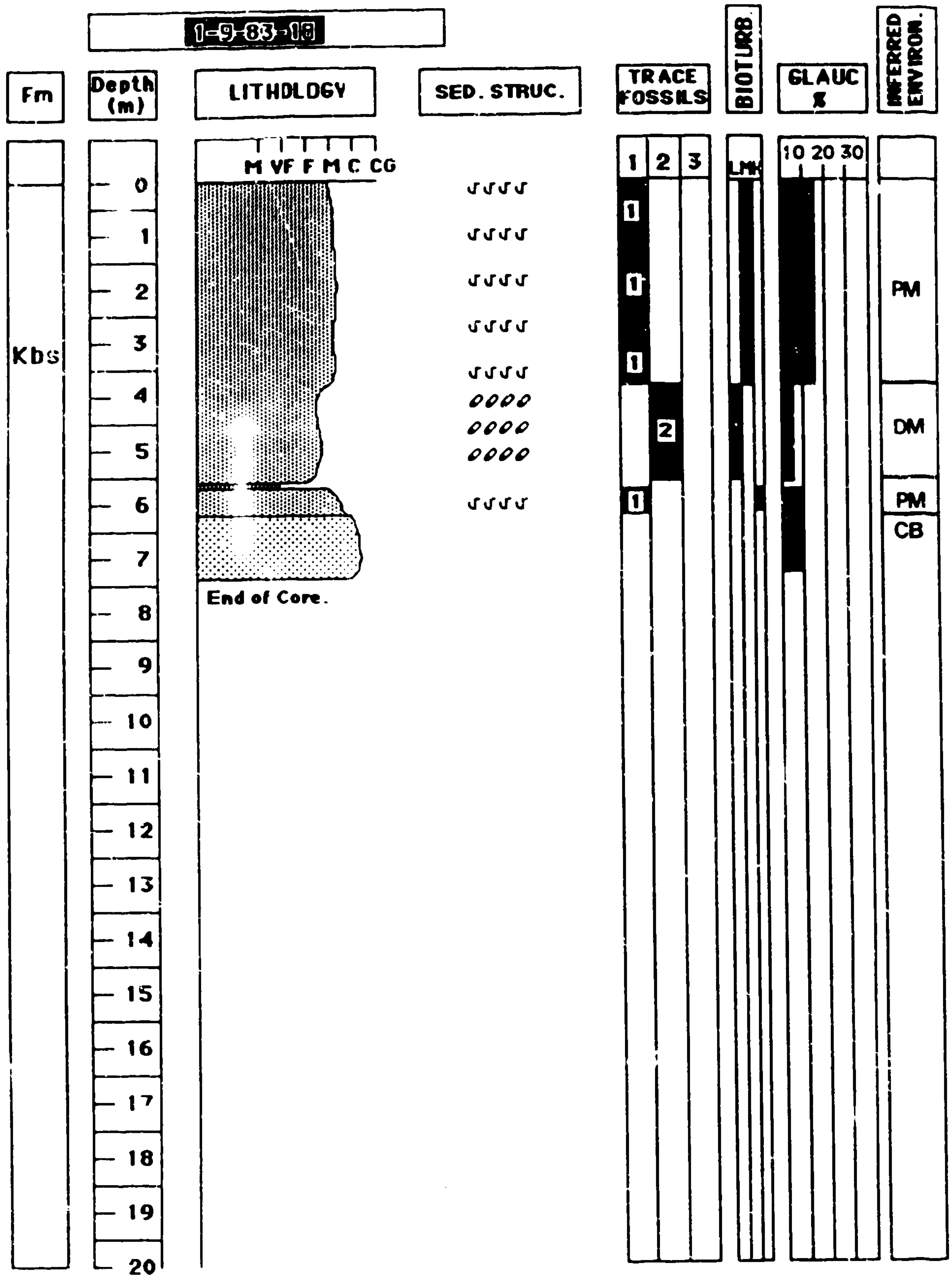
1
SF
CB
PM
DM
PM
End of Core.


1-9-83-18
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
0
1
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14
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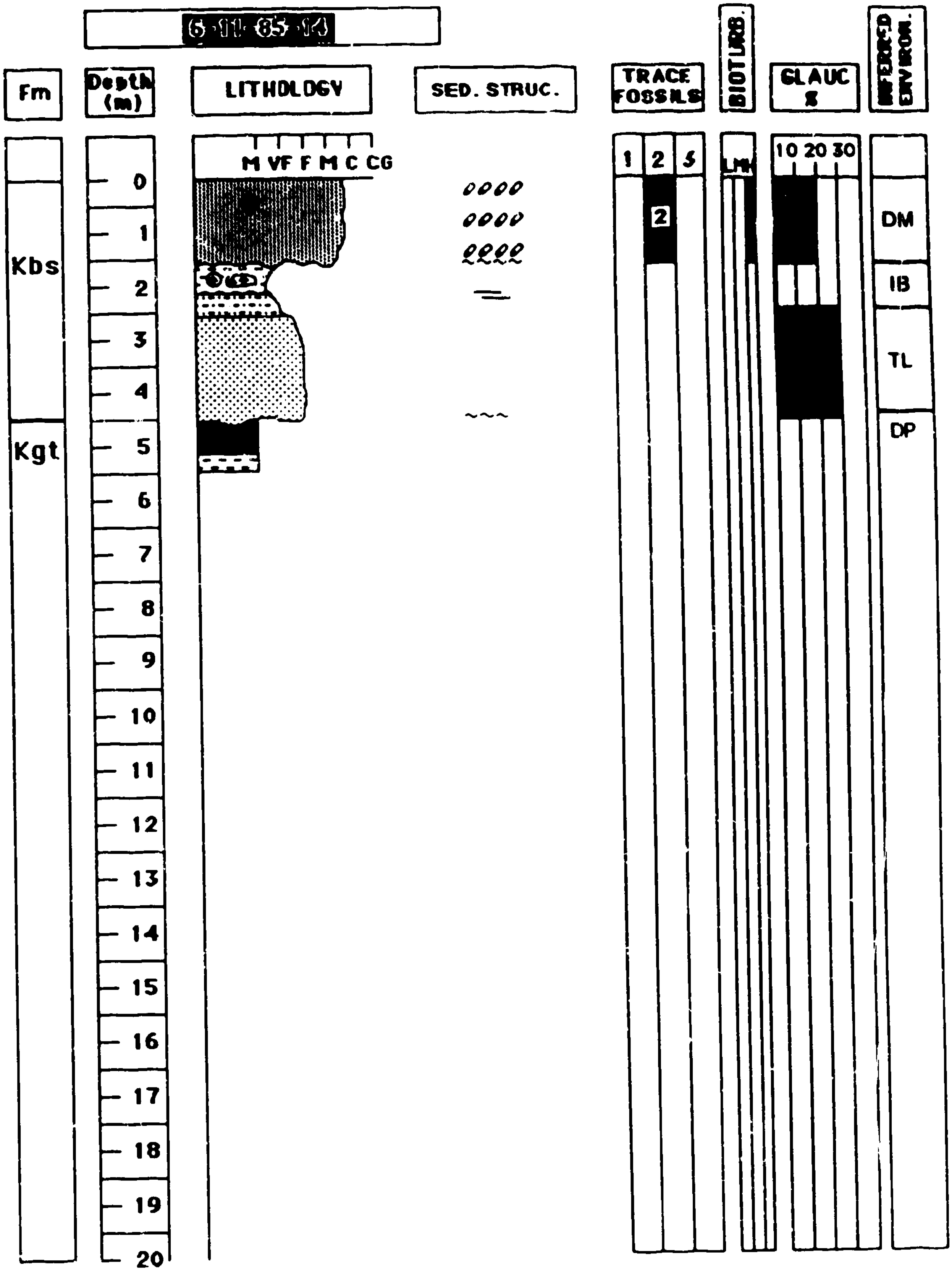
16
17
18
19
20
Kbs
End of Core.
PM
DM
PM
CB


6-11-85-14
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 5
L M H
10 20 30
Kbs
Kgt
0
1
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3
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14
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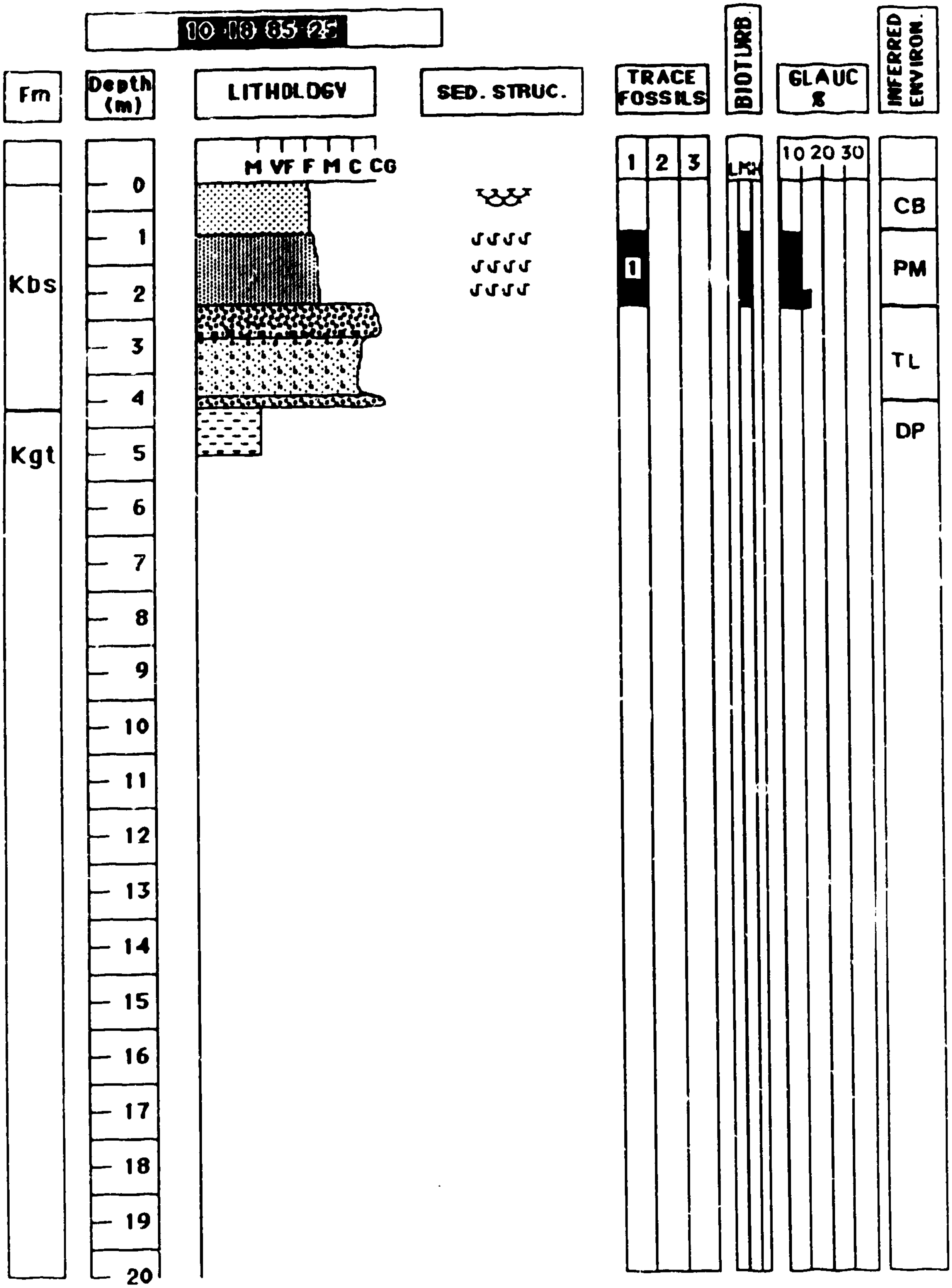
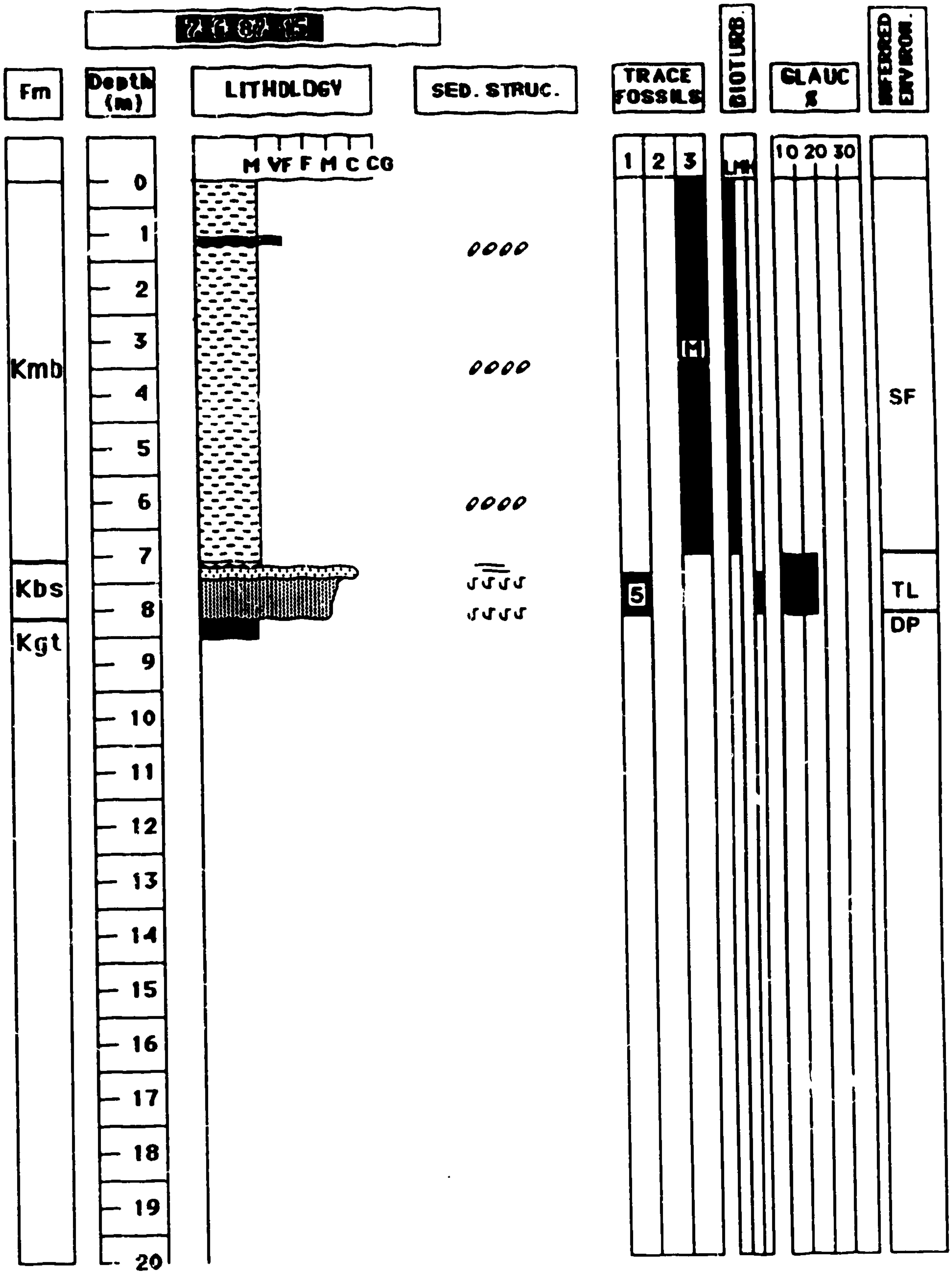
18
19
20
2
DM
IB
TL
DP


10-18-85-25
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
Kbs
Kgt
CB
PM
TL
DP



7-6-87-15
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
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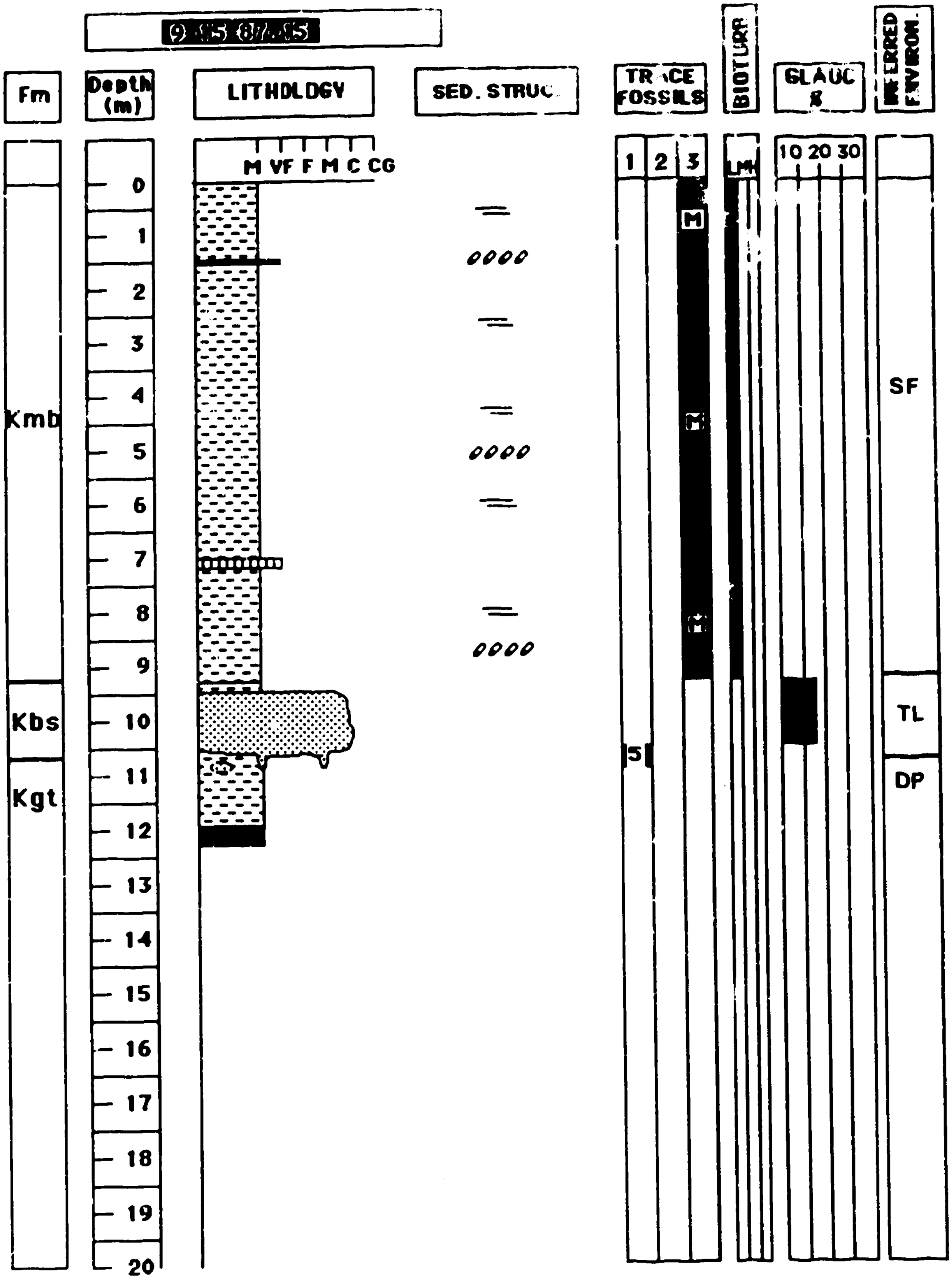
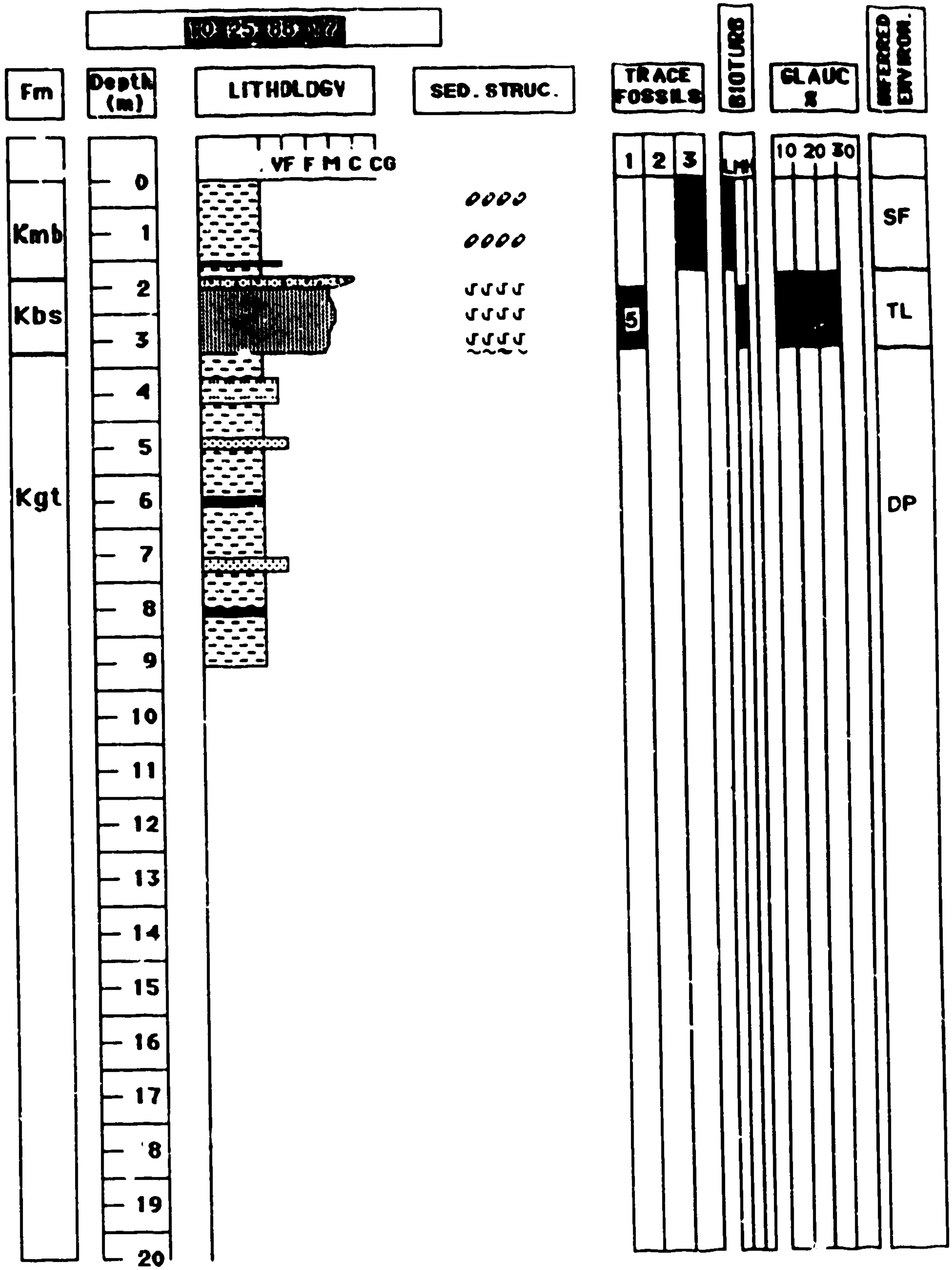
18
19
20
Kmb
Kbs
Kgt
1 2 3
LMH
10 20 30
M
5
SF
TL
DP


9 15 87-15
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
Kmb
Kbs
Kgt
M
5
SF
TL
DP



10 25 65 17
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB
GLAUC %
INFERRED ENVIRON.
VF F M C CG
1 2 3
L M H
10 20 30
Kmb
Kbs
Kgt
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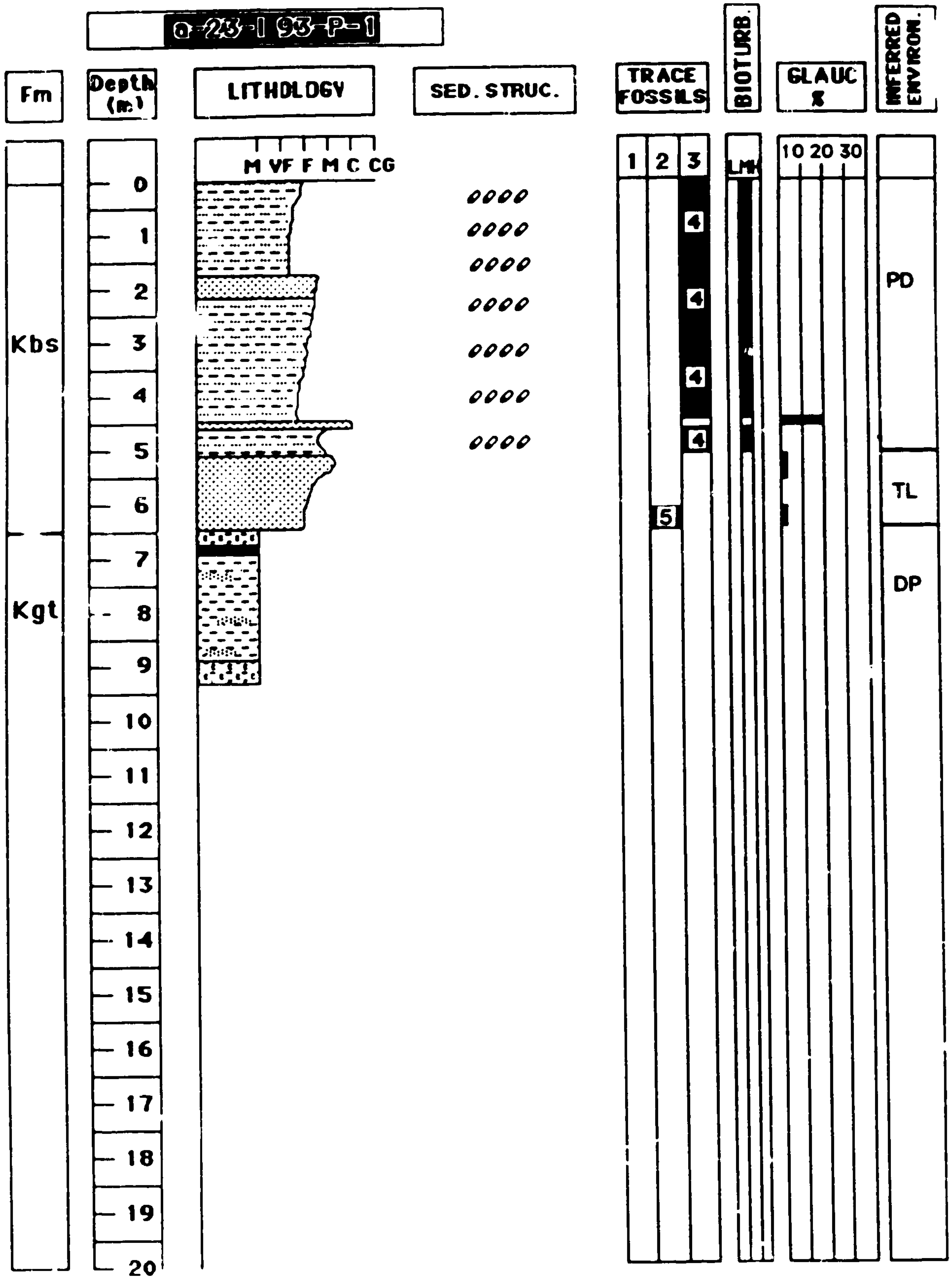
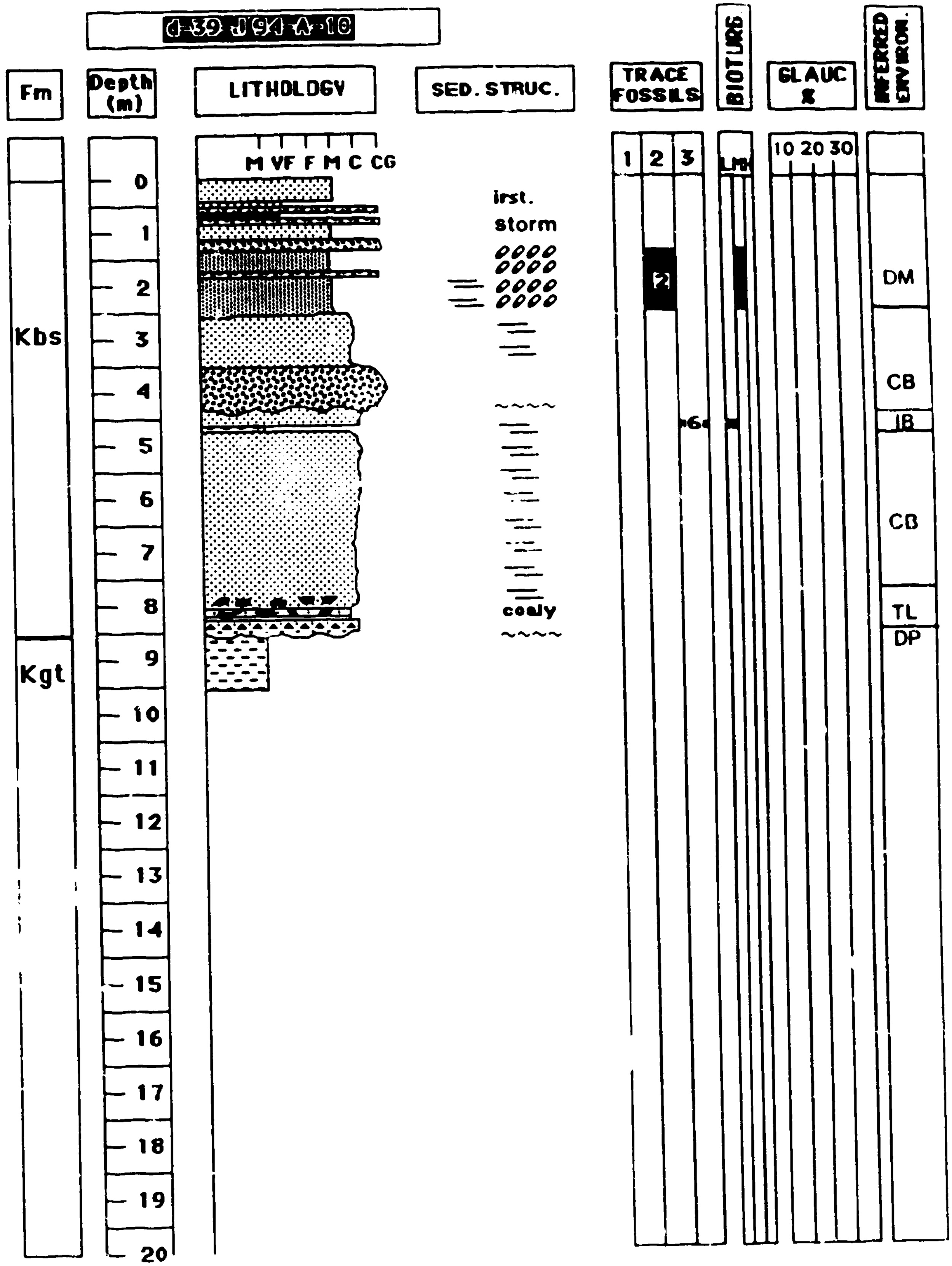
SF
TL
DP


a-23-I 93-P-1
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
Kbs
Kgt
4
4
4
4
5
PD
TL
DP


d-39-J 94-A-10
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
Kbs
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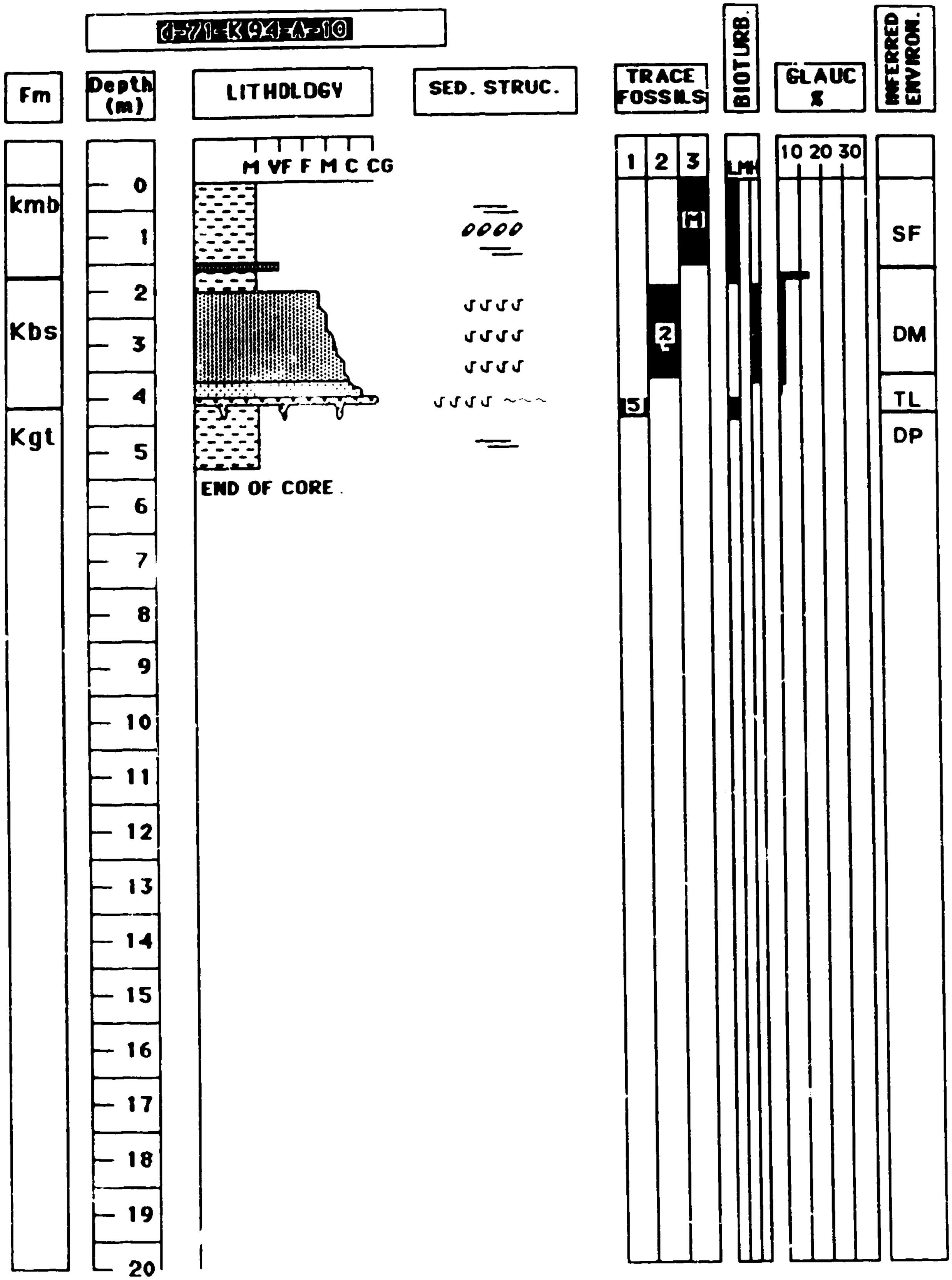
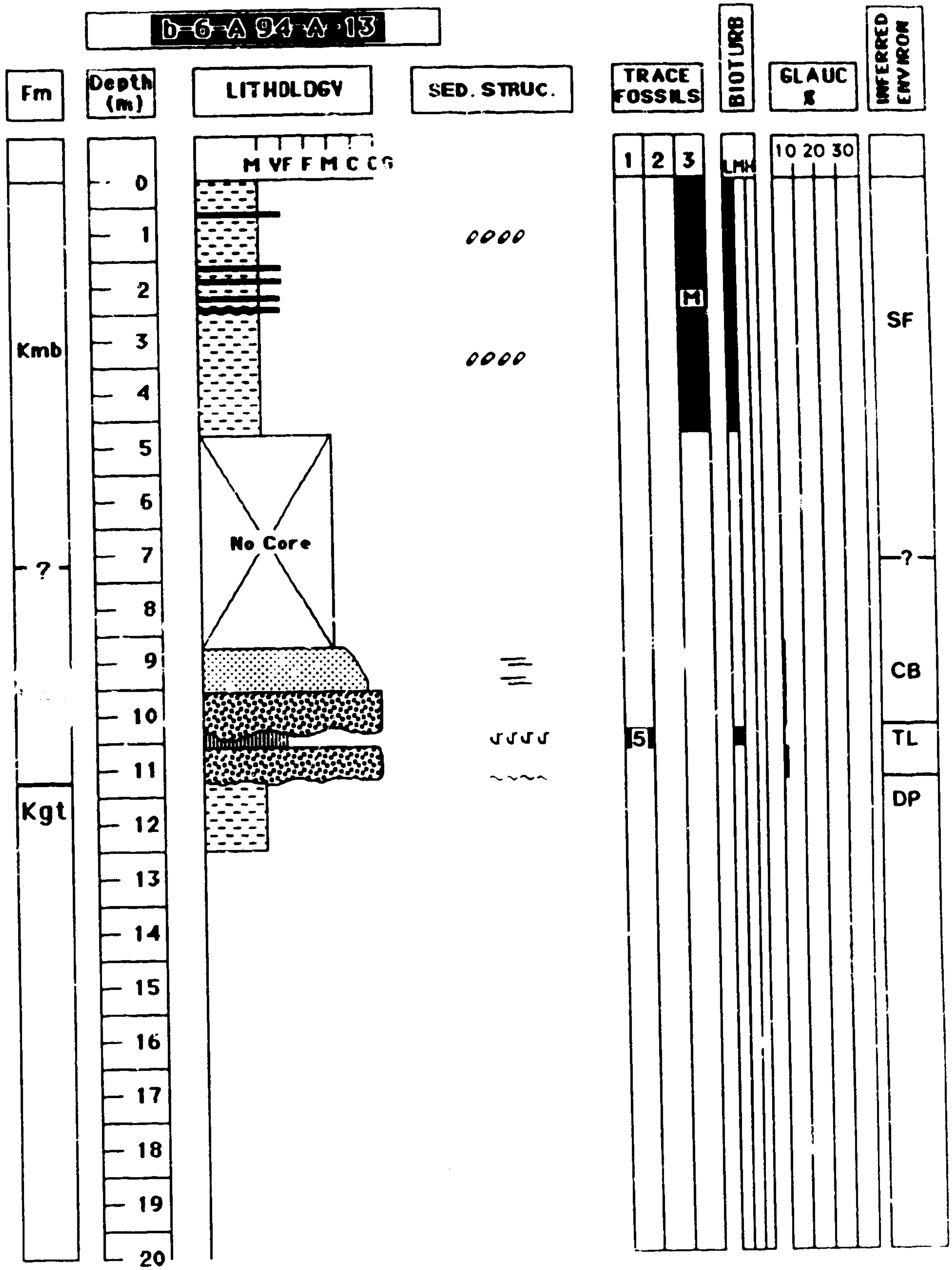
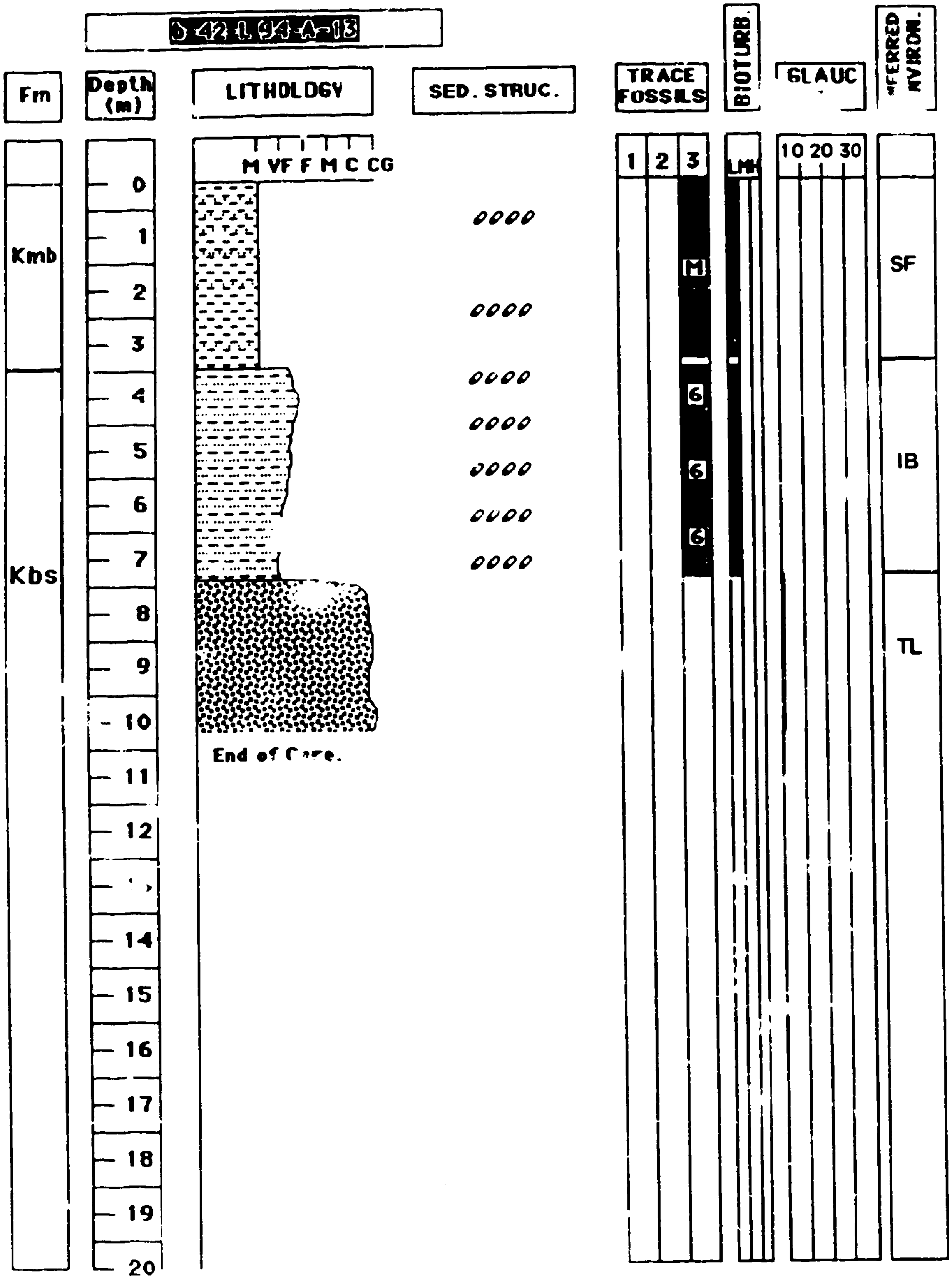

d-71-K 94-A-10
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
kmb
Kbs
Kgt
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
M
2
5
END OF CORE
SF
DM
TL
DP



b-6-A 94-A-13
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB
GLAUC %
INFERRED ENVIRON
M VF F M C CG
1 2 3
LMH
10 20 30
Kmb
Kgt
No Core
M
5
SF
CB
TL
DP



D-42-L/94-A-13
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC.
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
0
1
2
3
4
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6
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8
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16
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18
19
20
Kmb
Kbs
M
6
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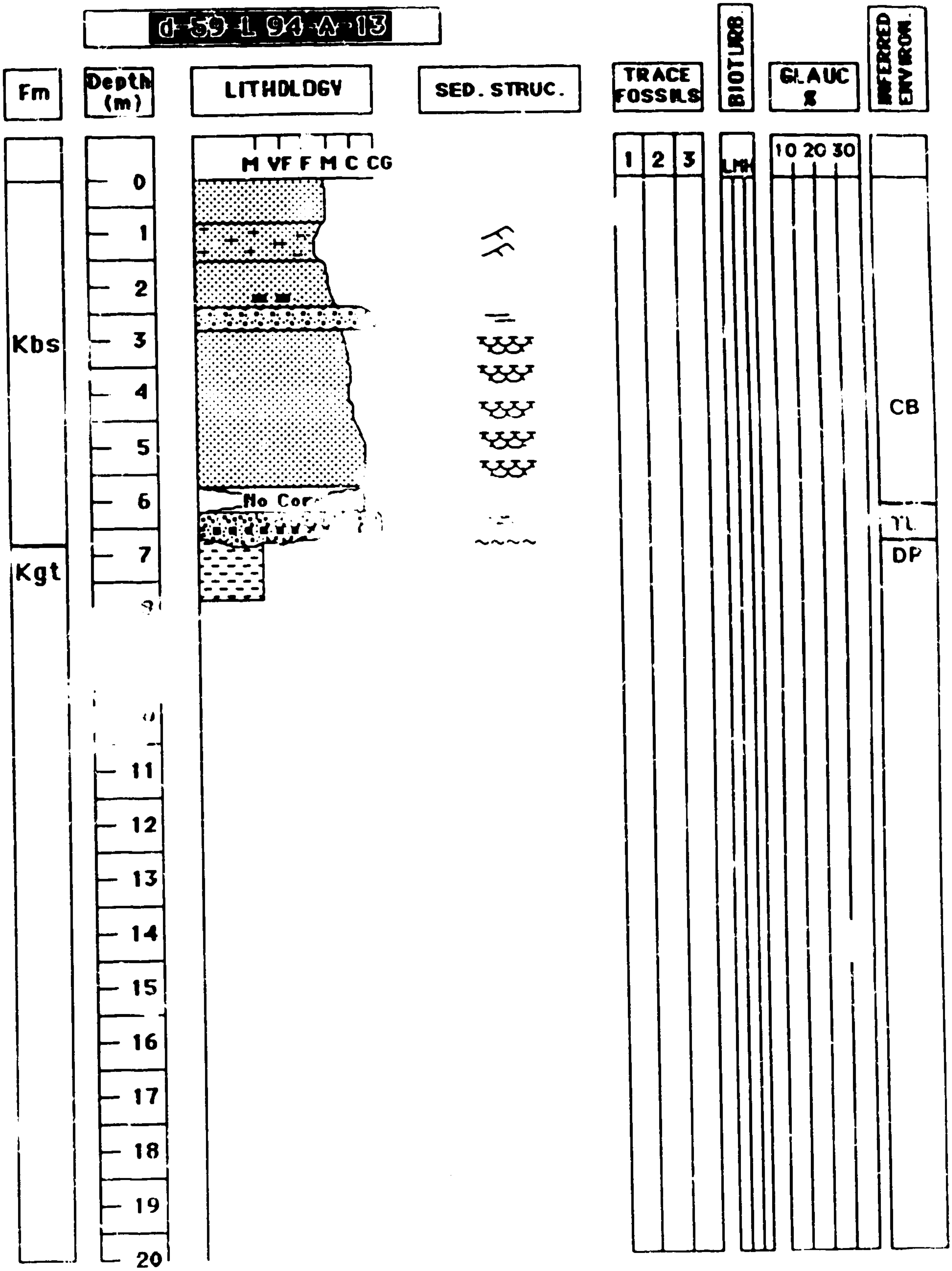
6
SF
IB
TL
End of Core.


d-59-L 94-A-13
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
Kbs
Kgt
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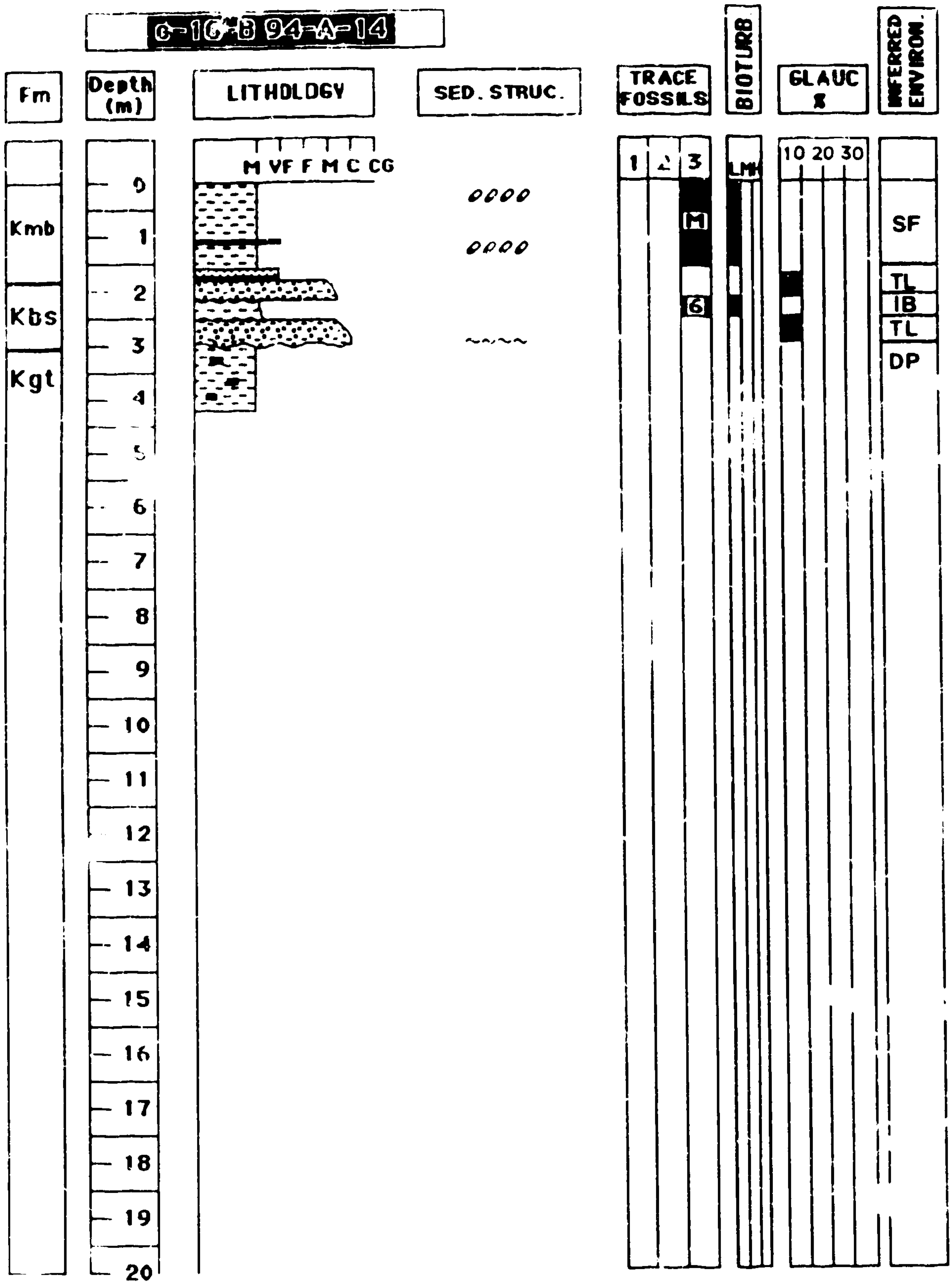
No Core
CB
TL
DP


C-16-B 94-A-14
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
Kmb
Kbs
Kgt
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M
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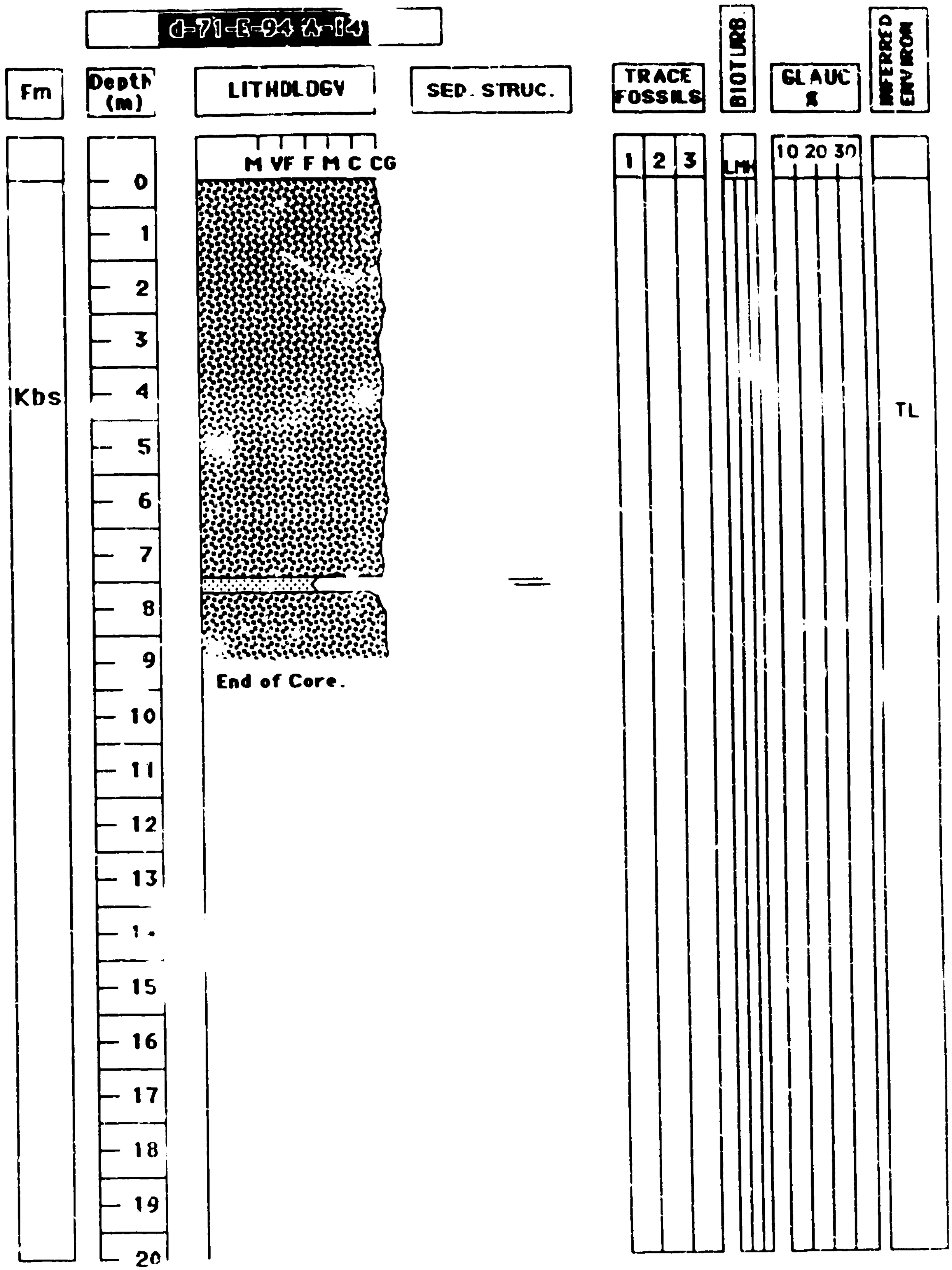
SF
TL
IB
TL
DP


d-71-E-94 A-14
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON
M VF F M C CG
1 2 3
L M H
10 20 30
Kbs
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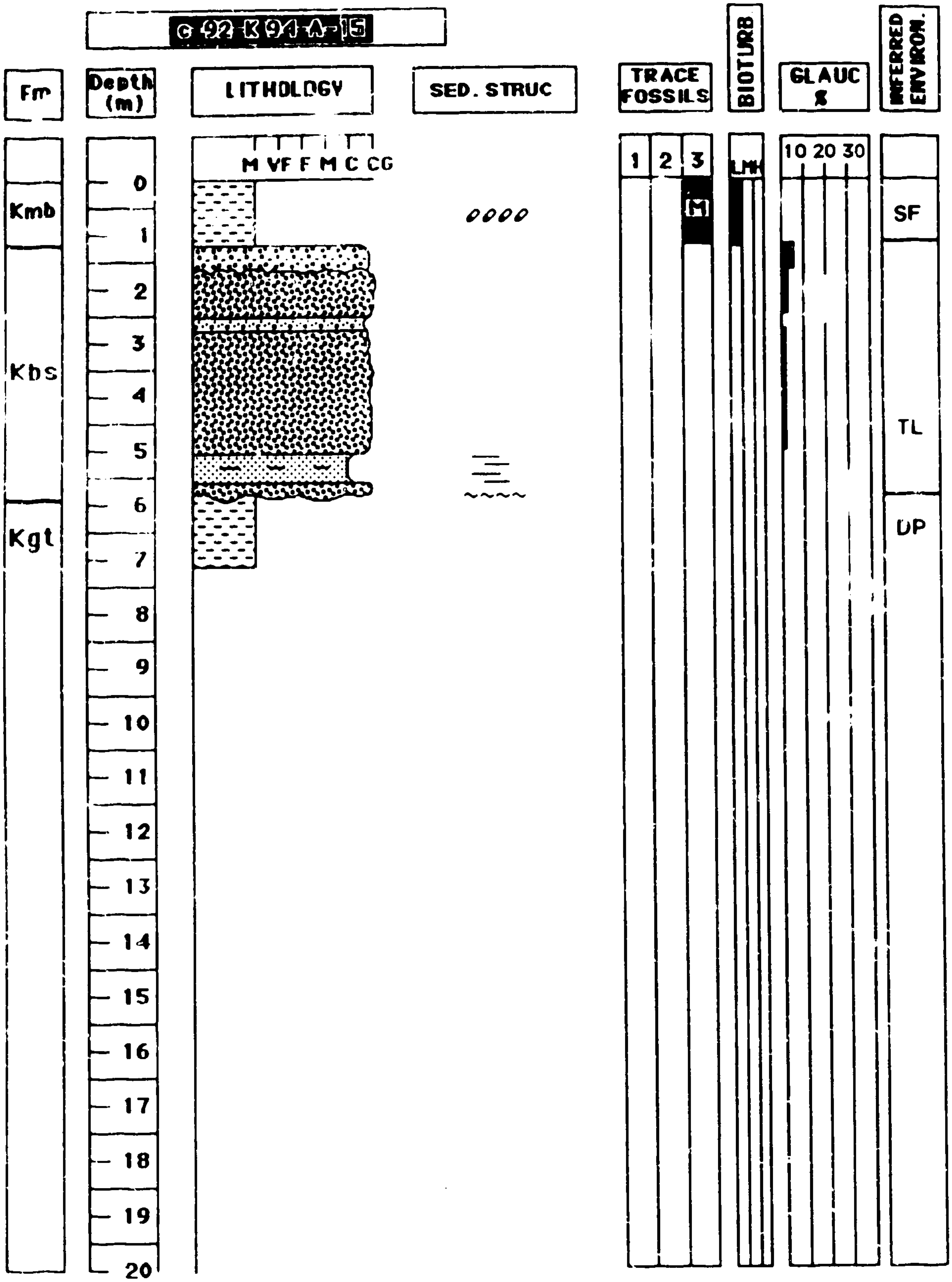
End of Core.
TL


G 92-K 94-A-15
Fm
Depth (m)
LITHOLOGY
SED. STRUC
TRACE FOSSILS
BIOTURB
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
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L M H
10 20 30
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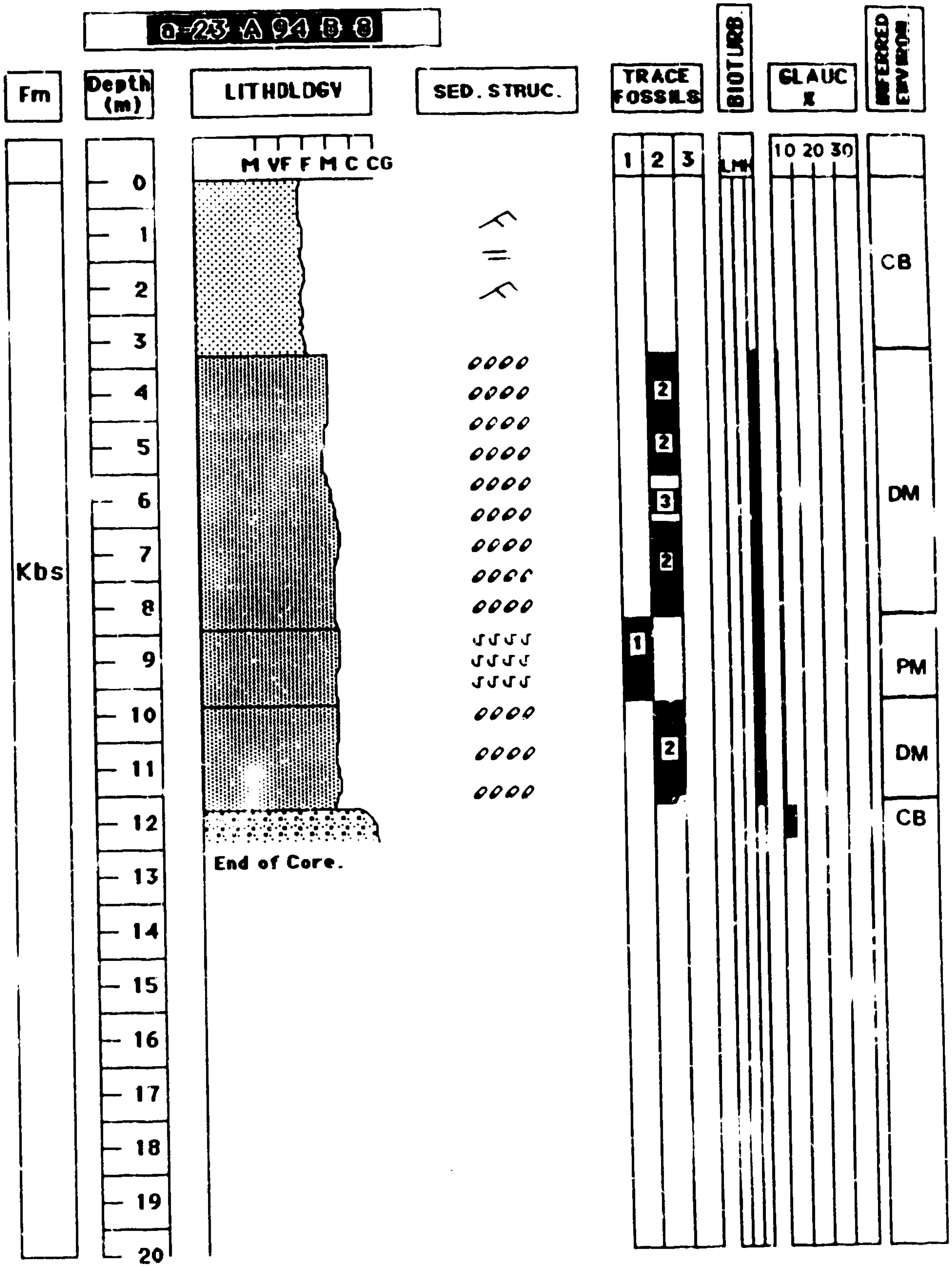
Kmb
Kbs
Kgt
SF
TL
DP



D-23 A 94 B 8
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
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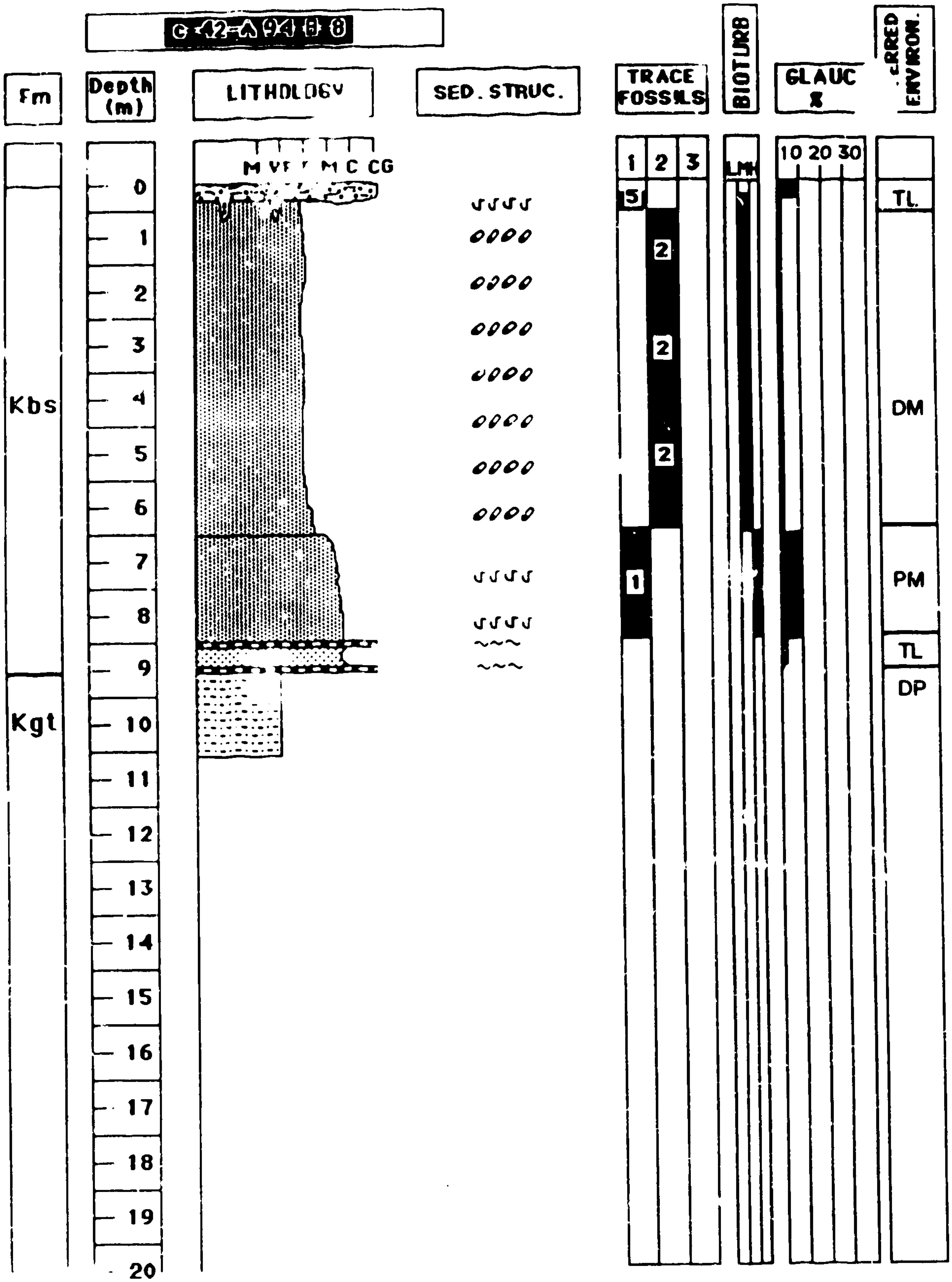
Kbs
End of Core.
CB
DM
PM
DM
CB


Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB
GLAUC %
ENVIRON.
M VF F M C CG
1 2 3
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Kbs
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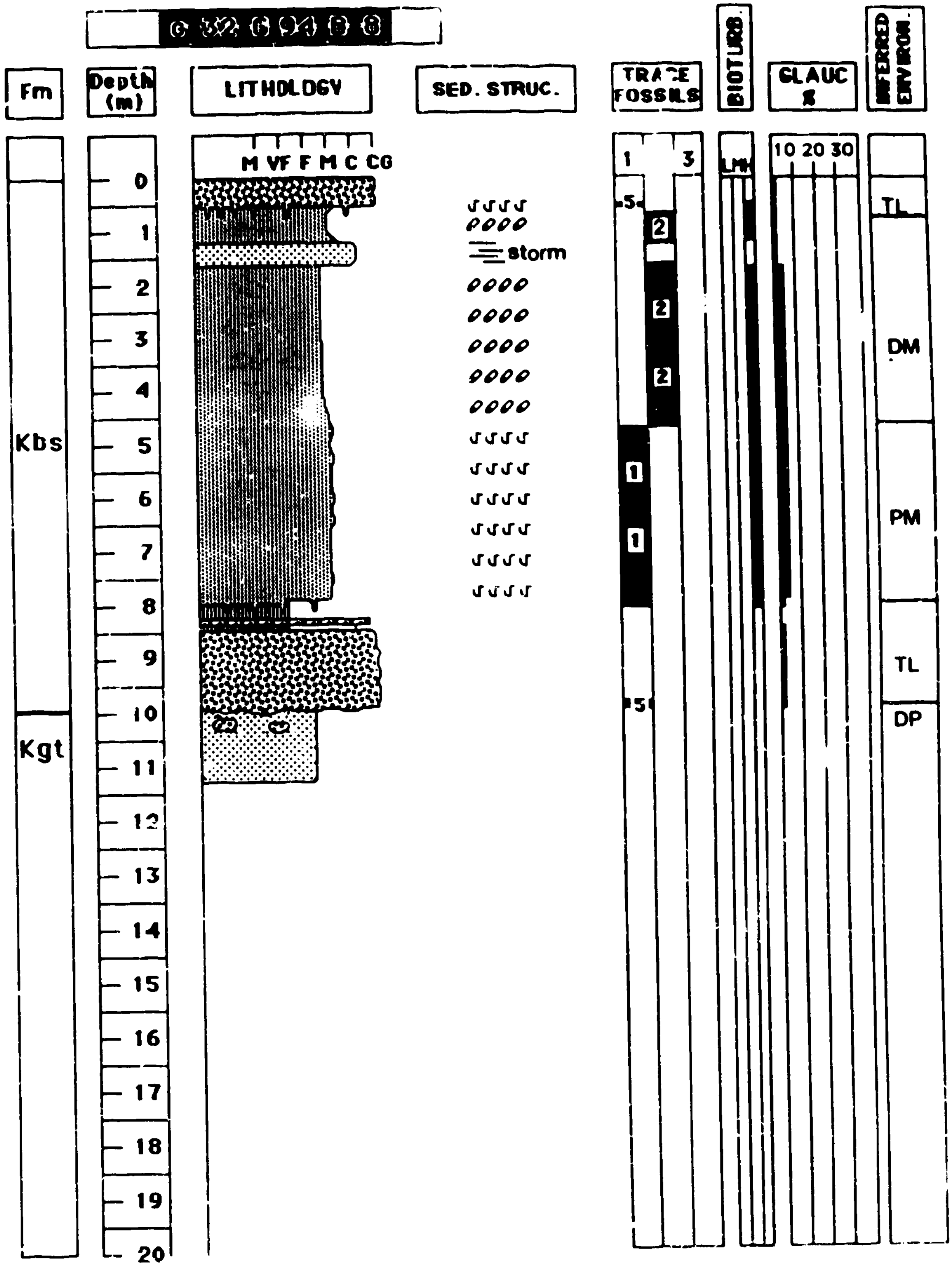

Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
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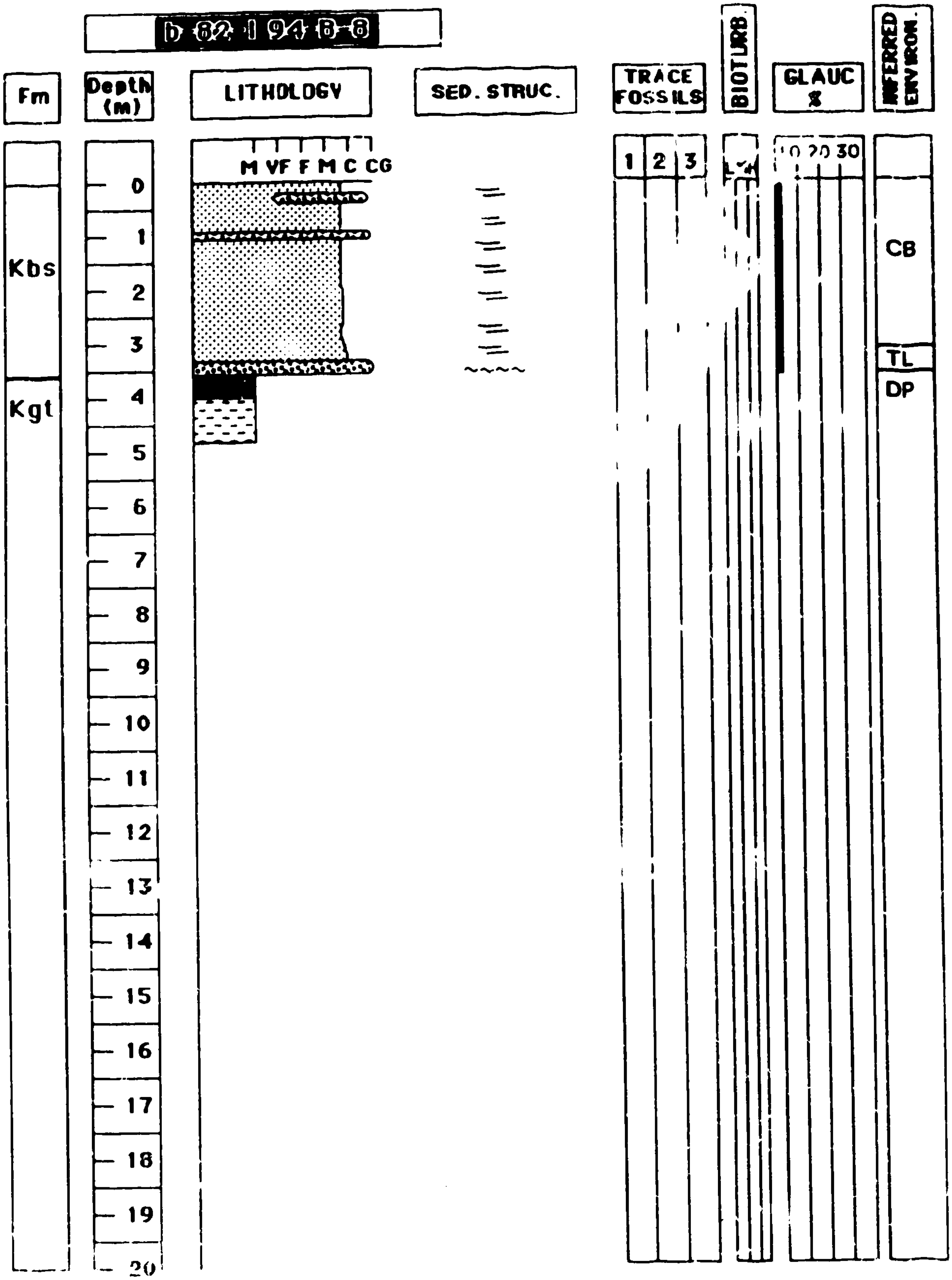
LMH
10 20 30
TL
DM
PM
TL
DP



b 82 I 94 B 8
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
10 20 30
Kbs
Kgt
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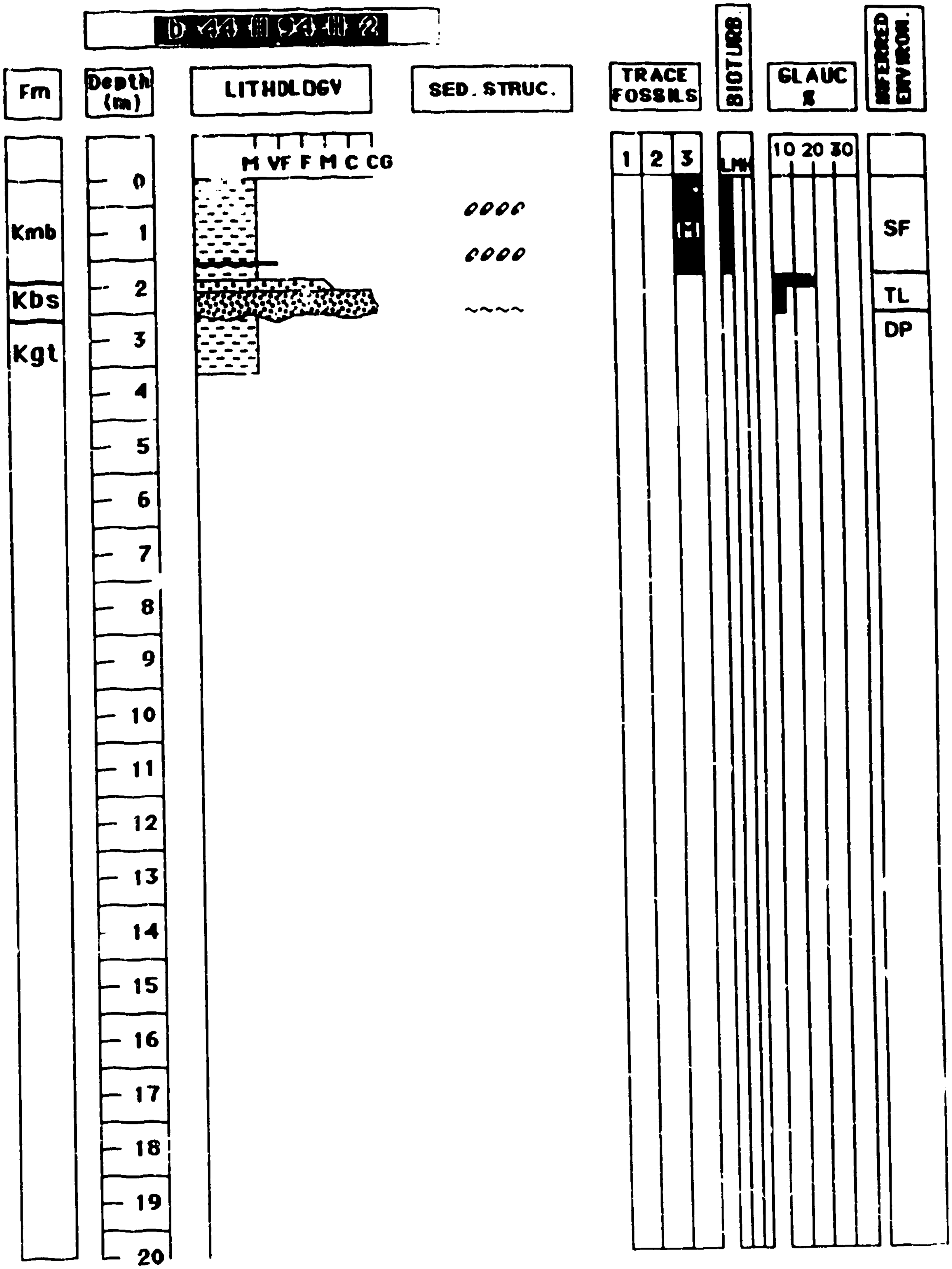
CB
TL
DP


D 44 H 94 H 2
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
Kmb
Kbs
Kgt
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SF
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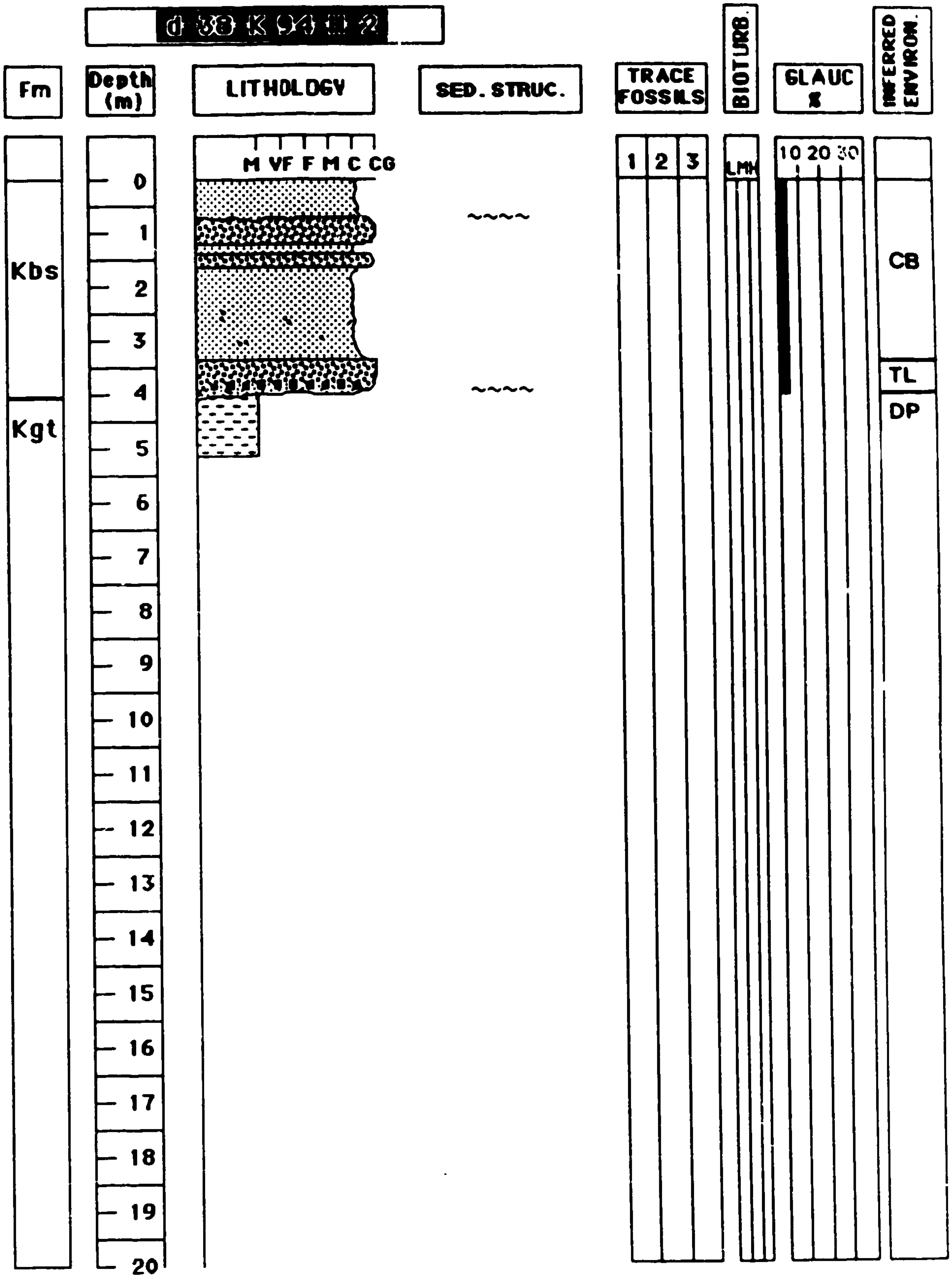
DP


d 38 K 94 III 2
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
Kbs
Kgt
0
1
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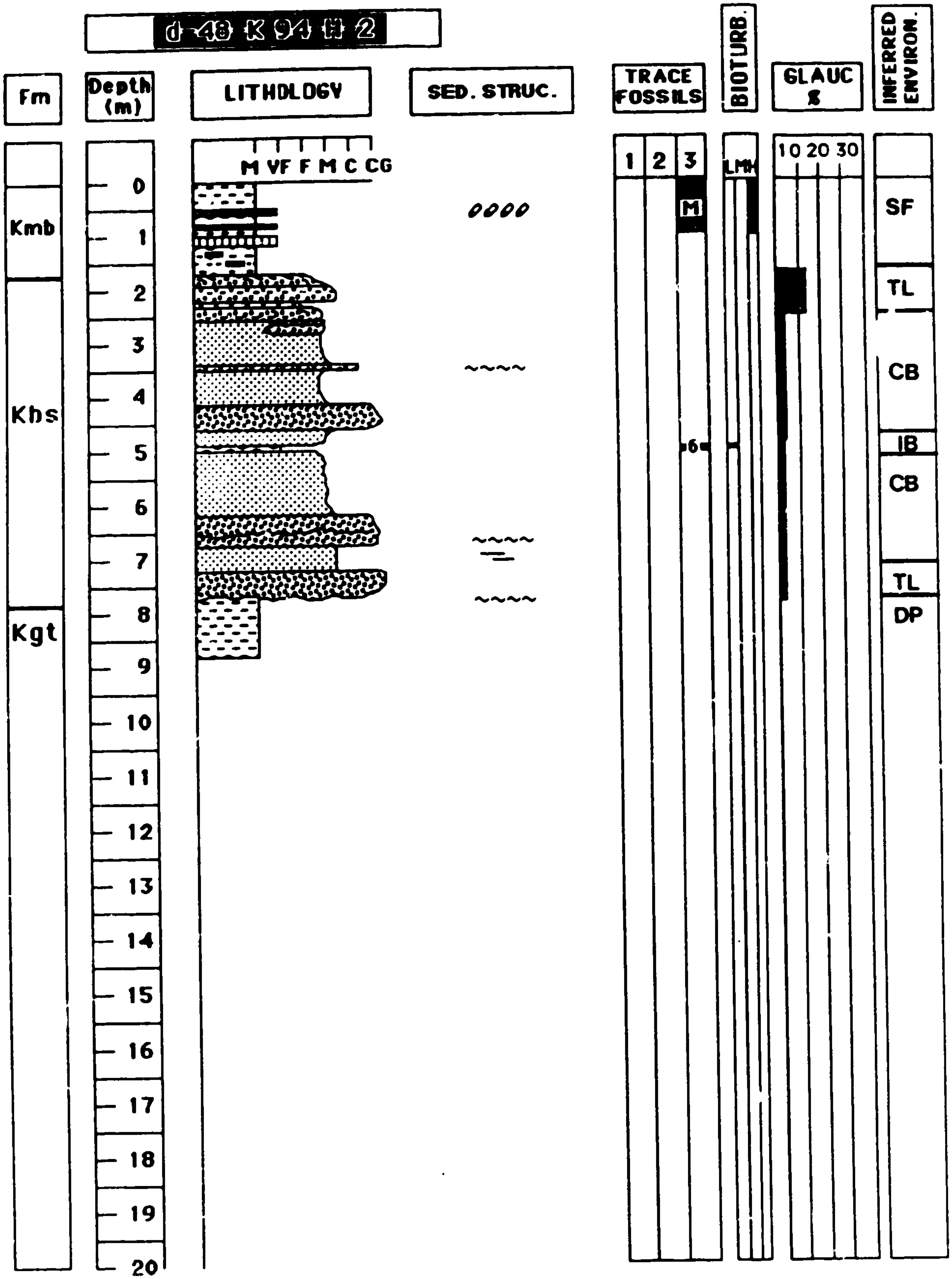
CB
TL
DP


d-48 K 94 H 2
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
Kmb
Khs
Kgt
0
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19
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SF
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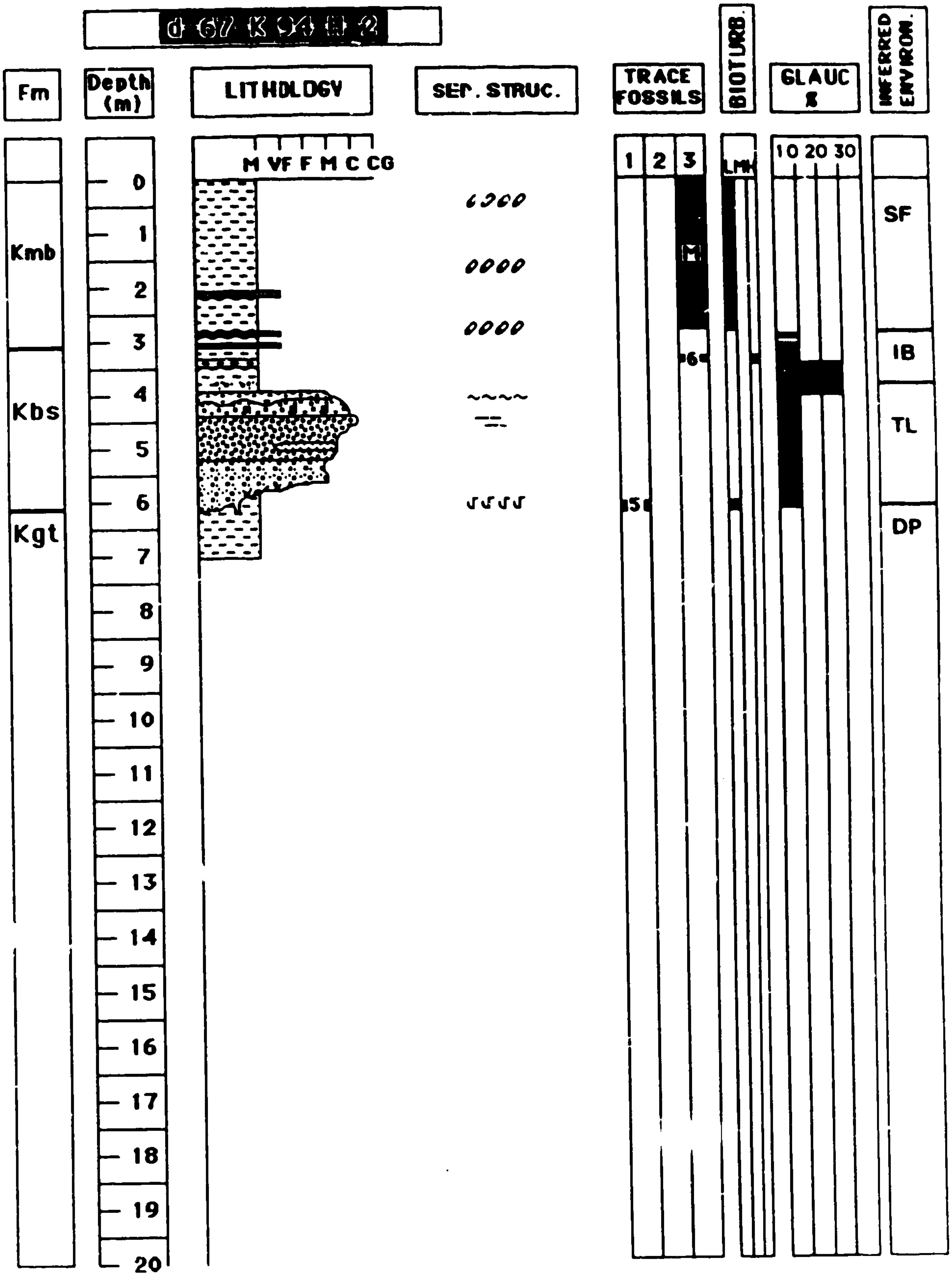
CB
IB
CB
TL
DP



d-67 K 94 H 2
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
Kmb
Kbs
Kgt
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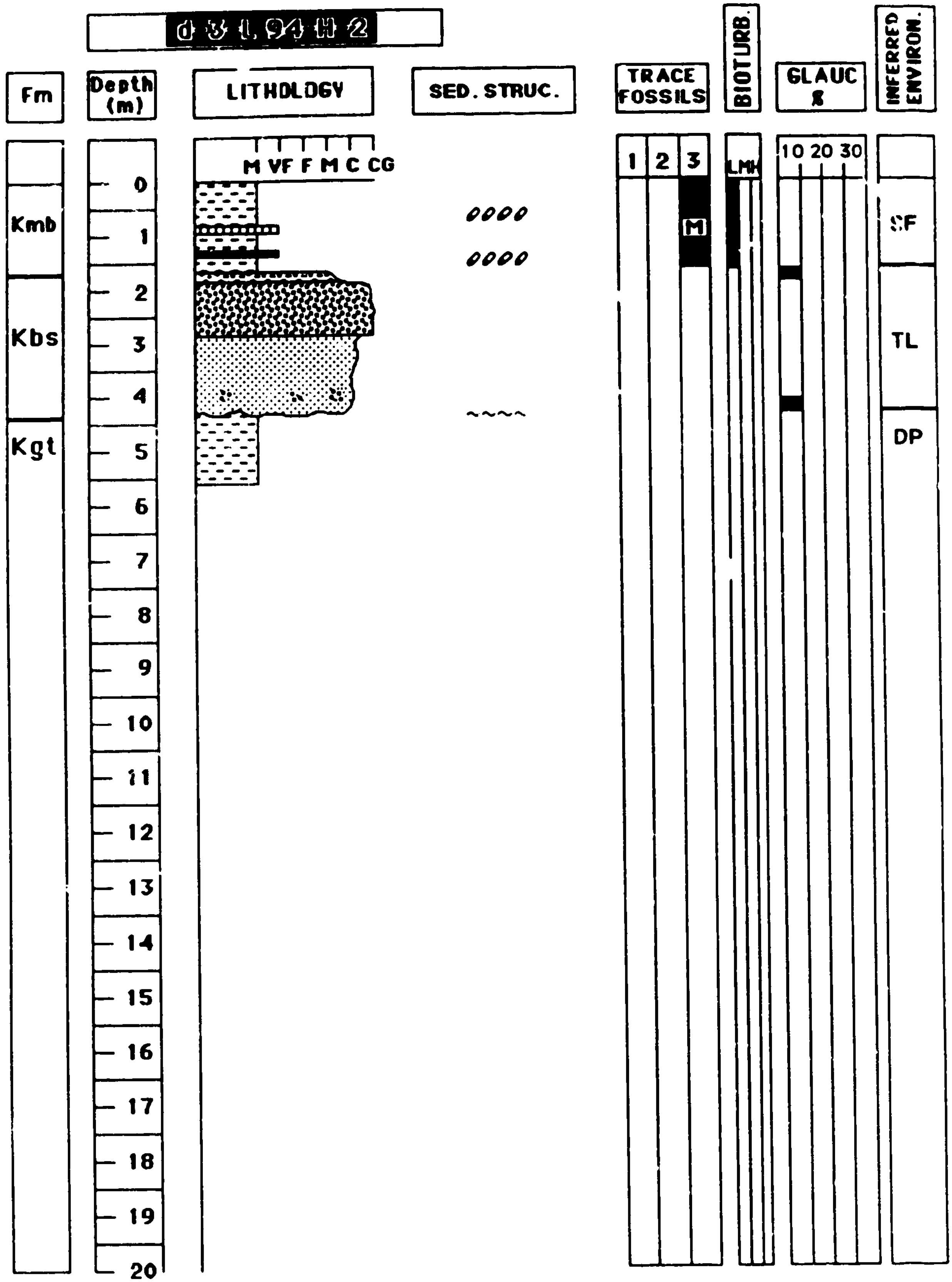
SF
IB
TL
DP


d 3 L 94 H 2
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
Kmb
Kbs
Kgt
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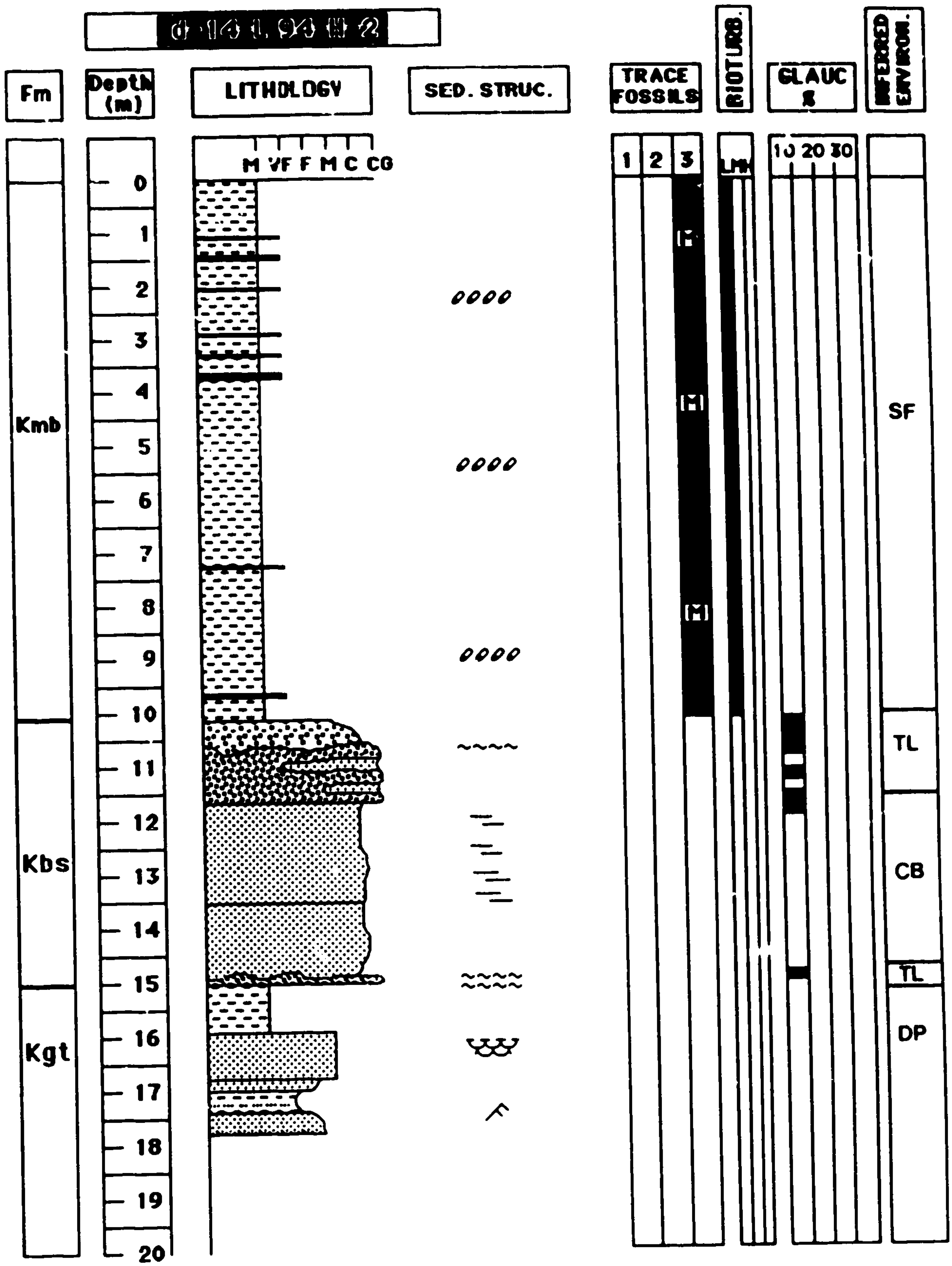
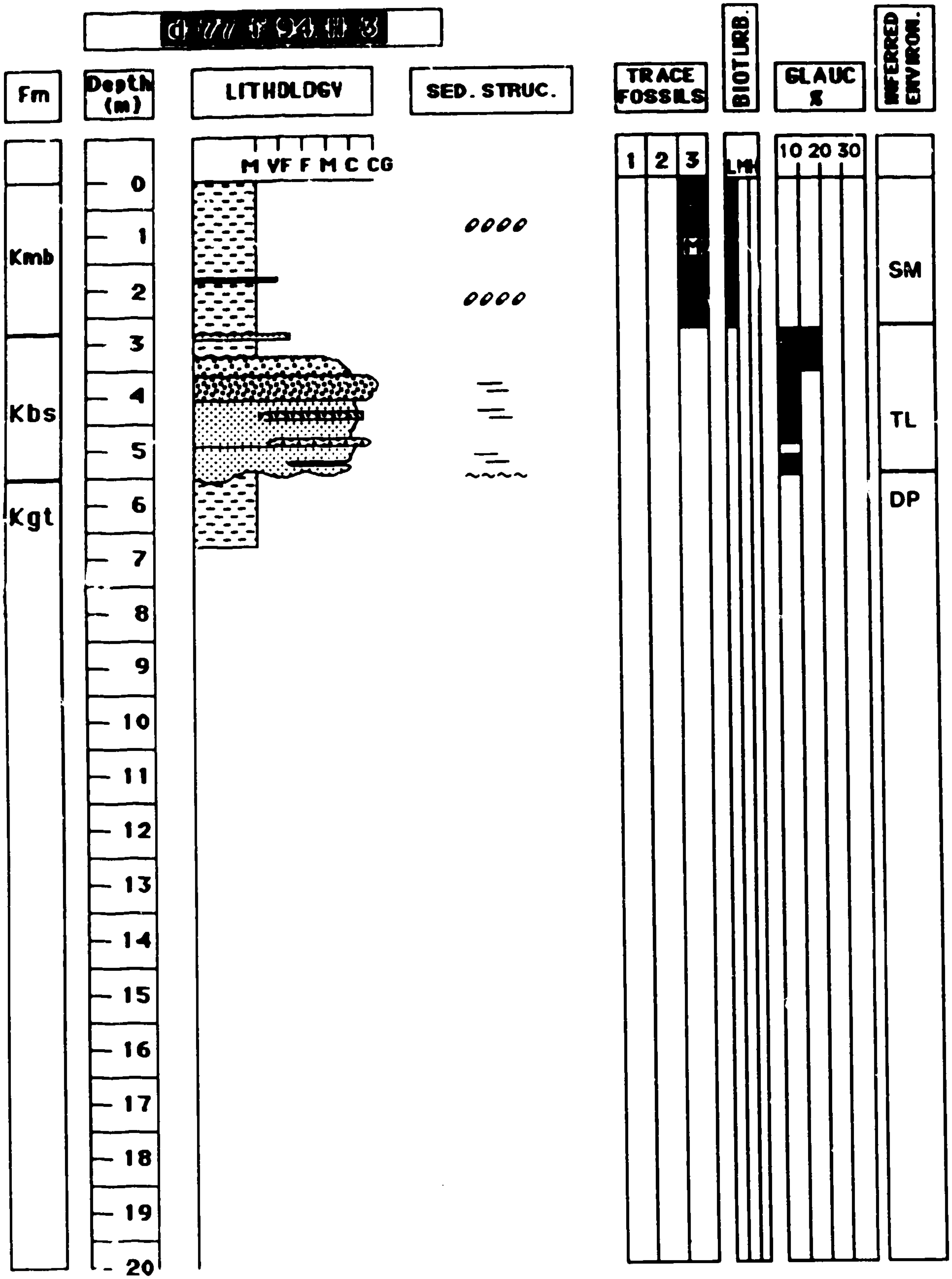
16
17
18
19
20
M
SF
TL
DP


d-14-L, 94 H 2
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
Kmb
Kbs
Kgt
M
M
M
SF
TL
CB
TL
DP



d-77-F/94-H-3
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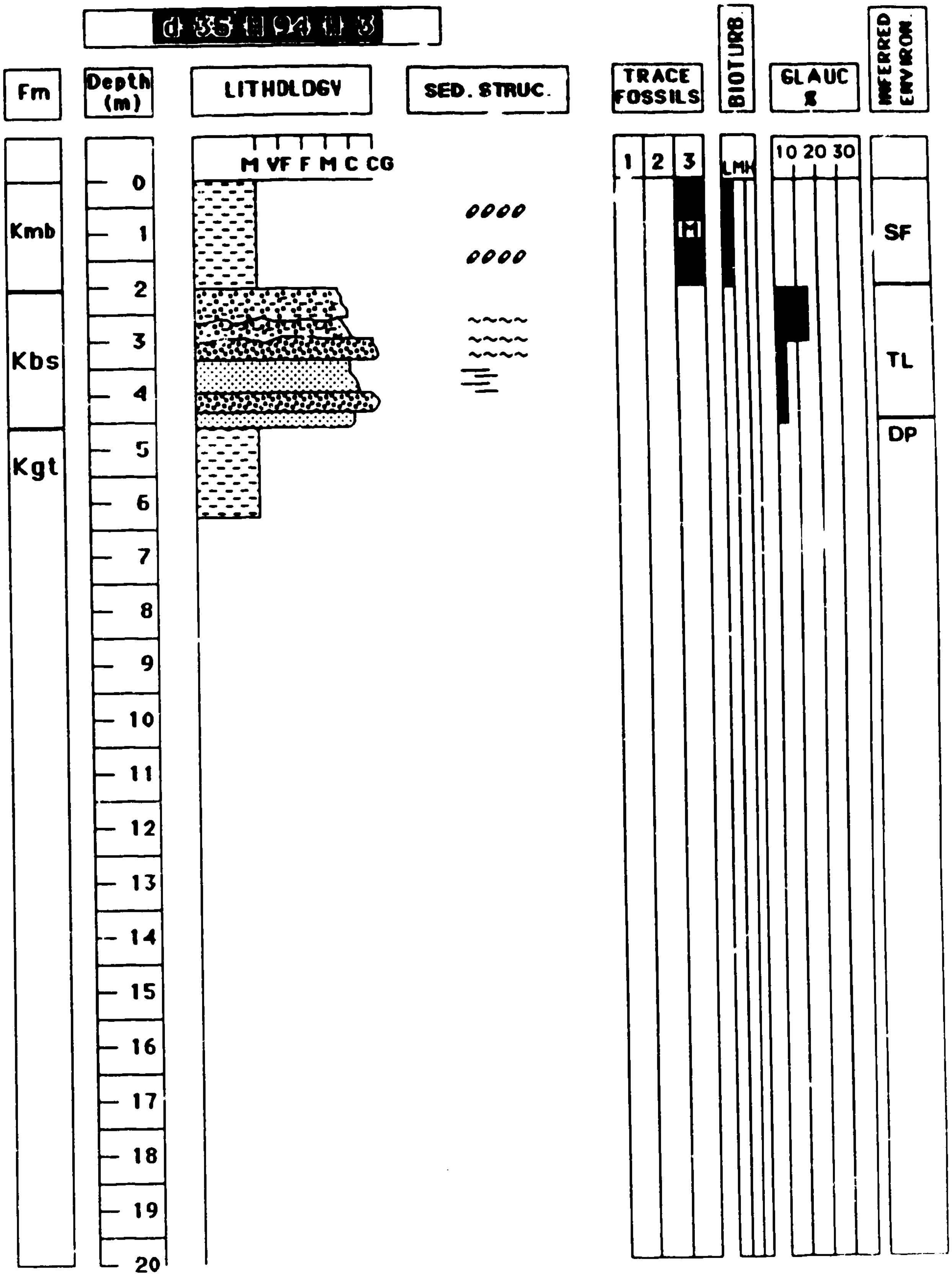
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
Kmb
Kbs
Kgt
SM
TL
DP


Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
Kmb
Kbs
Kgt
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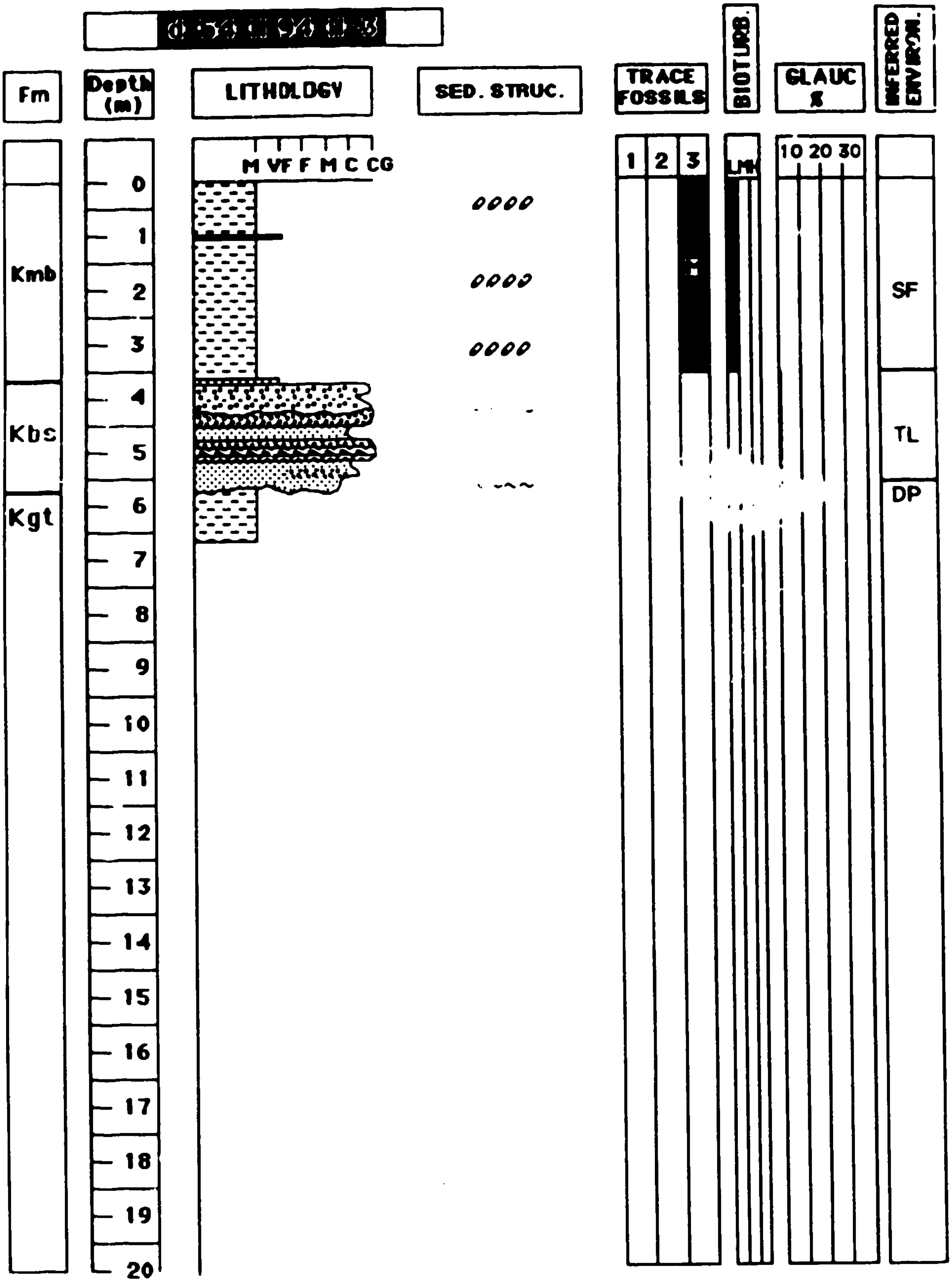
SF
TL
DP


Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
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1 2 3
LMH
10 20 30
Kmb
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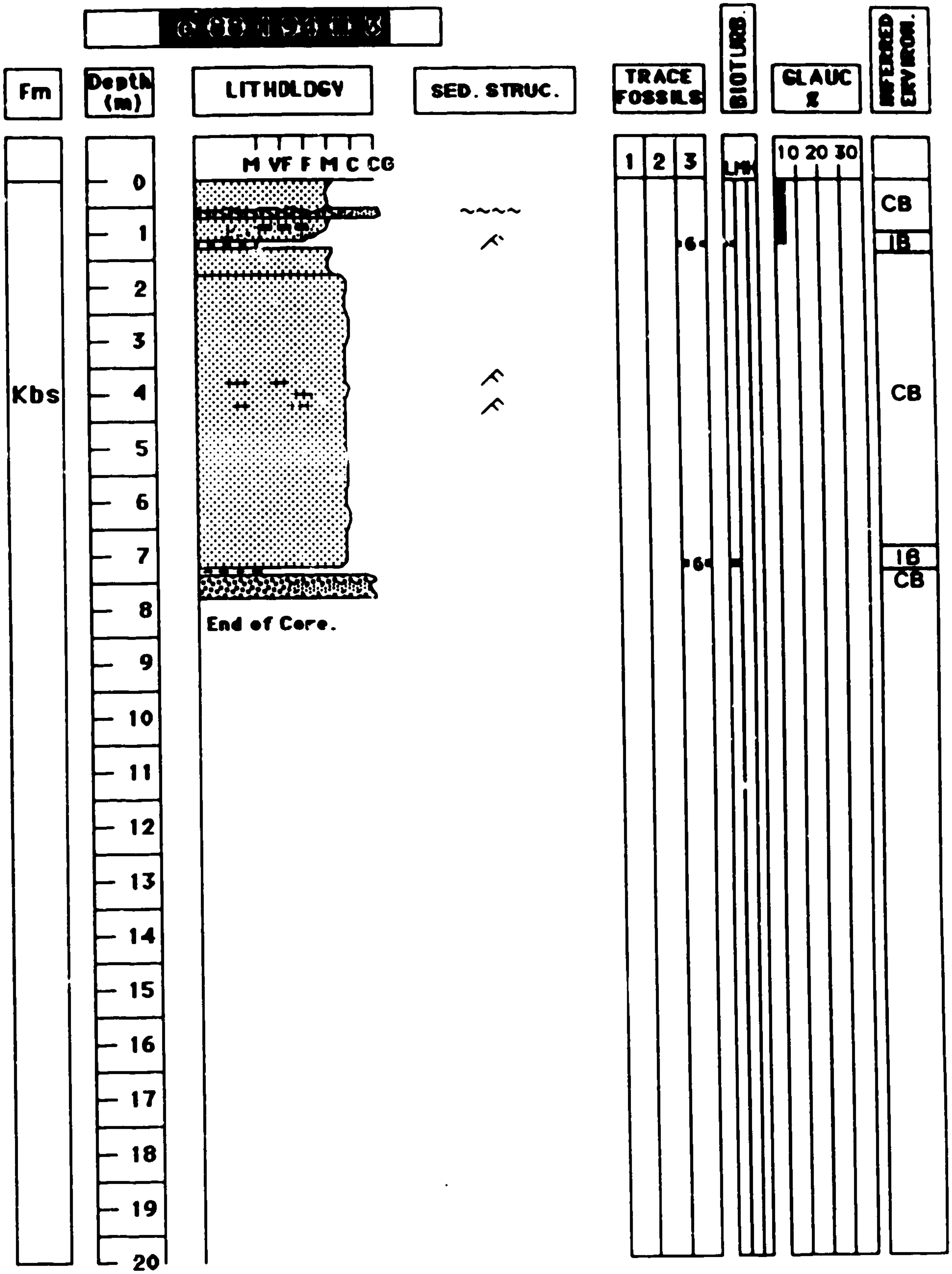
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CB
1 2 3
10 20 30
Kbs
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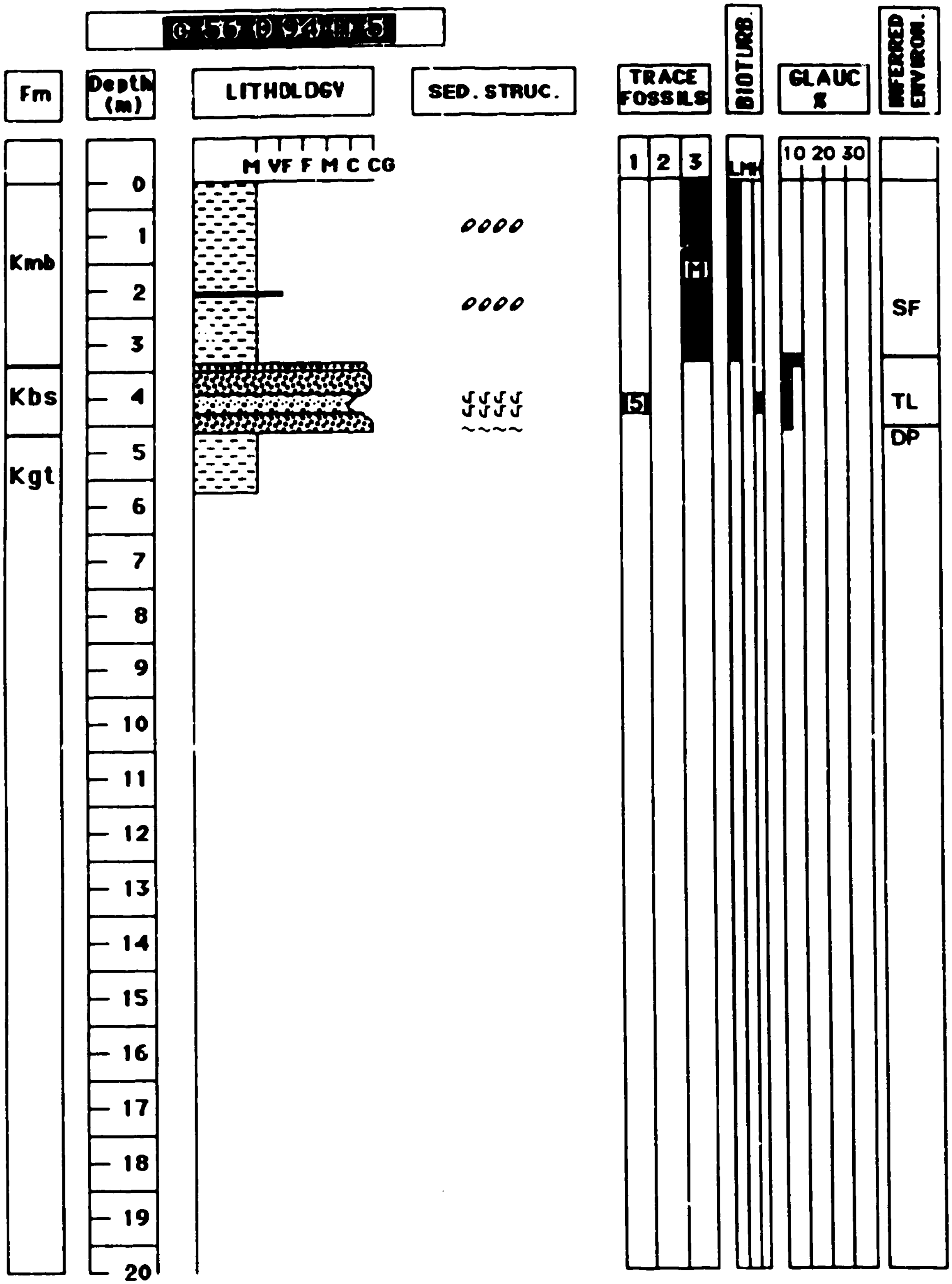
20
End of Core.
6
6
CB
IB
CB
IB
CB


Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
Kmb
Kbs
Kgt
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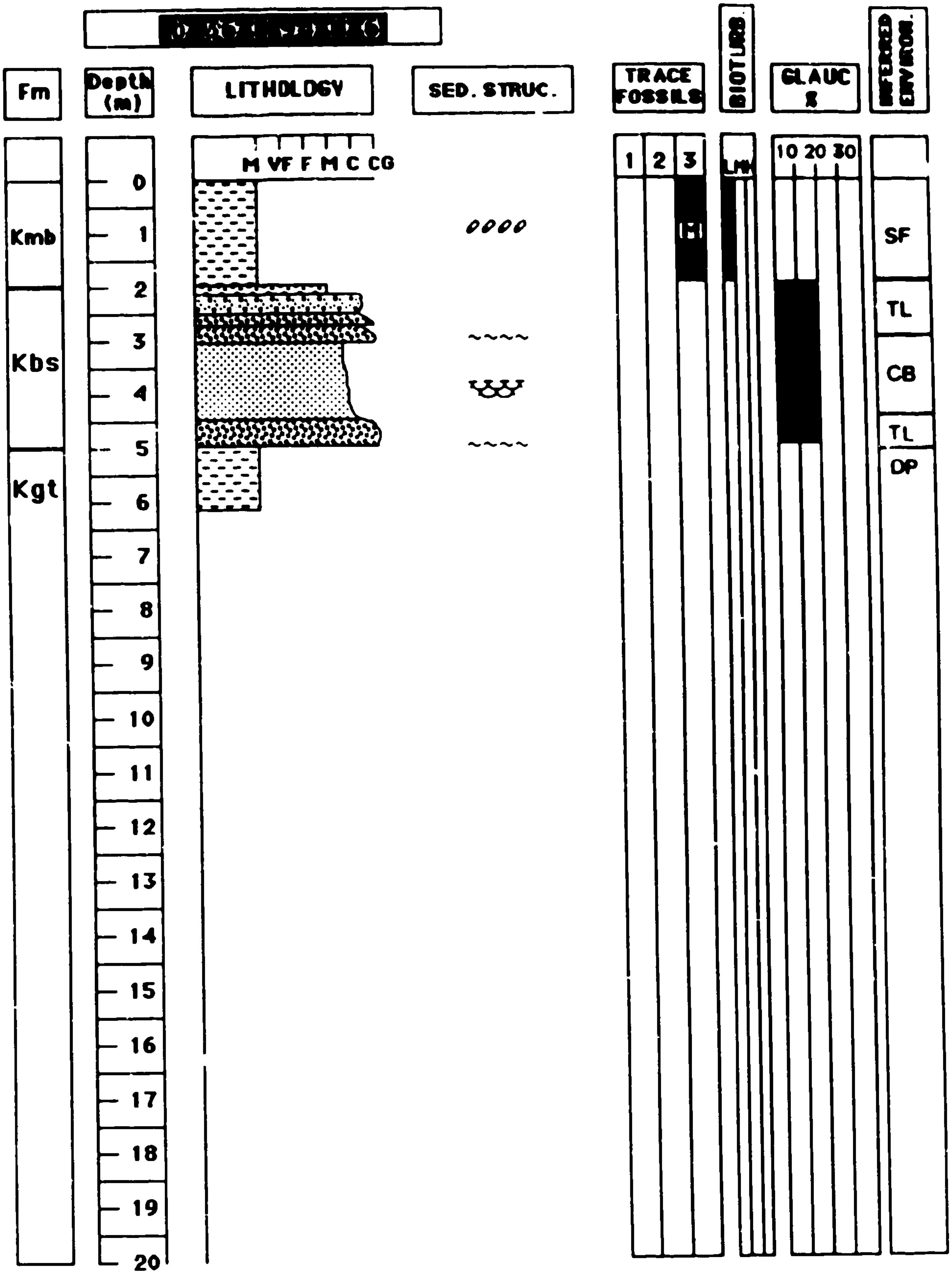
SF
TL
DP



Fm
Depth (m)
LITHOLOGY
M VF F M C CG
SED. STRUC.
TRACE FOSSILS
1 2 3
BIOTURB.
L M H
GLAUC %
10 20 30
INFERRED ENVIRON.
Kmb
Kbs
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SF
TL
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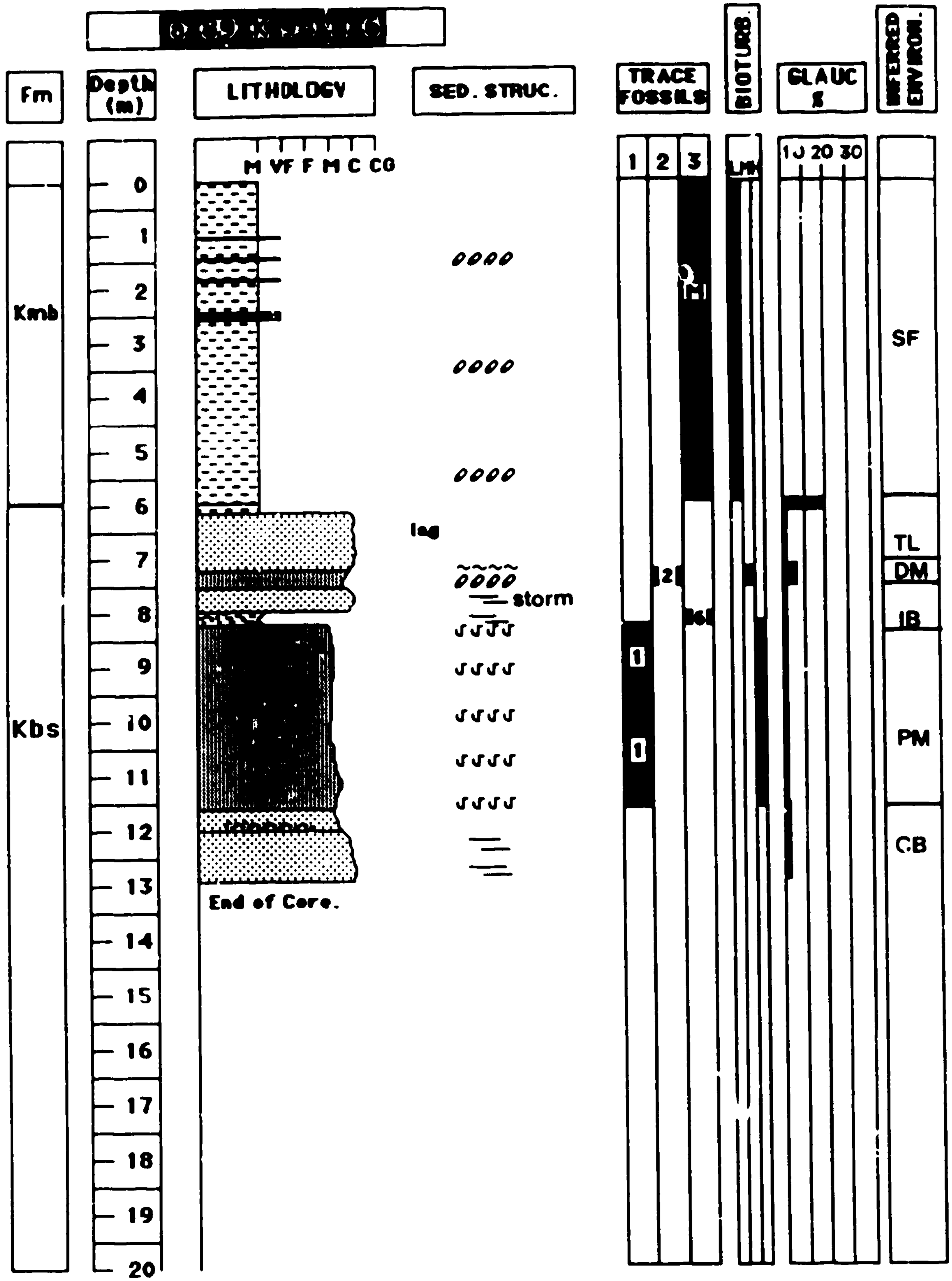
TL
DP


Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
10 20 30
Kmb
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lag
storm
End of Core.
SF
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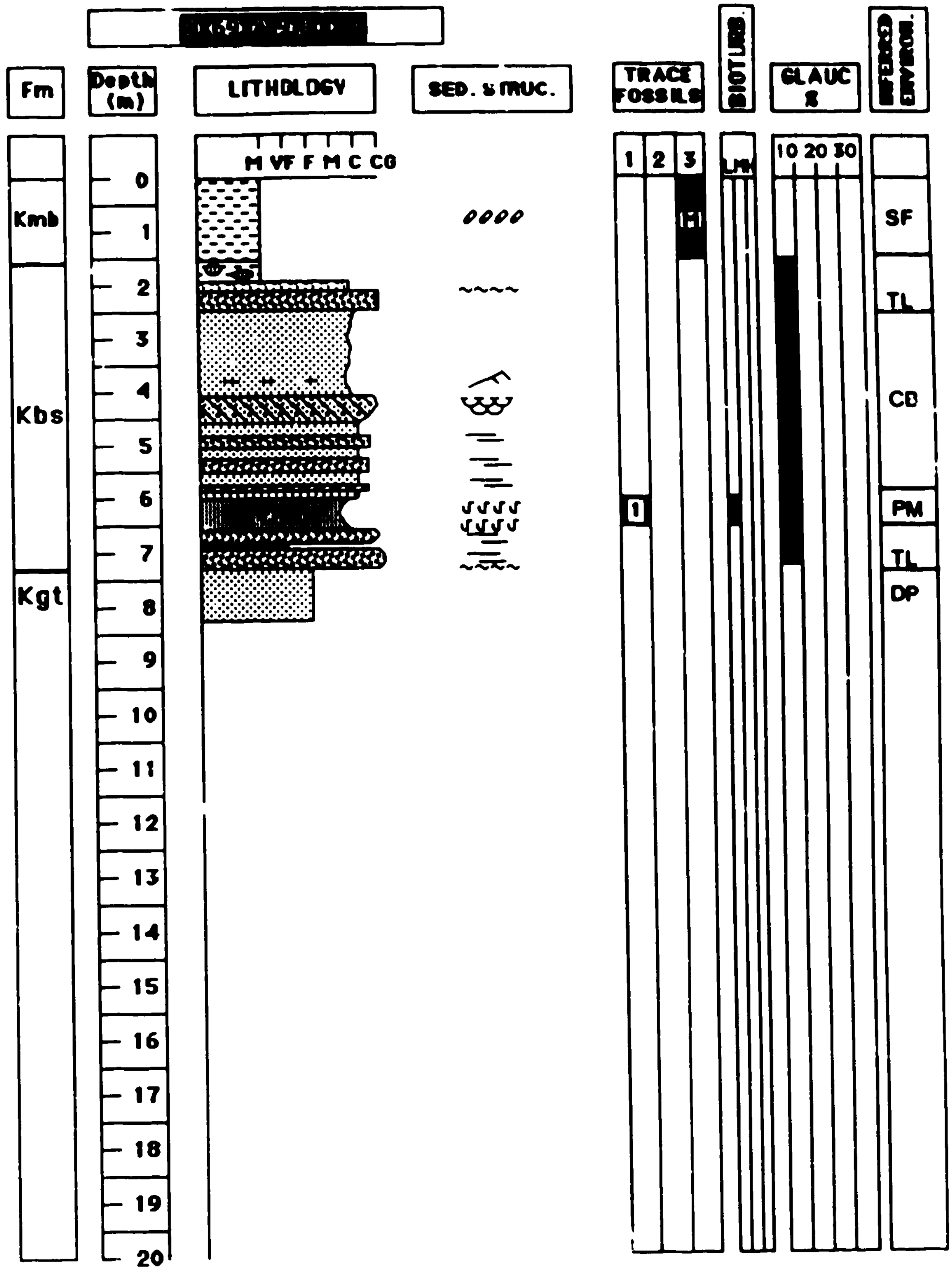
DM
IB
PM
CB


Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
Kmb
Kbs
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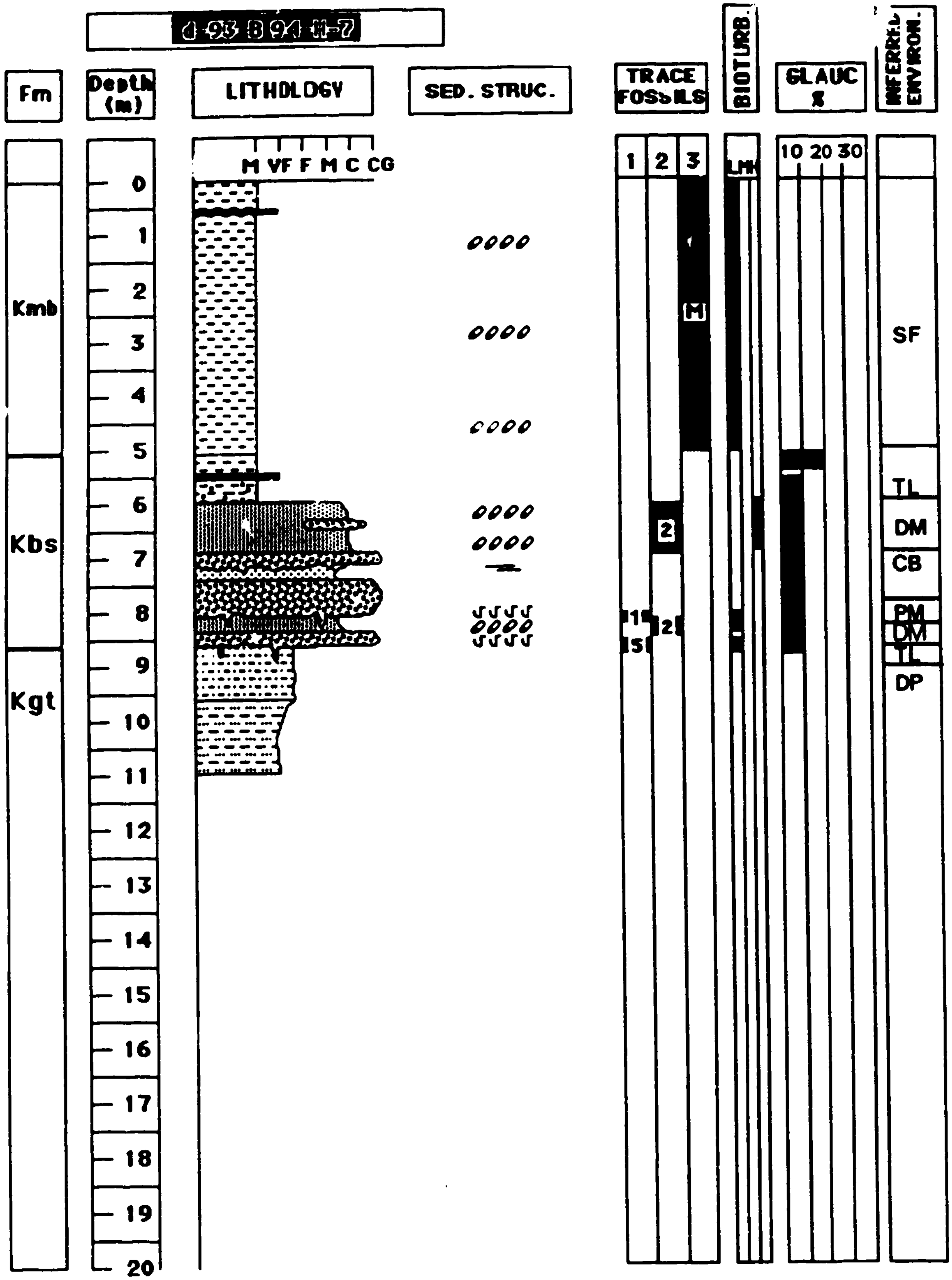
SF
TL
CB
PM
TL
DP



d-93-B 94-H-7
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
Kmb
Kbs
Kgt
M
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SF
TL
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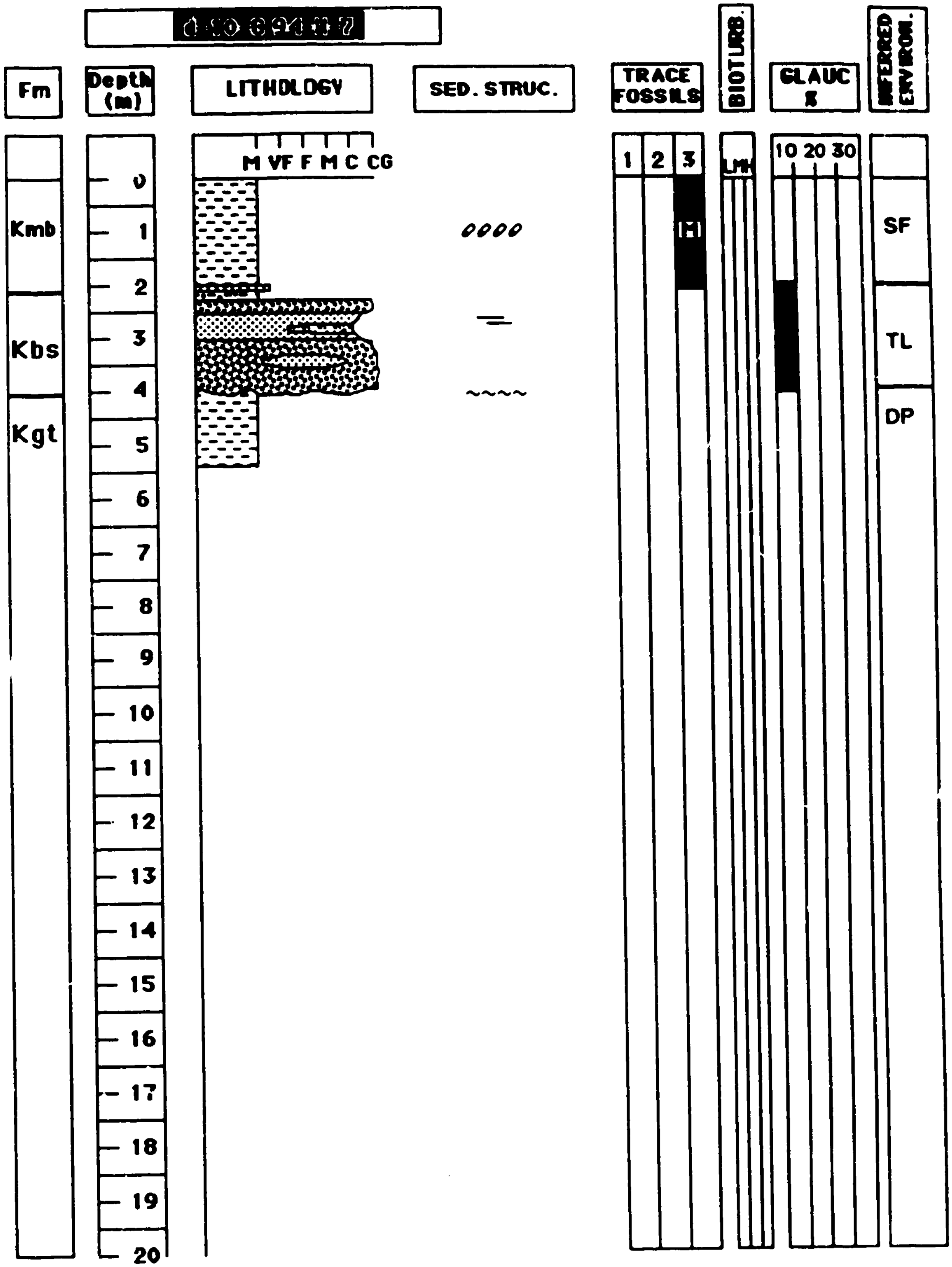
CB
PM
DM
TL
DP



Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
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10 20 30
Kmb
Kbs
Kgt
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SF
TL
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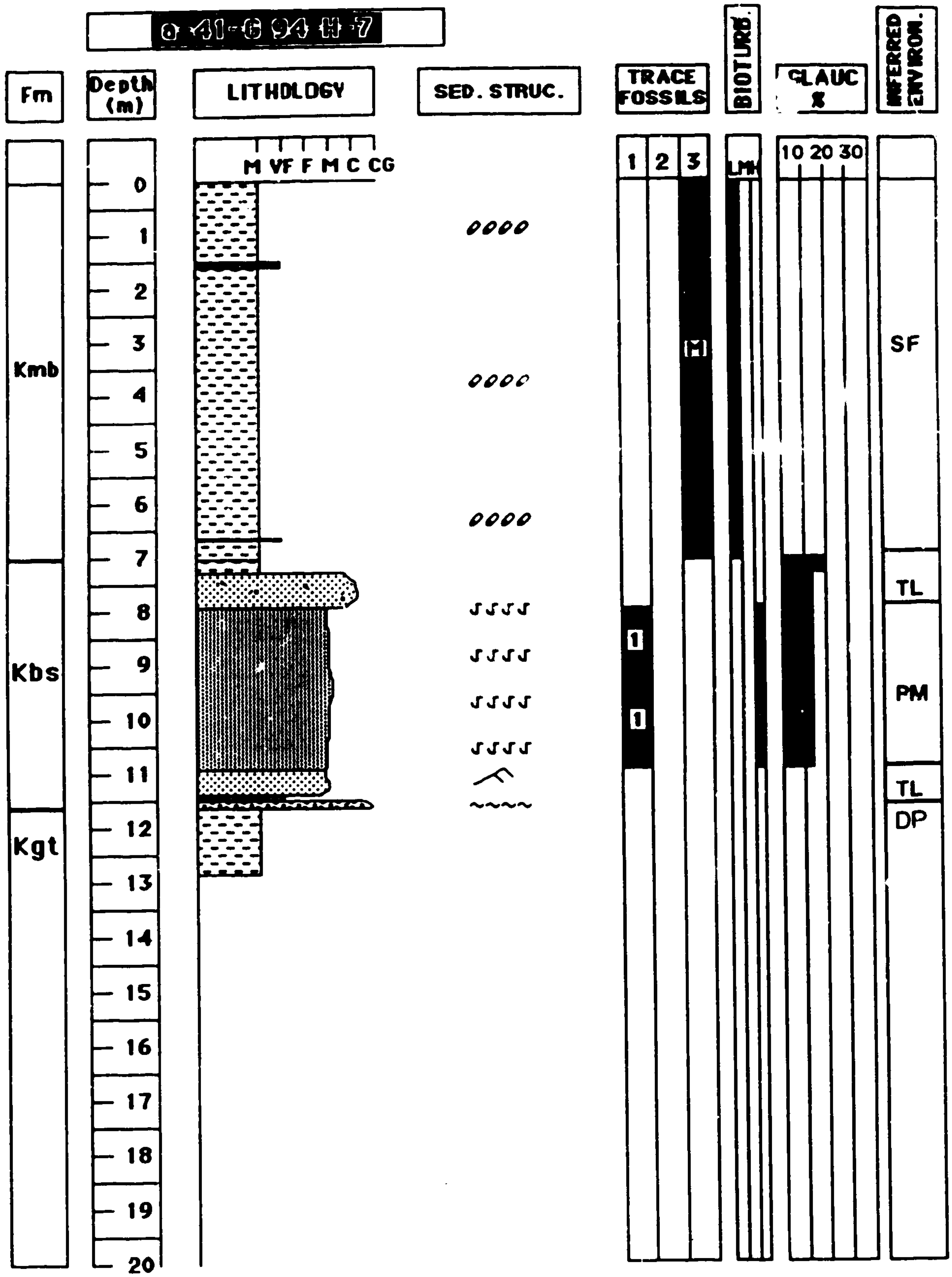
a-41-G 94 H 7
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
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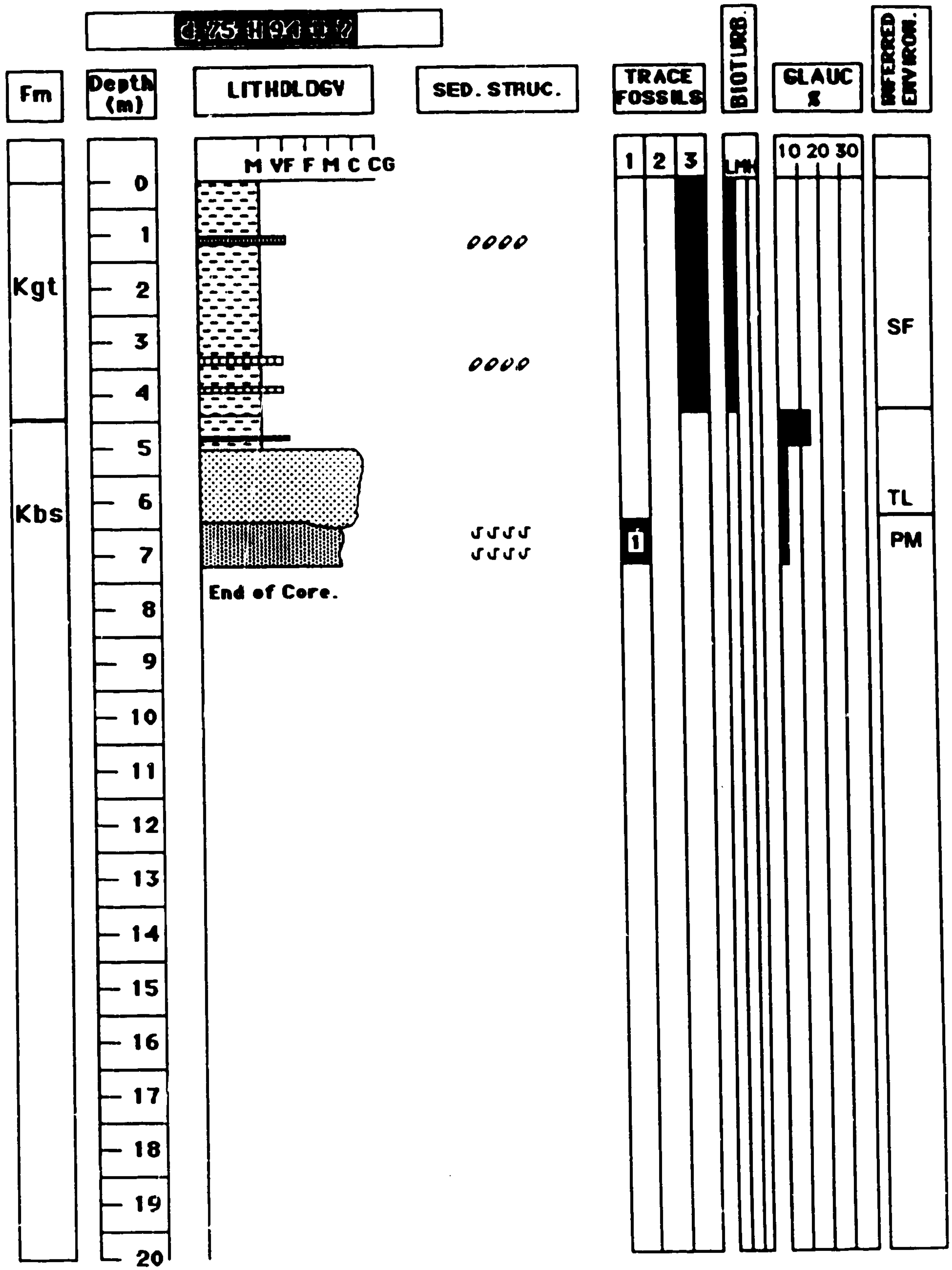
Kmb
Kbs
Kgt
SF
TL
PM
TL
DP



Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
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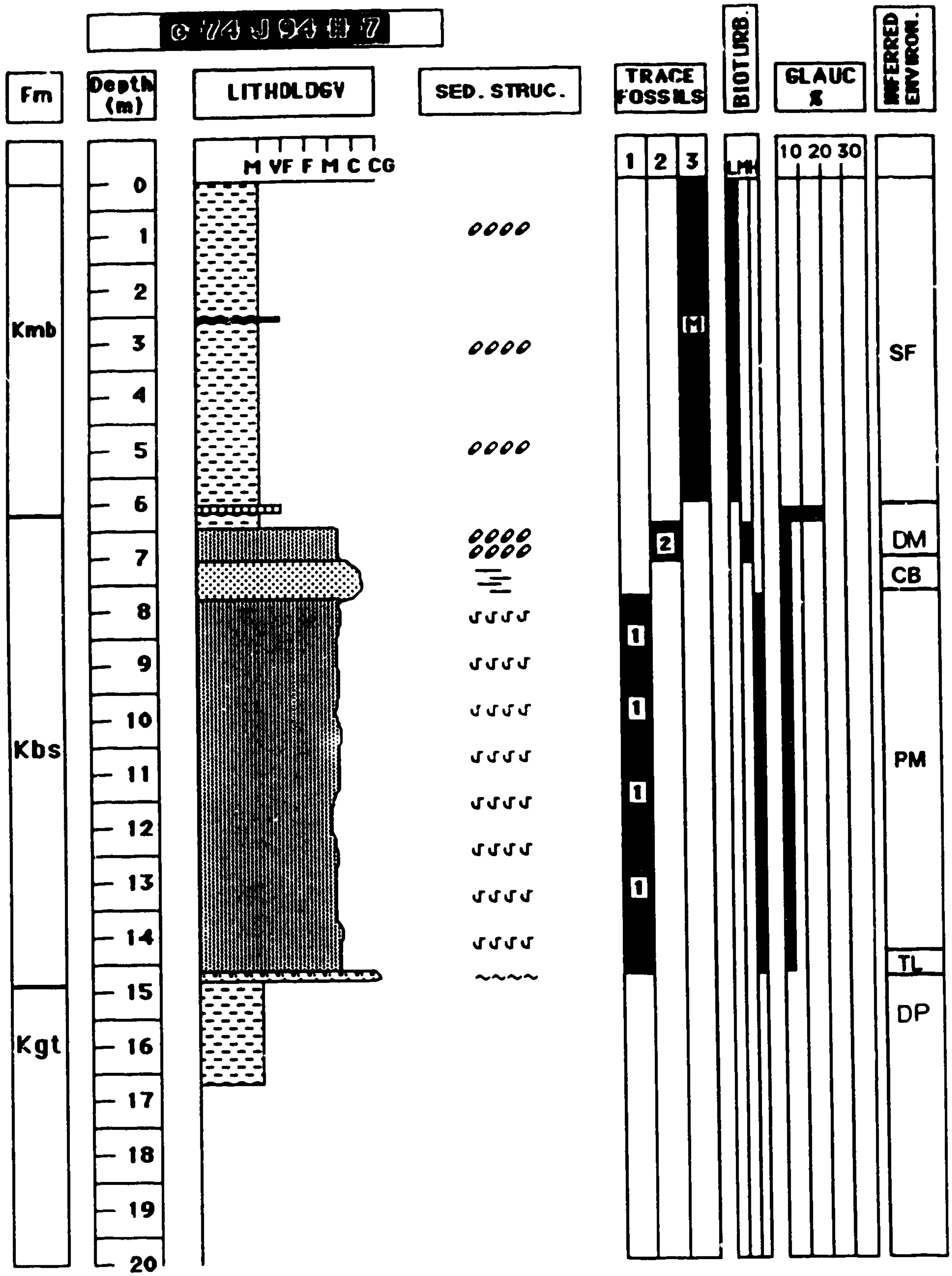
Kgt
Kbs
End of Core.
SF
TL
PM



C 74 J 94 H 7
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
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18
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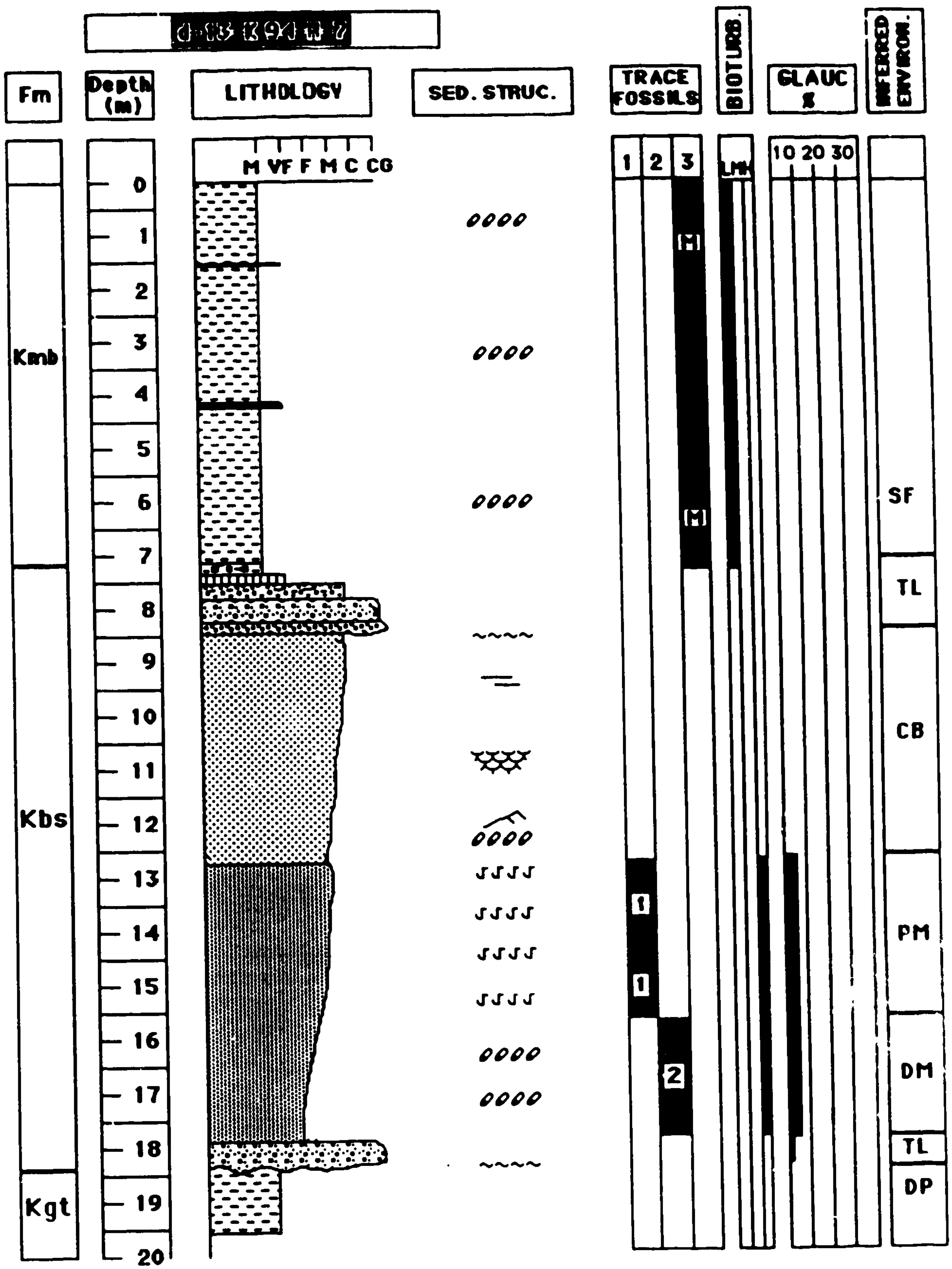
20
Kmb
Kbs
Kgt
1 2 3
M
2
1
L M H
10 20 30
SF
DM
CB
PM
TL
DP



Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
Kmb
Kbs
Kgt
M
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SF
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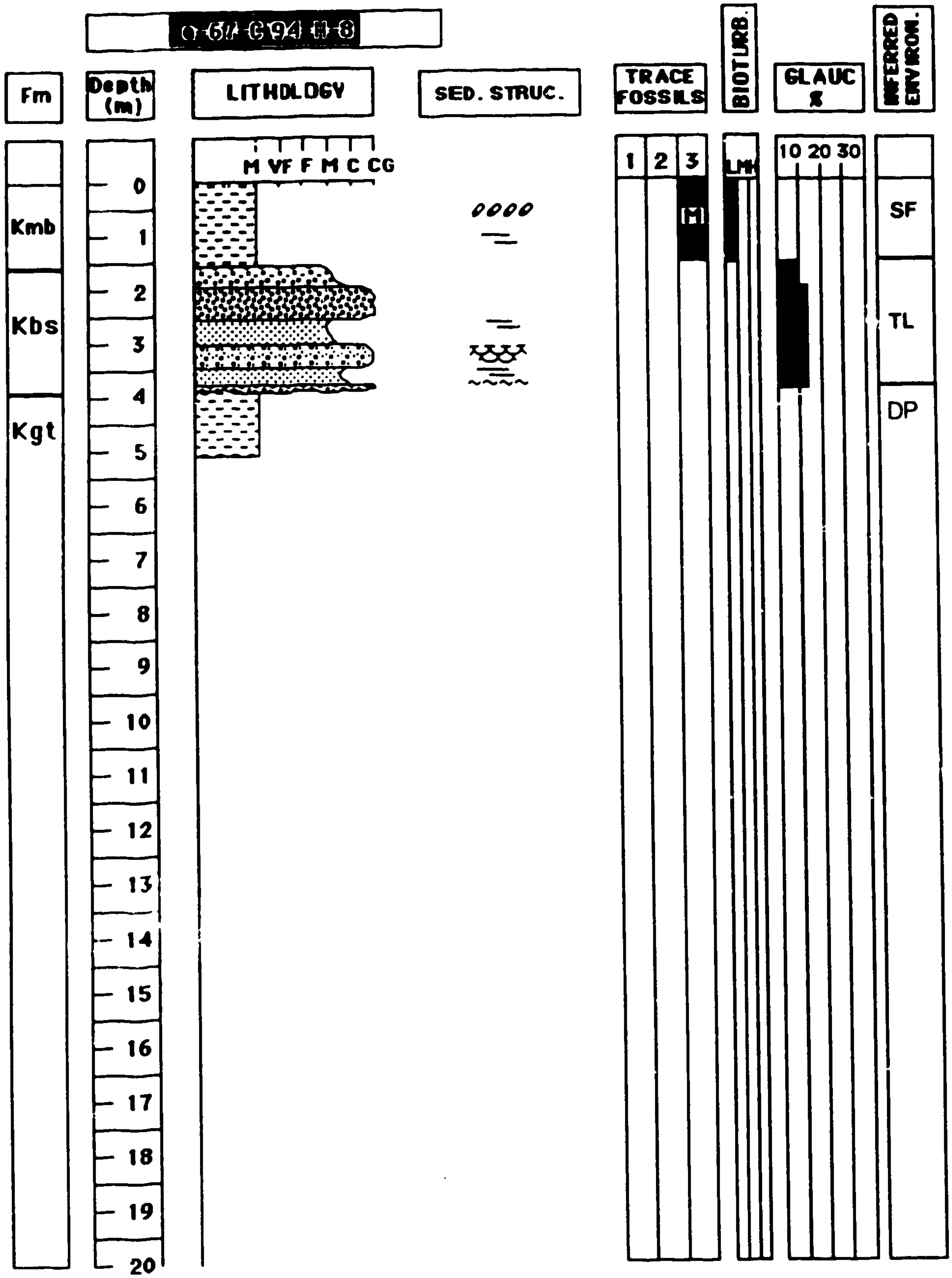
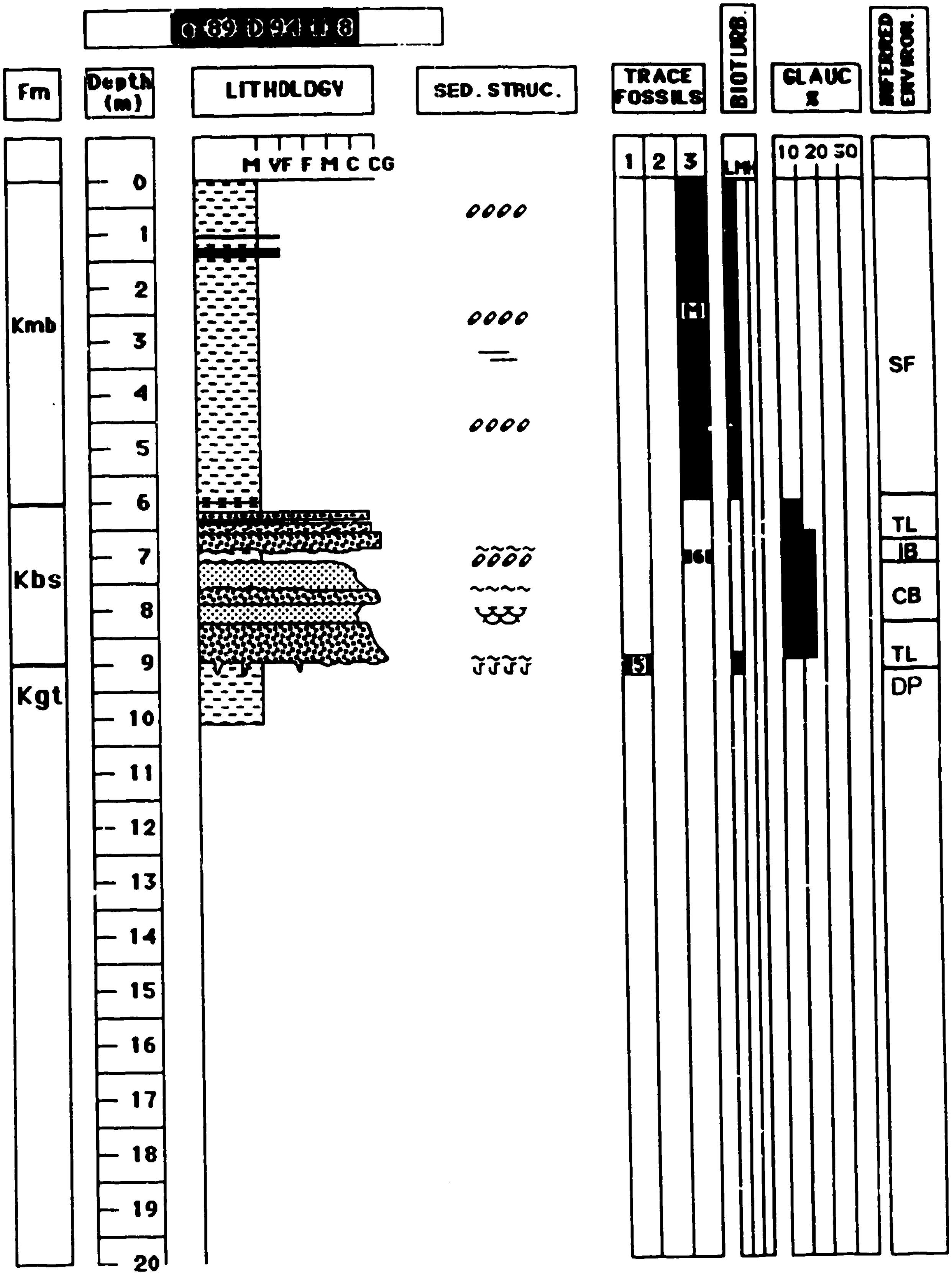
PM
DM
TL
DP



0-67-C94 H-8
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
Kmb
Kbs
Kgt
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
M
SF
TL
DP



Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
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Kmb
Kbs
Kgt
SF
TL
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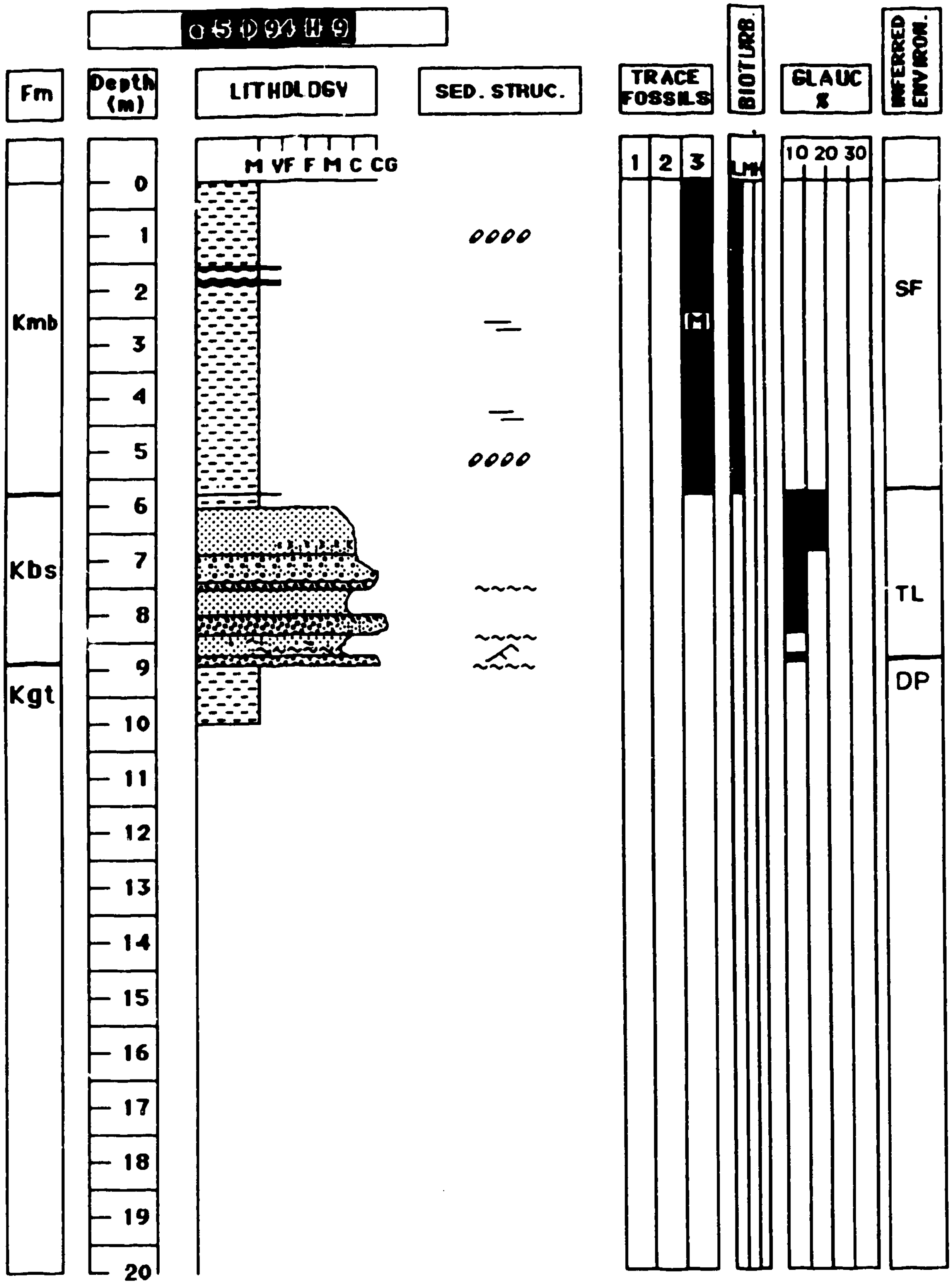
CB
TL
DP



05D94H9
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
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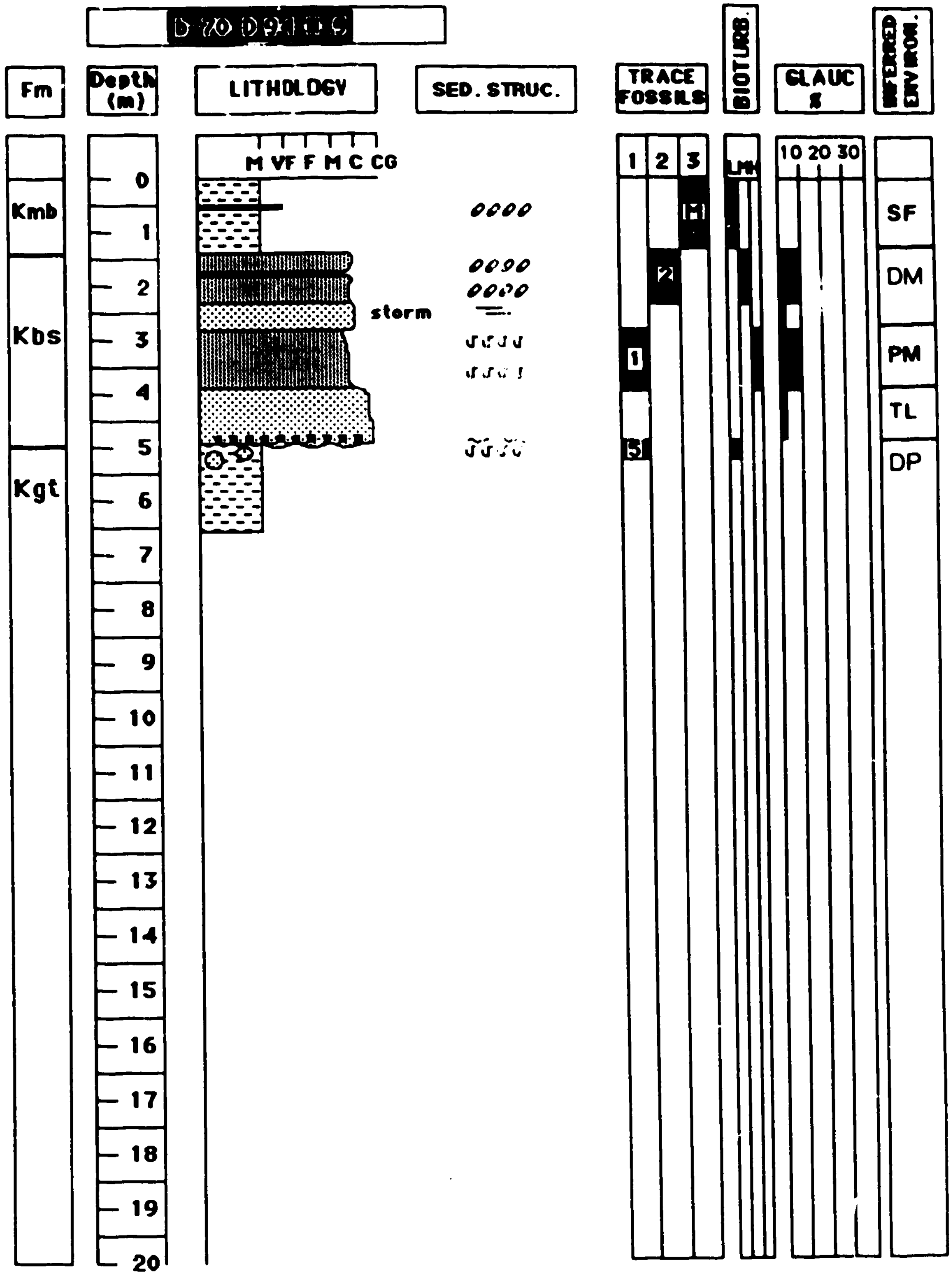
LMH
10 20 30
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
Kmb
Kbs
Kgt
SF
TL
DP


Fm
Depth (m)
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TRACE FOSSILS
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GLAUC %
INFERRED ENVIRON.
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1 2 3
10 20 30
Kmb
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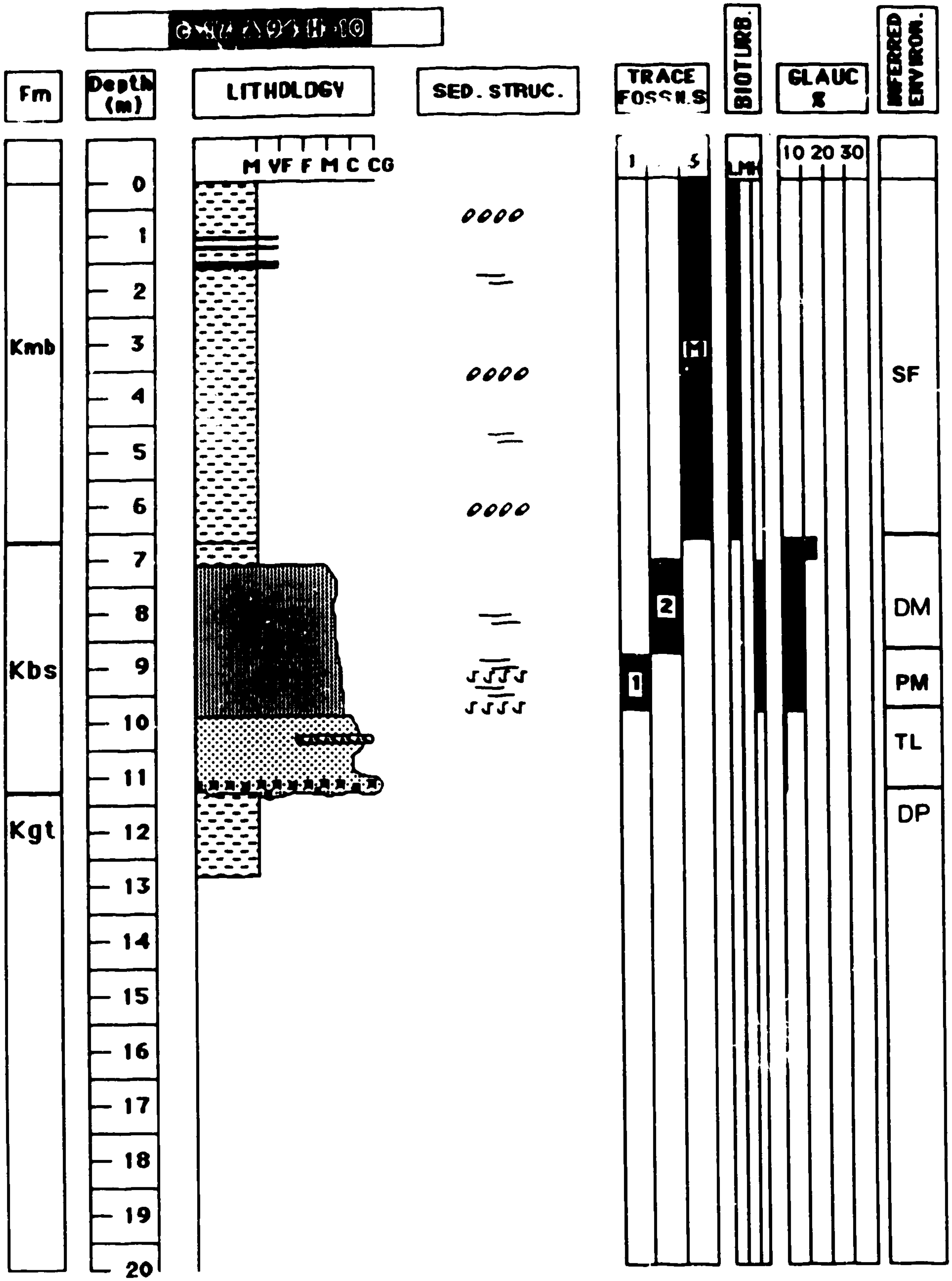

H-10
Fm
Depth (m)
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TRACE FOSSILS
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GLAUC %
INFERRED ENVIRON.
M VF F M C CG
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LMH
10 20 30
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Kmb
Kbs
Kgt
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TL
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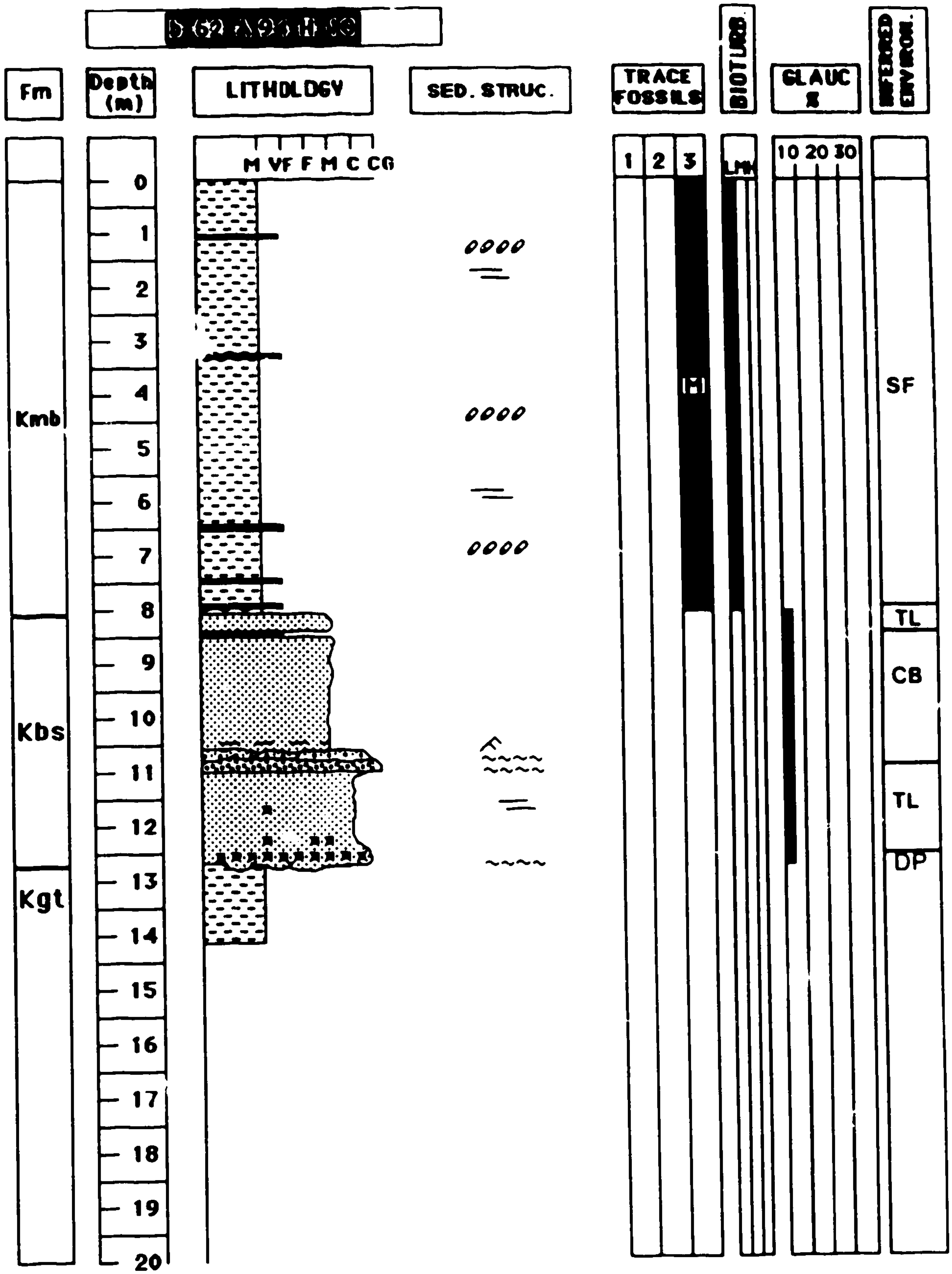

Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
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Kmb
Kbs
Kgt
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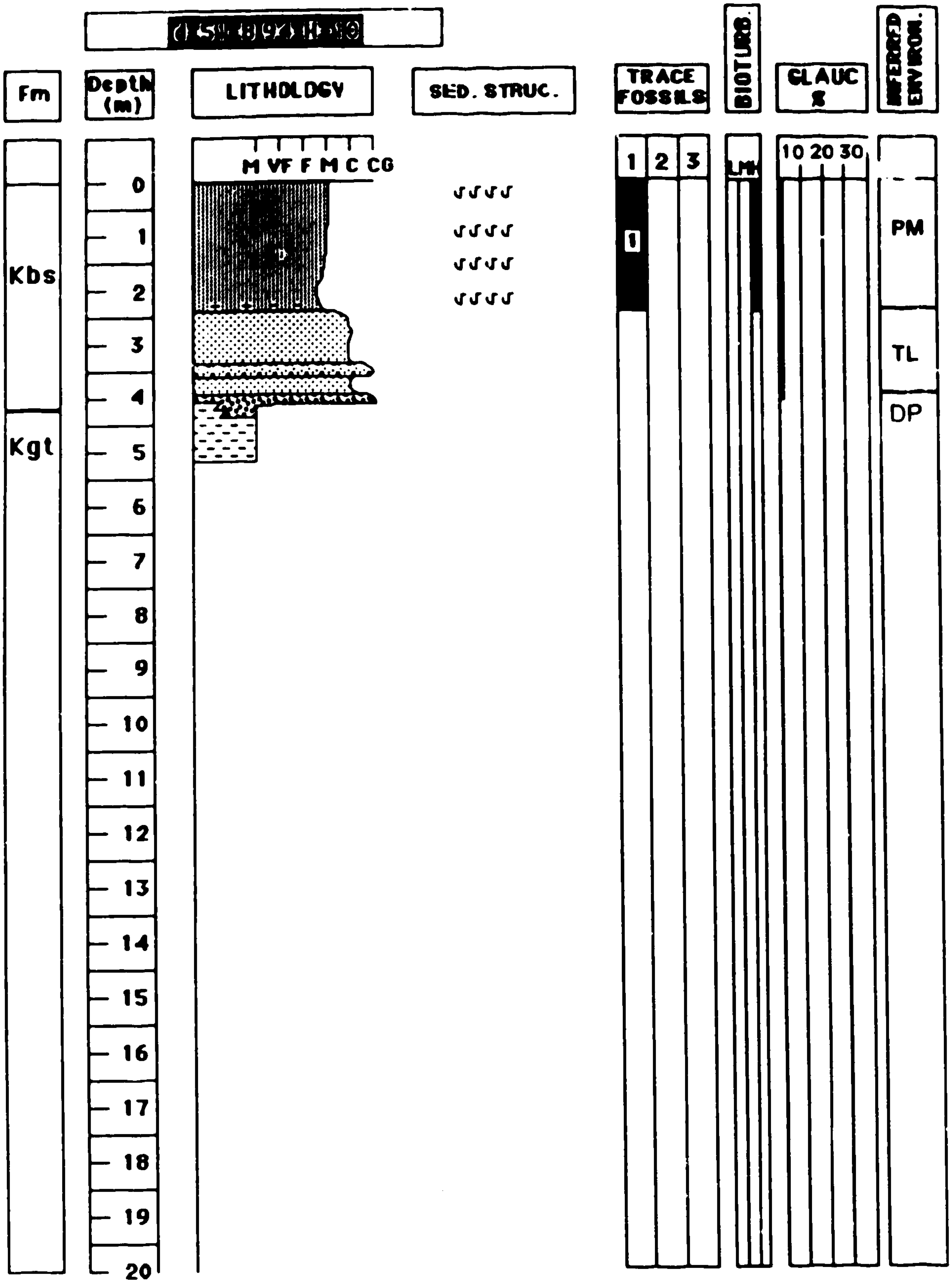
CB
TL
DP


Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
Kbs
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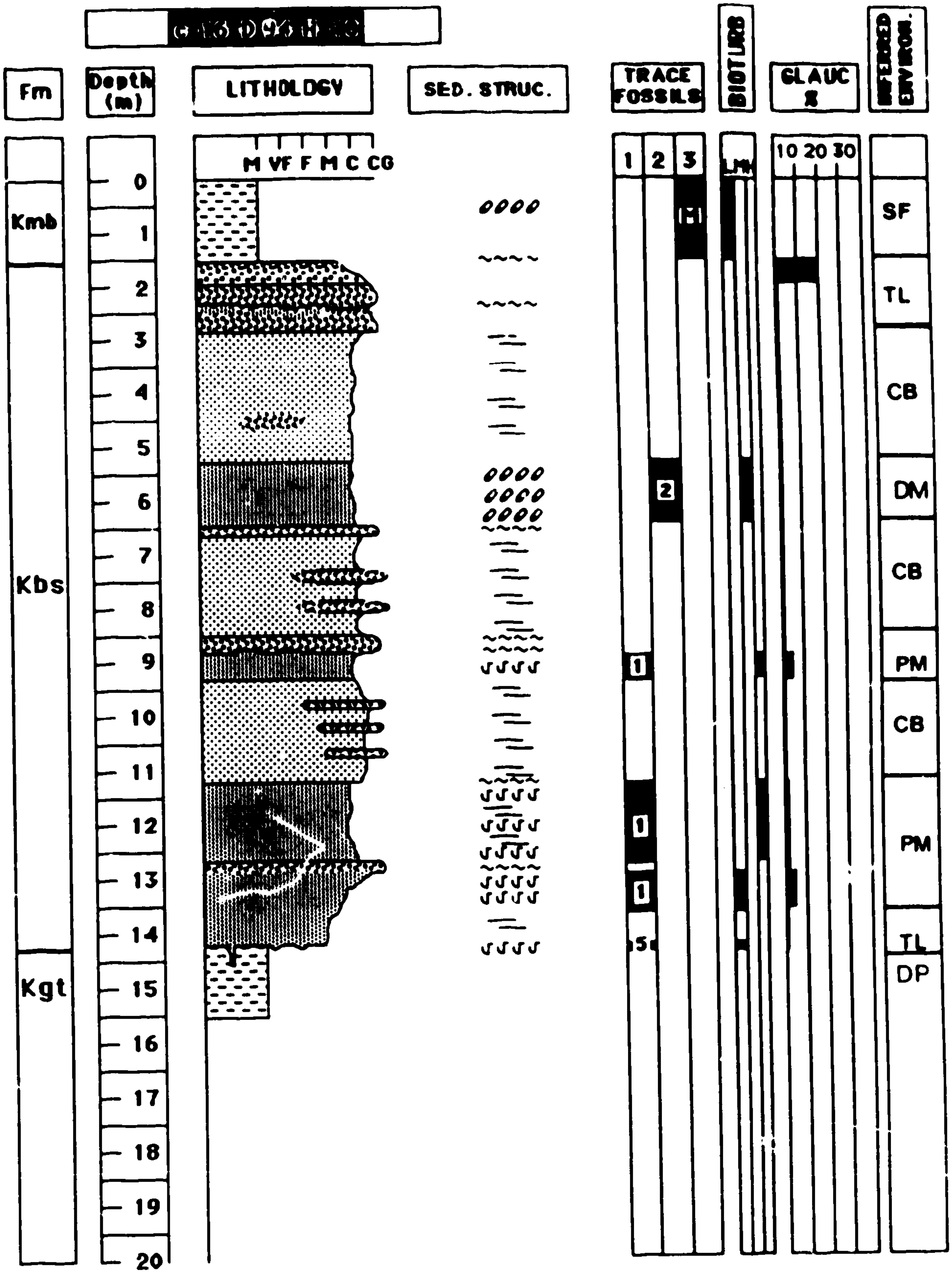
Fm
Depth (m)
LITHOLOGY
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TRACE FOSSILS
BIOTURB
GLAUC %
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1 2 3
10 20 30
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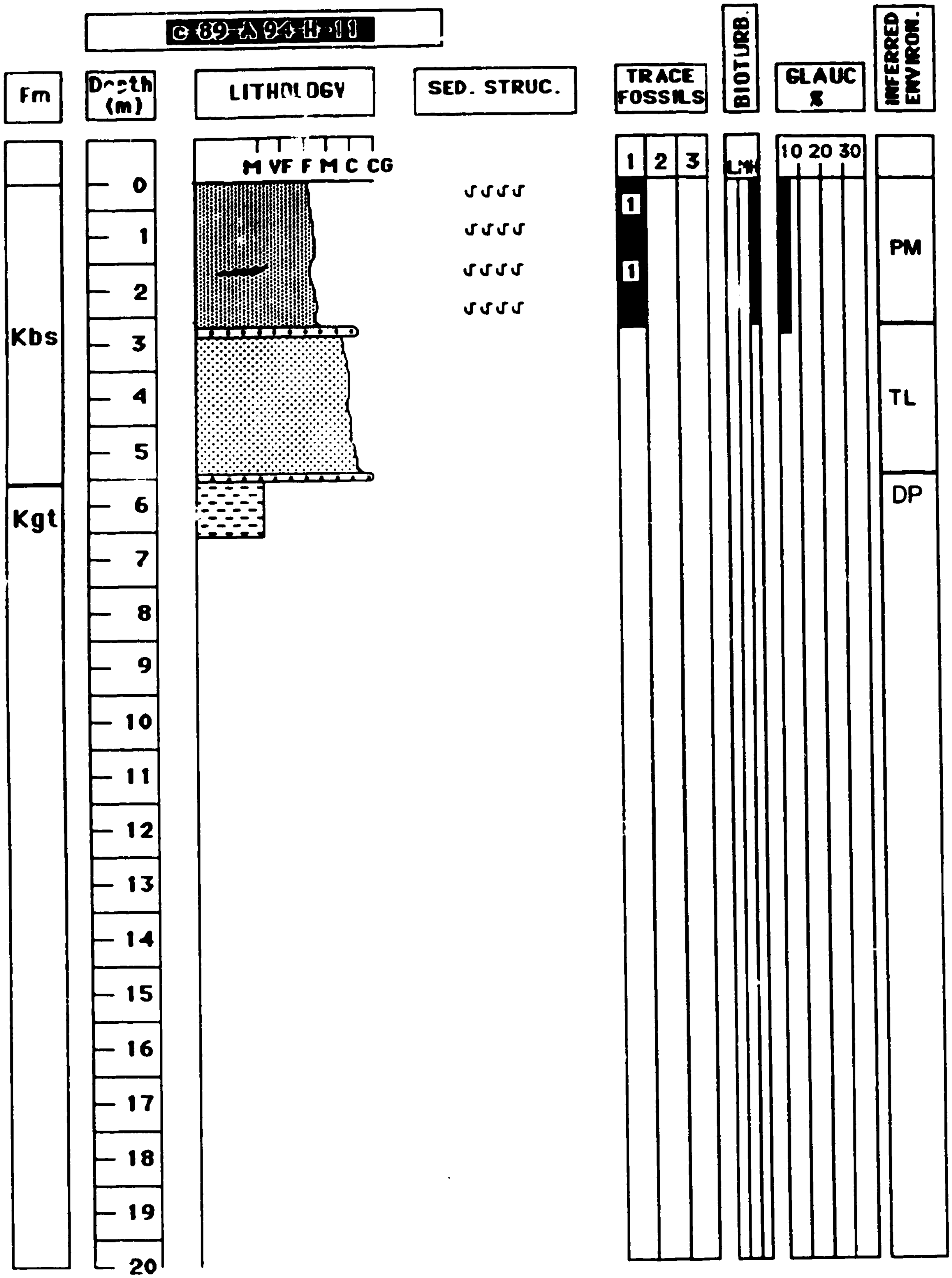

C-89-A 94-H-11
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
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GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
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Kbs
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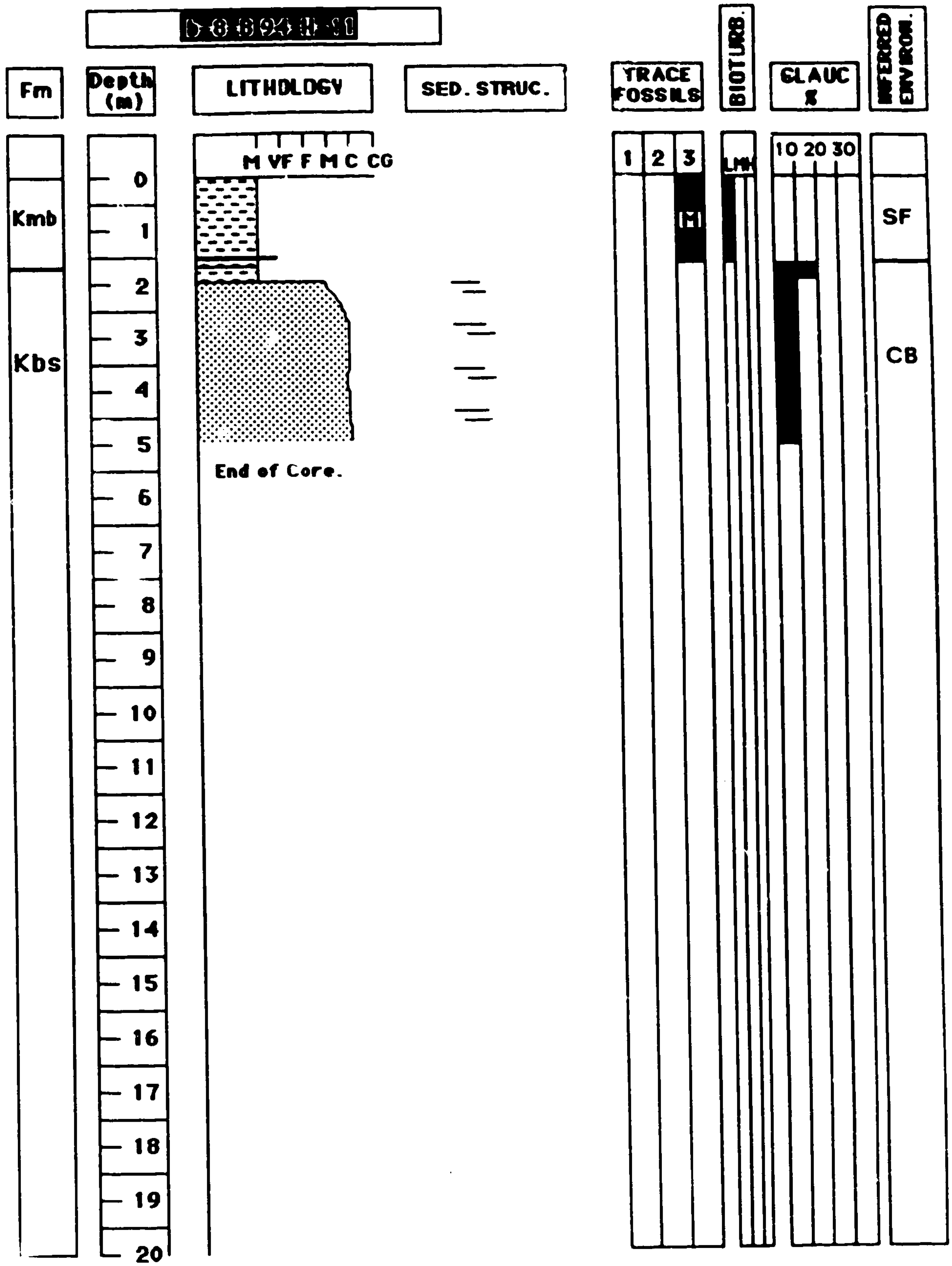

D-86-94-H-11
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
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GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
L M H
10 20 30
Kmb
Kbs
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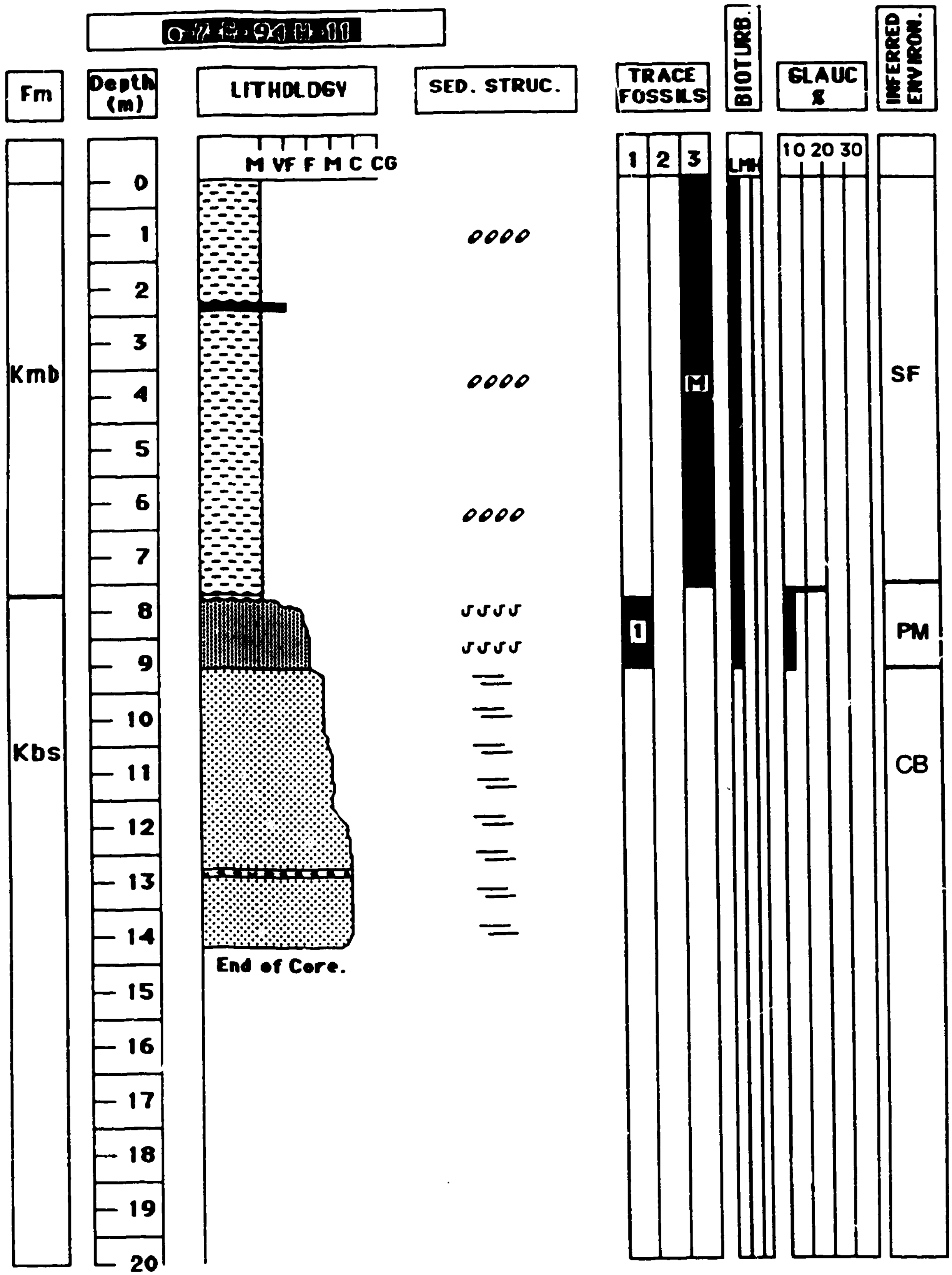
H
SF
CB
End of Core.



Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
Kmb
Kbs
M
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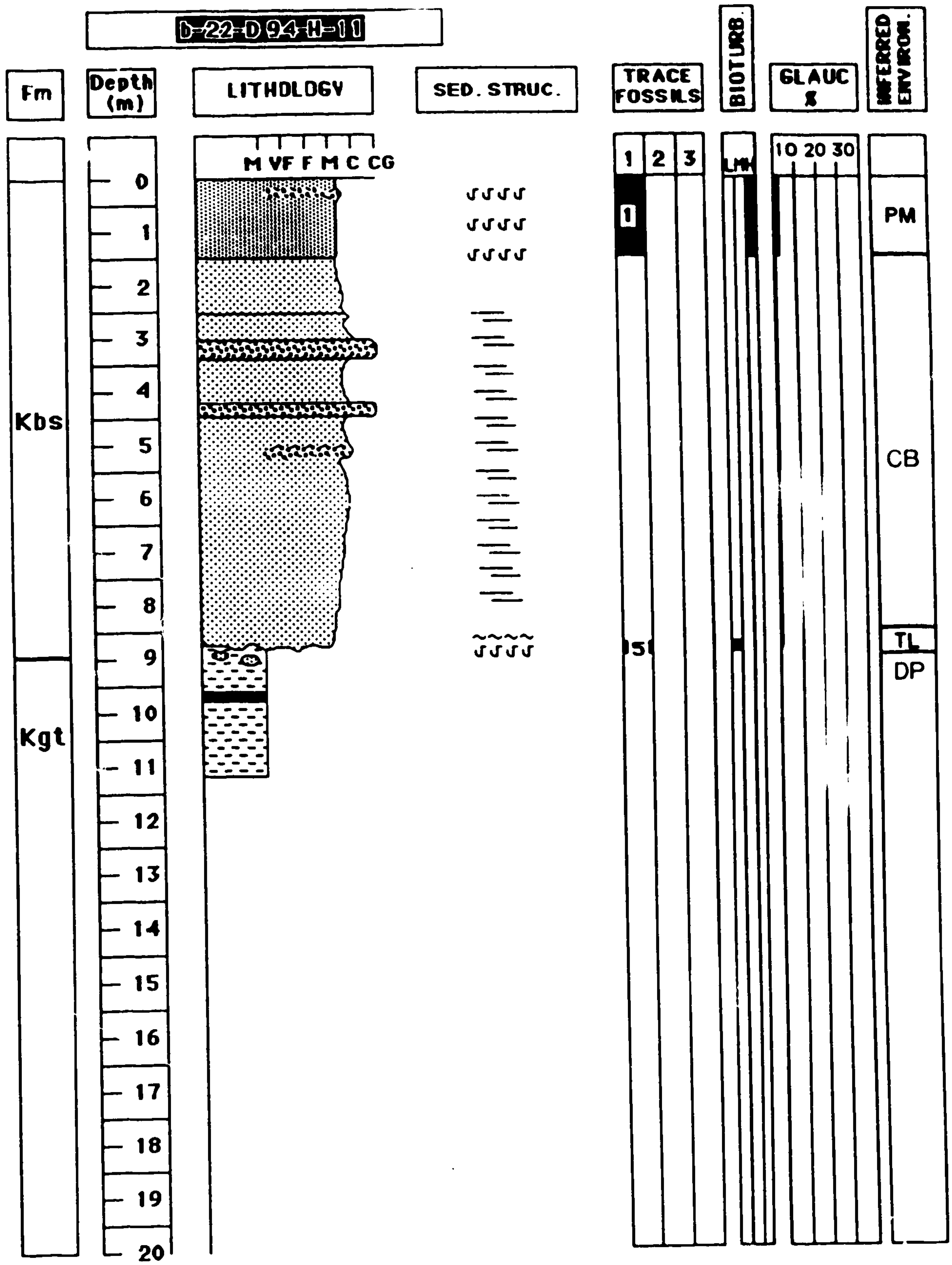
SF
PM
CB
End of Core.



D-22-D 94-H-11
Fm
Depth (m)
LITHOLOGY
SED. STRUC.
TRACE FOSSILS
BIOTURB.
GLAUC %
INFERRED ENVIRON.
M VF F M C CG
1 2 3
LMH
10 20 30
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
Kbs
Kgt
PM
CB
TL
DP