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GEOLOGICAL SURVEY

DEPOSITIONAL SYSTEMS AND THEIR PEAT/COAL-FORMING ENVIRONMENTS

By

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This report is preliminary and has not been reviewed for conformity with U.S. Geological Survey editorial standards and stratigraphic nomenclature.

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PREFACE

This notebook and the lecture series are a compendium of research investigations performed since mid-FY 1977 under the Branch of Coal Resources Project 9420-01977 on "Environments of Coal Deposition in the U.S. Western Interior Coal Basins." The research investigations were greatly facilitated by the research contributions of university faculty members and graduate students. Dr. Victor V. Cavaroc, Professor at North Carolina State University, conducted research investigations in the Gallup sag and Dr. Edward S. Belt, Professor at Amherst College, worked in the Williston Basin. Douglas A. Canavello, La Ree Lynn, Walter E. Marley, William J. Muldoon, Eugene D. Pait, Stuart Strum, John C. Toth, and Peter D. Warwick, all graduate students from North Carolina State University, conducted thesis work in the Powder River Basin, Wasatch Plateau, Raton Basin, and Williston Basin. Michael F. Erpenbeck, Casey L. Lepp, and Stephen M. Tur, graduate students from Texas Tech University, performed thesis work in the San Juan Basin, Raton Basin, and Williston Basin. John C. Hohman, a graduate student from Colorado State University, worked on a thesis in the Gallup sag. Rickey J. Wells, a graduate student from the University of Kentucky, worked in the Raton Basin. The contributions of these researchers to the overall knowledge on environments of coal deposition presented in this notebook and the lecture series are greatly appreciated.

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CHAPTER I

FACIES ANALYSIS

INTRODUCTION

In a simplified view, a fluvial system pertains to rivers and their channel networks, interfluves, and associated processes. Schumm (1977) described an idealized fluvial system as consisting of three geomorphologic zones in a downstream direction:

- Zone 1 -- This zone is the drainage basin, watershed, or source area where sediment production and/or sediment storage occur.
- Zone 2 -- This zone is the transfer zone wherein a stable channel network input of sediment can equal output.
- Zone 3 -- This zone is the sediment sink or area of deposition.

A single process dominates each zone, with zone 1 mainly dominated by sediment production, zone 2 predominantly by sediment transportation, and zone 3 chiefly by sediment deposition. Zone 3, which is a primary concern of geologists, includes depositional areas such as alluvial fans, alluvial plains, deltas and related coasts, and deeper waters. Thus, Schumm's (1977) geomorphologic classifications of the fluvial system include deltas and related coastlines as well as marine environments, which are otherwise classified as separate depositional systems by other workers (Fisher and others, 1969; Leblanc, 1972). Fisher and others (1969) defined depositional systems as assemblages of process-related sedimentary facies. This definition implies that a depositional system is an assemblage of interrelated environments and their associated processes, with each environment being physically, chemically, and biologically distinct. Correspondingly, deltas and coastlines (e.g. barrier and back-barrier systems), as well as the rivers that feed them, make up an interrelated depositional continuum or a spectrum of fluvial systems (fig. 1).

SEDIMENTARY FACIES

This notebook and lecture series will concentrate on the characteristics of deposits of ancient fluvial systems and their coal-forming environments. Although numerous depositional systems exist, only a few contain important coal-forming environments such as the alluvial, deltaic, and coastal barrier systems (fig. 1). Environments of these depositional systems are not difficult to identify in Recent sediments; however, environments of these depositional systems in the sedimentary record are best recognized by their sedimentary facies. A sedimentary facies is described (Reading, 1978; Selley, 1980) as a body of sedimentary rock that exhibits significantly distinctive properties (e.g. mineral composition, texture, lithology, geometry, sedimentary structures, paleocurrent patterns, and fossils). Thus, no two sedimentary facies are identical; however, gradational transitions between facies are common. A rock unit or a group of rock units defined by these distinctive properties is termed a lithofacies. Although the term facies may be used to describe a rock unit or a group of rock units in either an observational sense (e.g. "limestone facies," "detrital facies") or tectonic sense (e.g. "flysch facies"), it is preferred to be used in an environmental or process sense (e.g. "fluvial facies," "marine facies"). Thus, facies is intended to be used for products of an environment and/or subenvironment characterized by a specific process or interrelated processes.

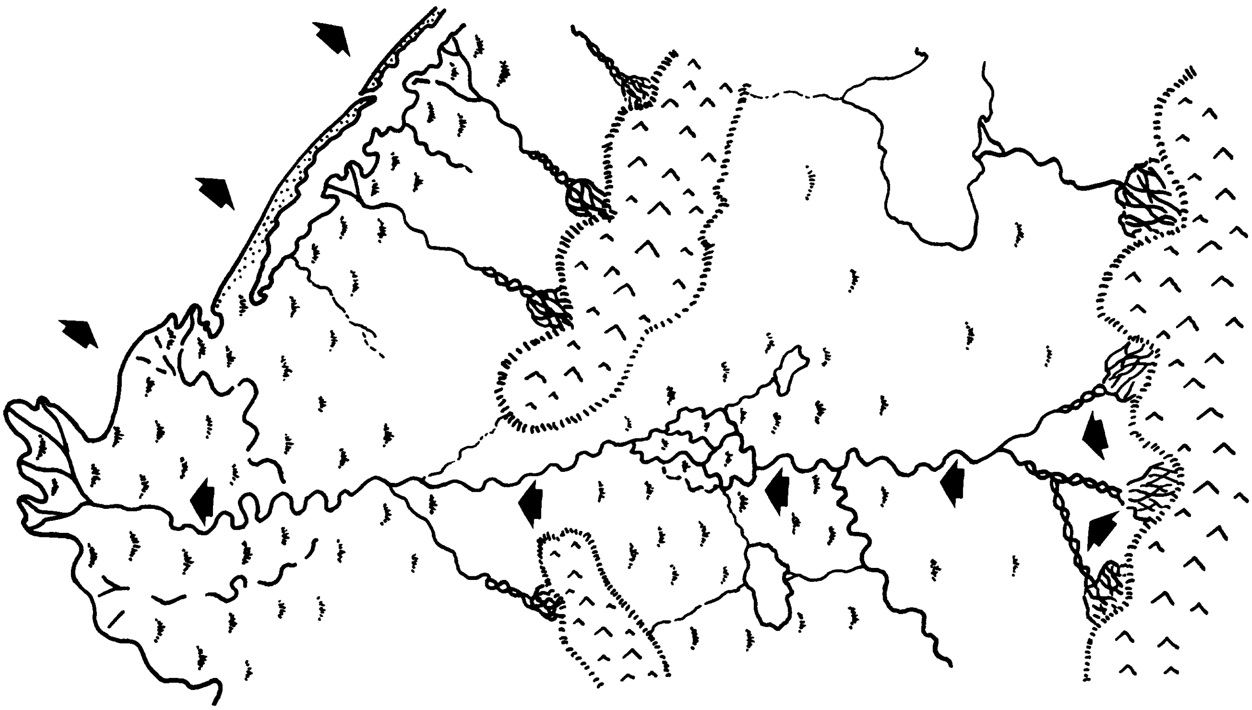

Figure 1.--An idealized spectrum of fluvial systems illustrating a depositional continuum from drainage basins through coastal areas and their coal-forming environments.

In order to interpret facies of a depositional system which consists of interrelated environments, it is necessary that facies be analyzed in terms of their relationships in space. The significance of facies relationships was first recognized in Walther's Law of Facies (1894). This law states that facies developed in a continuous vertical stratigraphic sequence was formed in laterally juxtaposed environments and that facies in vertical contact must be the product of geographically juxtaposed environments. Thus, according to this concept the environments of a depositional system may be determined if the geographic distribution and vertical succession of its constituent facies are known. This leads to analysis of facies associations and sequences that are environmentally related. A facies association refers to facies that tend to occur together; whereas, a facies sequence refers to a series of facies that pass vertically and laterally from one to another. An integration of the facies associations and sequences gives rise to construction of a facies model that simplifies understanding a complex of facies relationships and helps us visualize the processes and resulting environments.

CRITERIA FOR FACIES RECOGNITION

Recognition of facies to determine depositional environments can be grouped into petrographic and stratigraphic criteria. These criteria vary on the basis of whether the facies analysis is evaluated either from laboratory or field data generated from surface and subsurface investigations.

Petrographic Criteria

The petrographic criteria, which are generated from laboratory work, mainly consist of derived properties of sedimentary rocks that include mineral composition and texture. These properties are functions of the depositional environment, transportation, and source rocks of the sedimentary rocks. Mineral composition and texture are better indicators of depositional histories for detrital rocks than carbonate rocks; these properties in the latter are more susceptible to destruction by diagenesis.

Mineral Composition.--Mineral content as an indicator of depositional environments focuses primarily on diagnostic minerals that make up the framework mineral grains such as quartz, feldspar, and rock fragments. These minerals have been shown to be most sensitive to depositional processes characteristic of certain environments (fig. 2) as described by Folk (1968), Berg (1970), Pettijohn and others (1973), Davies and Ethridge (1975), and Flores (1978). The sensitivity of these framework mineral grains to specific depositional processes is recorded by their relative abundance. Depositional significance of the relative abundance of the framework mineral grains are best evaluated by statistical methods suggested by Krumbein and Graybill (1965), Griffiths (1967), and Davis (1973).

The clay matrix constituents of detrital rocks have limited use as environmental indicators (Grim, 1958; Weaver, 1958). Their limited utility is probably due to the complexity of their chemistry which reflects not only their diagenetic history but also weathering conditions, source rock, and environment of deposition. Perhaps, in coal-forming environments differentiation between freshwater and brackish-water swamps can best be analyzed by the clay mineralogy such as illite-montmorillonite, kaolinite, and boron content of clays of associated detrital rocks (Degens and others, 1957; Walker, 1968;

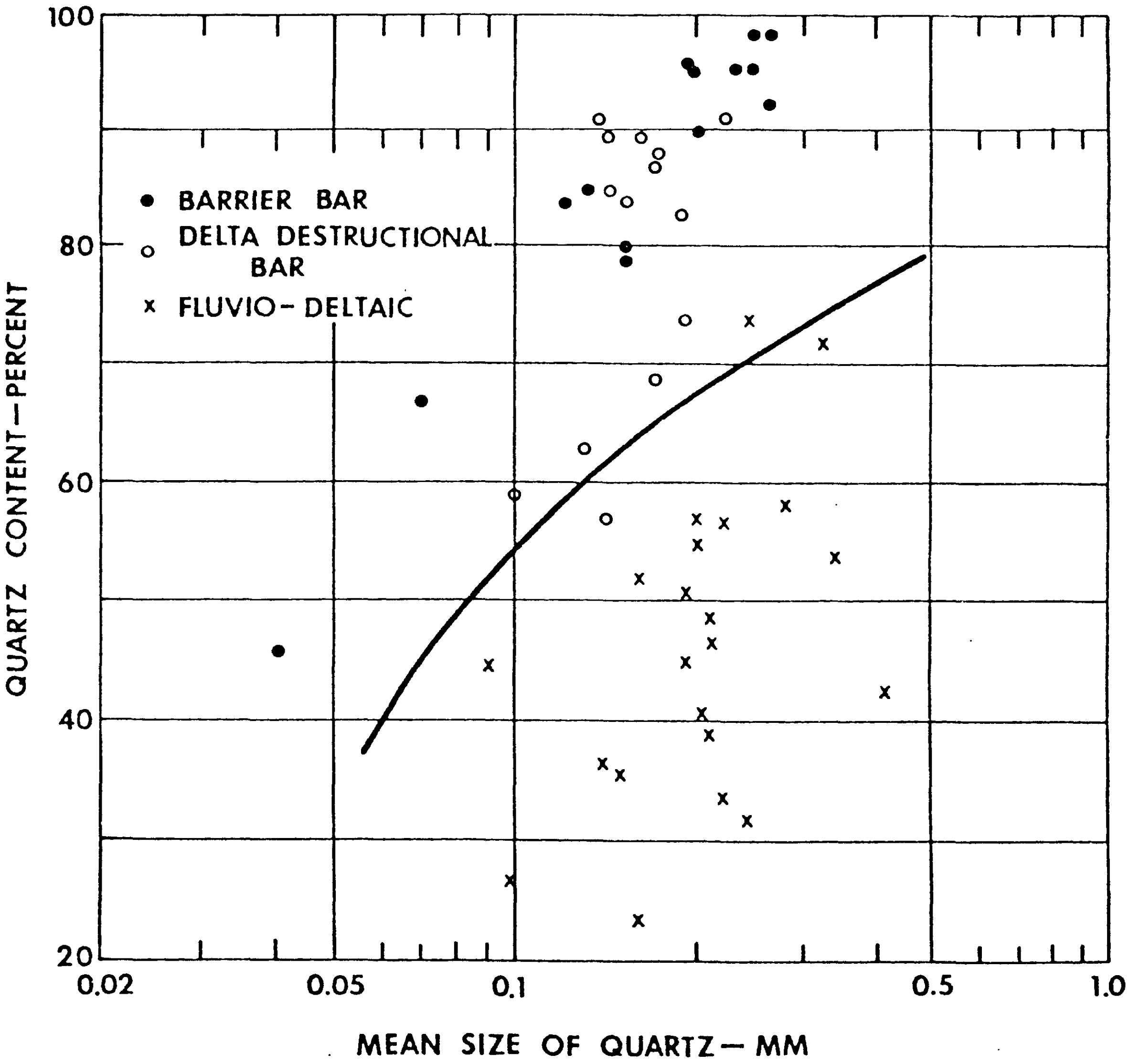

Figure 2.--Mineral composition and grain size directly related to environment of deposition. Adapted from Davies (1976).

Shimp and others, 1969; Bohor and Gluskoter, 1973; Selley, 1976; Ryer and others, 1980).

Texture.--There have been numerous attempts to use texture to determine depositional environments of sedimentary rocks. Textural parameters that are used in evaluating depositional environments of detrital rocks are skewness and kurtosis (Folk, 1966; Moiola and Weisner, 1968), log probability plots of grain-size distribution (Visher, 1969; Glaister and Nelson, 1974), and grain shape and roundness (Reineck and Singh, 1980). The philosophy behind the study of these parameters is that a detrital rock deposited in a specific environment can be identified from another by statistical analysis of its texture. In general, grain size of a detrital rock is an important indicator of the energy level of the depositional environment. That is, the coarser the grain size, the higher the current energy that deposited the detrital rock. In addition, prolonged application of the high energy level produces better sorted detrital rocks. However, these relationships are dependent on grain-size distribution of parent materials in the source area. Thus, like mineral composition, the use of texture to diagnose environment of deposition is limited. Reviews of problems of this approach are reported by Folk (1966) and Selley (1976).

Stratigraphic Criteria

Diagnosis of depositional environments of sedimentary rocks on the basis of stratigraphic criteria consists chiefly of critical evaluation of field observations of the lithology, geometry, sedimentary structures, paleocurrent patterns, and fossils.

Lithology.--The lithology of sedimentary rocks is one of the easiest stratigraphic criteria to observe in the field. Lithology is defined in part by the AGI Glossary of Geology as "the study of rocks based on megascopic examples of samples." This study, at first hand, is best performed in the field. This notebook will not attempt to provide a lithological classification of sedimentary rocks because it is adequately reviewed by Jackson (1970), Blatt and others (1972), and Pettijohn (1975). More specifically, classification of sandstones based on mineralogical content and amount of clay matrix (fig. 3) is reviewed in detail by Krynine (1948), Gilbert (1955), McBride (1963), Dott (1964), Folk (1968), and Pettijohn and others (1973). Although there is a lack of complete philosophical agreement among different classifications of sandstones, one concensus is their use of three common mineral species or group of rock fragments as components of poles of a triangle. On the basis of relative abundance of these mineral constituents, the triangle is internally subdivided according to sandstone types (e.g. arkose, graywacke, quartz arenite). These sandstone types, in turn, can be subclassified according to the detrital clay content of their matrix (e.g. wackes) and the amount of labile or unstable rock fragments (e.g. litharenite, lithic arkose).

Classification of fine-grained detrital rocks or "mudrocks" is best described by Potter and others (1980). Classification of "mudrocks," which is based on relative abundance of "silt-size" and "clay-size" particles as well as on fissility and nonfissility, is more simplistic than the classification of sandstones. Classification of carbonate rocks based on mineralogical composition was developed by Folk (1959), and a classification based on depositional texture is described by Dunham (1962). Both types of classification are readily adaptable to field usage.

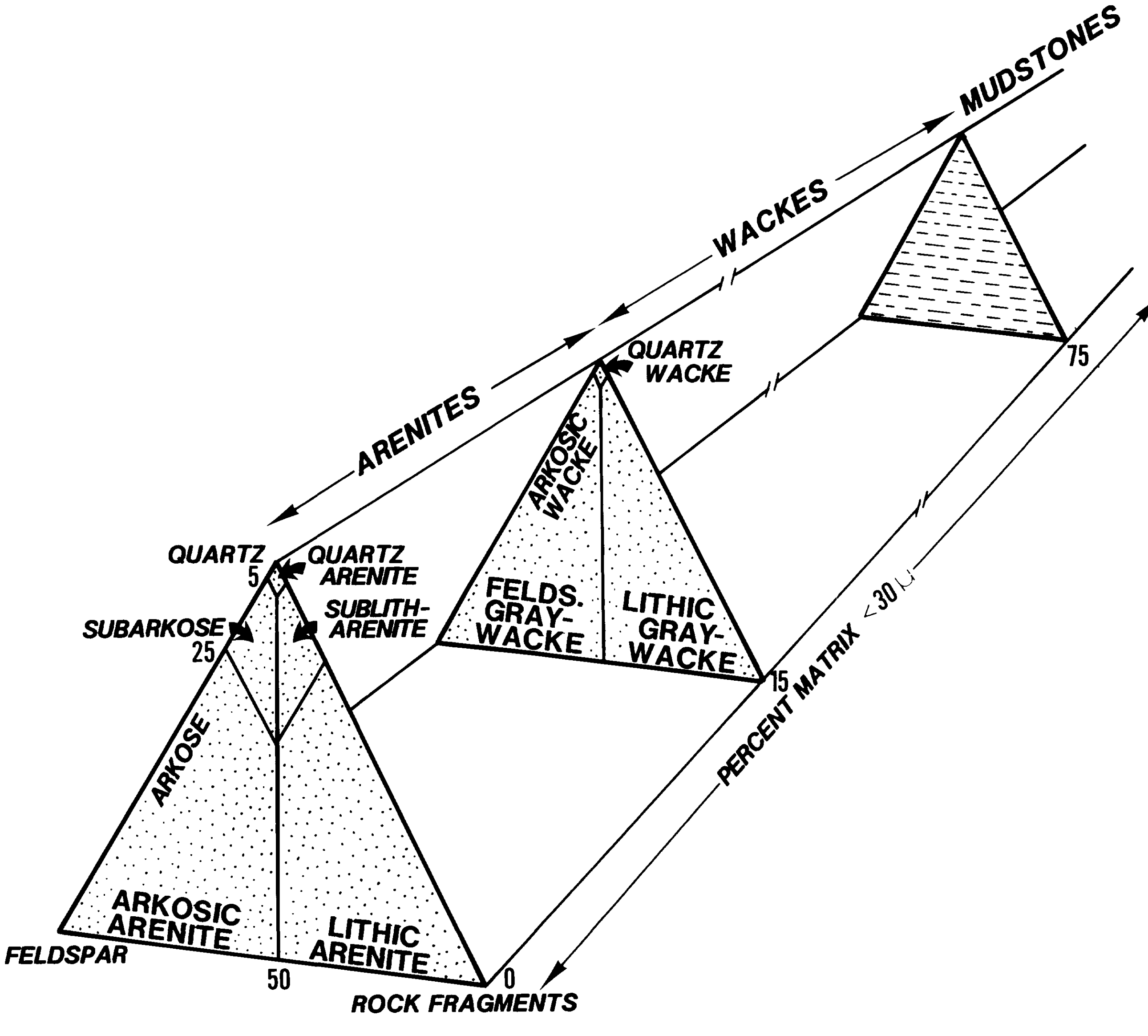

Figure 3.--Classification of sandstones based on framework mineral constituents and detrital clay matrix. Adapted from Dott (1964).

Geometry.--Sedimentary rocks exhibit a variety of shapes as functions of paleotopography or geomorphology of the depositional environment (e.g. channel geometry), nature of sediment transport and deposition (e.g. blanket geometry), and proximity to source areas (e.g. fans, aprons). The shape of a deposit of a depositional environment can be expressed in terms of the relationship between different size factors. According to Krynine (1948), the relationship between area and volume (a fundamental shape factor) can best be expressed by a ratio between width and thickness which defines types of sedimentary bodies into:

1. Blanket shape (1000 to 1).
2. Tabular shape (1000 to 1 and 50 to 1).
3. Prism shape (50 to 1 and 5 to 1).
4. Shoestring shape (less than 5 to 1).

Interest in geometry of sedimentary rocks has concentrated on shapes of sandstone bodies first recognized by Rich (1938) and later described in detail by Potter and Pettijohn (1963), Leblanc (1972), Shelton and others (1972), and Weimer (1973) on the basis of genetic and geometrical dimension. Pettijohn (1975) reviewed in detail various types of external morphology of sandstone bodies and devised a generalized geometrical classification:

1. Linear or shoestring sandstone.
2. Complex and bifurcating linear sandstone.
3. Wedge-shaped sandstone.
4. Sheet sandstone.

Shoestring and bifurcating sandstones are illustrated in figure 4. Although geometry of sedimentary facies is an important criterion, it is only when syndepositional shape of the environment is preserved that geometry is diagnostic. This condition is exemplified by anastomosed patterns of alluvial channels (fig. 4), a radiating pattern of deltaic distributary channels, and a sheetlike geometry of a barrier coastline.

Sedimentary Structures.--Perhaps the most important indicator of depositional environments of sedimentary rocks is sedimentary structures. Of the data that can be generated from field observations, sedimentary structures are the most clearly seen and are best studied in outcrops. Sedimentary structures can indicate deposition of sedimentary rocks in either glacial, aqueous, or subaerial environments (Conybeare and Crook, 1968), and they provide evidence of depth and energy conditions of the depositional environments as well as velocity, hydrodynamics, and directions of currents that flowed across the environment (Middleton, 1965; Allen, 1967, 1970; Harms and others, 1975; Reineck and Singh, 1975). Interpretation of sedimentary structures has greatly benefited from flume experimental studies (Middleton, 1965; Harms and others, 1975) and from studies in Recent sediments (Allen, 1970; Coleman and others, 1964, 1974). Sedimentary structures can be genetically and morphologically classified into three types:

1. Hydrodynamic or current-produced structures.
2. Rheologic or hydroplastic synsedimentary-deformation structures.
3. Biogenic or organism-produced structures.

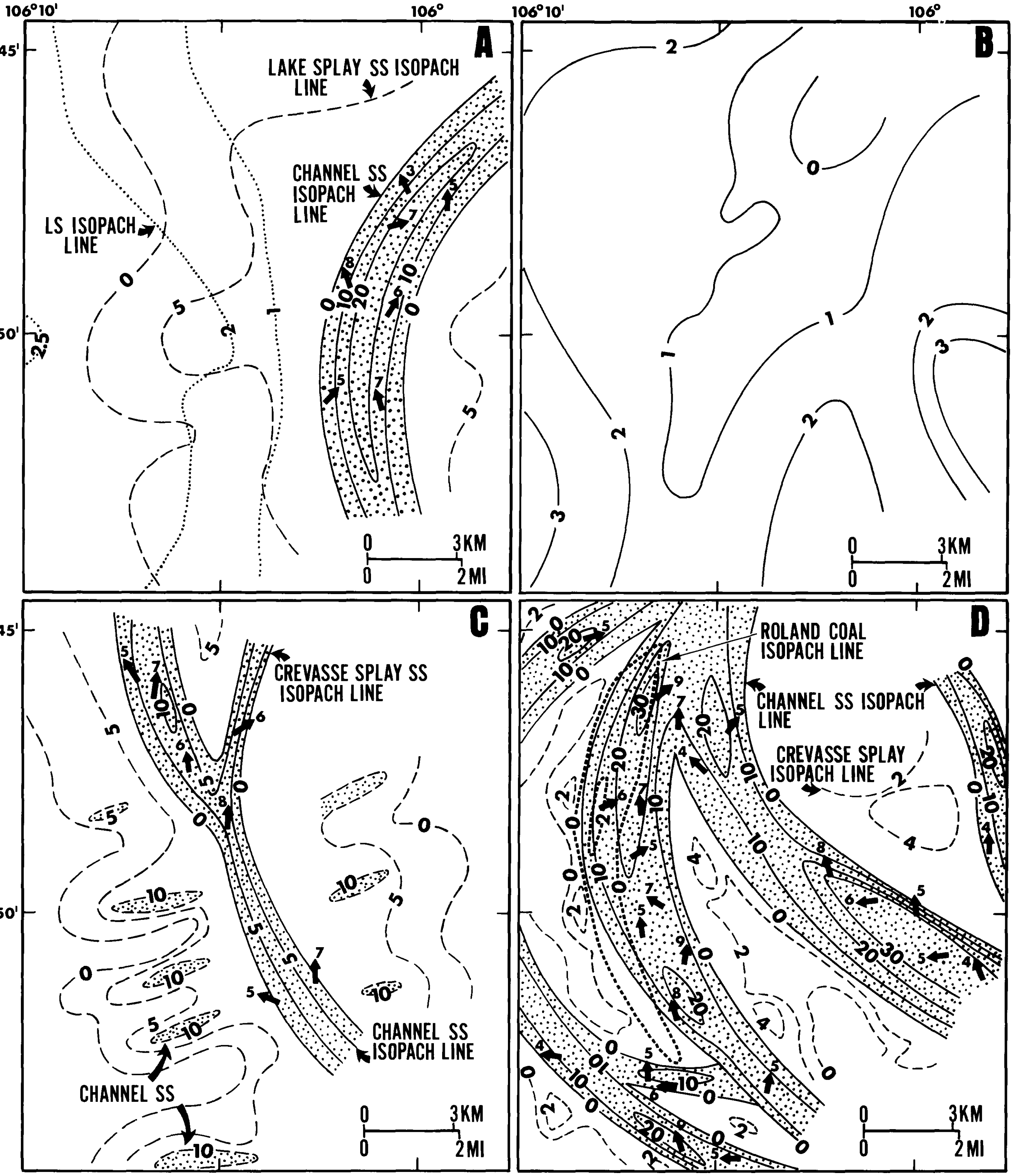

Figure 4.--Geometry of channel sandstones (D) in anastomosed fluvial deposits of the Upper Tongue River Member of the Fort Union Formation. Adapted from Flores and Hanley (1983).

Table 1 summarizes the classification of sedimentary structures. The hydrodynamic-propagated structures can be divided into internal and bedform subtypes. The synsedimentary deformation structures primarily formed during postdeposition. These structures, otherwise known as soft sediment deformation structures, can be grouped into two genetic classes:

1. Class I - displays vertical reorientation of bedding.
2. Class II - illustrates lateral rearrangement of rock units.

The first class is formed by sand collapse into water-saturated mud (e.g. ball-and-pillow structures, diapirs, and contortions) triggered by earthquake, current turbulence, and hydrostatic pressure. The second class includes slumps and slides as well as microfolds and microfaults generated by slope failure, storm, and differential loading and/or compaction. Because biogenic structures are grouped as ichnofossils and largely unrelated to hydrodynamic and rheologic conditions, they will be treated as fossil components of sedimentary facies. The descriptions, processes, and associations of sedimentary structures of various genetic rock types are treated separately in chapter II of this notebook.

Paleocurrent Patterns.--Sedimentary structures indirectly provide a guide to mapping paleocurrents as well as determining paleoslope and depositional strike. Of the stratigraphic criteria for recognition of facies, all but the paleocurrent patterns are directly observed in the field. Paleocurrent patterns are derived from interpretation of measurements of dip direction or inclination of foresets of trough and planar cross laminations. Dip direction must be interpreted from foresets of either one of the internal structures (e.g. trough cross lamination) or the other (e.g. planar cross lamination) but never from both types of cross lamination. Otherwise, lumping measurements from all types of internal structures will yield confused paleocurrent patterns that are impossible to interpret (Allen, 1967). In addition, paleocurrent analysis of facies should deduce the paleocurrent directions from a statistically prudent number of observations at each sample point or locality (Potter and Olson, 1954; Potter and Pettijohn, 1963). Normally, half a dozen or so measurements of cross laminations in a univariate condition will suffice per locality; however, 20 or more measurements are necessary for multivariate conditions (Potter and Pettijohn, 1963). These observation points are integrated to prepare a regional paleocurrent map in relation to other facies (see fig. 4). The paleocurrent pattern from each sample locality is integrated with those of other sample localities and summarized in windrose diagrams (Selley, 1968).

Paleocurrent windrose diagrams consist of three patterns: unimodal, bimodal, and polymodal or random. The modality of these paleocurrent patterns is a function of the frequency distribution of current azimuths. A bipolar pattern, which is a variant of bimodal pattern, shows a frequency distribution of current azimuths in which a principal mode is directly opposed (180) by a lesser mode forming a "bow-tie" windrose pattern. Table 2 summarizes the patterns of paleocurrents of fluvial systems based on data from Selley (1968). Although current directions measured from trough and planar cross laminations proved valuable to the interpretation of depositional environments, other vector measurements from ripple laminations, flutes, grooves, parting lineations, oriented fossils, and pebble and grain orientation proved to be the less useful tools. Thus, great care should be taken in selecting the types of

Table 1.--Classification of sedimentary structures

Hydrodynamic structures	Synsedimentary deformation structures	Biogenic structures
A. Internal structures	A. Vertical rock reorientation	A. Borings
1. Trough or festoon cross lamination	1. Ball-and-pillow structures	B. Tracks and trails
2. Planar or tabular cross lamination	2. Diapirs	C. Casts and molds
3. Ripple lamination (symmetrical and asymmetrical)	3. Contortions	
4. Parallel (horizontal) and subparallel lamination	B. Lateral rock reorientation 1. Slumps and slides	D. Fecal pellets
5. Convolute lamination	2. Microfolds	
B. Bedding (on surface) structures	3. Microfaults	
1. Dunes (megaripples)		
2. Sandwaves		
3. Ripples (lunate, linquoid, rhomboid, and straight-crested)		
4. Swash marks		
5. Pits and prints		
6. Parting lineations		
7. Mud cracks		
C. Bedding (on sole) structures		
1. Flute marks		
2. Tool marks		
3. Load casts		
4. Scour marks		

Table 2.--Paleocurrent patterns in the fluvial systems

[Modified from Selley (1980)]

Environment	Local current vector	Regional current pattern
A. Alluvial		
1. Braided fluvial channels	Unimodal, low variability	Often fan shaped
2. Meandering fluvial channels	Unimodal, high variability	Slope-controlled, often centripetal basin fill
B. Deltaic	Unimodal	Regionally radiating
C. Barrier coastlines	Bimodal (due to tidal currents), sometimes bipolar or polymodal	Generally consistently oriented longshore, onshore or offshore

sedimentary structures and to measure paleocurrent directions, as well as in the application of statistical analysis of paleocurrent vectors for recognizing ancient depositional environments and their paleogeographies. The relationship between paleocurrents and paleogeographies is discussed by Potter and Pettijohn (1963) and Pettijohn (1975).

Fossils.--The study of fossils is an integral part of facies analysis and one of the most important tools for identifying depositional environments of sedimentary rocks. The method of study of body fossils (megafossils and microfossils) used for recognizing depositional environments relates to paleoecology. Paleoecology deals with an investigation of the habits and habitats of fossil organisms in relation to their living counterparts. Paleoecological interpretation of paleoenvironments that controlled geographic and time-stratigraphic distributions of fossil organisms requires integration of lithostratigraphic, sedimentologic, and depositional-process data (Flores and Hanley, 1983). Thus, paleoecological study is an interdisciplinary effort between a sedimentologist and paleoecologist. A paleoecologist cannot disregard lithologic investigation, nor, vice versa, a sedimentologist cannot disregard fossil content of facies. Only in a joint effort can both investigators find a key to the solution of problems on the modes and conditions of fossil organisms, on one hand, and the nature of depositional environments, on the other. This interdisciplinary relationship is evident since specific paleoecological precursors concern the interpretation of the history of fossil organisms from time of death to final burial (biostratonomy) and the post-burial (diagenetic) condition. Post-mortem depositional processes reflected by rock type and fabric, sedimentary structures and stratification sequences, and biofabric (size, shape, orientation, and distribution of fossil organisms) are demonstrated by Flores and Hanley (1983) in alluvial deposits. These sedimentological and biological parameters ascertain whether remains of fossil organisms are of a life association (paleobiocoenosis) or are an accumulation of organic remains merely buried together (paleothanatocoenosis). Investigations of paleoecology have been reviewed by Hedgpeth and Ladd (1957), Ager (1963), Imbrie and Newell (1964), and Hecker (1965).

Trace fossils or ichnofossils are useful in interpretations of depositional environments because they occur in situ and are diagnostic of certain environments. Seilacher (1964) recognized five classes of trace fossils based on the behavior of organisms:

1. Resting marks made by mobile animals lodging on the bottom.
2. Crawling trails made by mobile animals while moving over sediment surface.
3. Shelter structures or burrows that are permanent and made either by mobile or semi-attached animals.
4. Feeding structures or burrows made by sessile bottom feeders.
5. Grazing sinuous trails or burrows of mud-eating organisms at or below the sediment-water interface.

Biogenic structures can also be classified on the basis of their relation to the bedding, geometrical configuration and ornamentation or internal structures (Pettijohn, 1975). The use of these classifications permits facies to be defined in terms of ichnofacies in which each ichnofacies consists of a suite of trace fossils that occur in a characteristic sedimentary facies

(Seilacher, 1967; Rodriquez and Gutschick, 1970; Heckel, 1972). A generalized classification of biogenic structures is given in table 1.

PALEOGEOGRAPHIC RECONSTRUCTION

The final stage of facies analysis is a reconstruction of a paleogeographic map or paleogeography (fig. 5) that depicts the orientation and approximate sites of environmental belts. Perhaps, Sorby (1857) best described what is known today as a paleogeographic reconstruction. Sorby wrote:

> "The examination of modern seas, estuaries, and rivers shows that there is a distinct relation between the physical geography and the currents present in them; currents so impress themselves on the deposits formed under their influence that their characters can be ascertained from those formed in the ancient periods. Therefore their physical geography can be inferred within certain limits."

Moore (1949) suggested that sedimentary facies be considered to comprise any areally segregated part of a time-stratigraphic rock unit in which physical and organic characters differ significantly from those of another part or parts. Thus, since an assemblage of facies reflects areal variation of environments of depositional systems, the most effective method of showing geographic facies variations is by maps. Most facies maps are what have been identified as lithofacies maps. The construction of facies maps and their paleogeographic interpretations have been given a great deal of attention in a large number of publications (Sloss and others, 1949; Krumbein, 1952; Krumbein and Sloss, 1963; Reading, 1978; Walker, 1979). The fundamental idea of paleogeographic reconstruction of sedimentary rocks basically follows the concept of "consanguinity" utilized in igneous petrology (Iddings, 1892). That is, igneous rocks that are members of a group associated in time and space and with a community of characters or family likeness make up a "consanguineous association." Thus, the use of consanguineous association is extended into sedimentary rocks such that the cogenetic association reflects facies association. The spatial arrangement of time-stratigraphic facies associations and their patterns permit an interpretation of paleogeography.

Paleogeographic reconstruction can be complemented by paleocurrent analysis (Pettijohn, 1962; Potter and Pettijohn, 1963). Paleocurrent analysis contributes to paleogeographic reconstruction in ascertaining sedimentary strike, delineating paleoslope, and inferring the trend of paleoshoreline as well as defining the trend and location of margins of a depositional basin. The ultimate test of paleogeographic interpretation of sedimentary rocks is its comparison to Recent sediments and their environments. This approach of Huttonian "uniformitarianism," which has been the watchword of field geologists for a century and a half, still remains the most definitive explanation for the rock record of past events.

SUMMARY

In summary, this introductory chapter has reviewed how depositional environments may be deduced from facies sequences, associations, and relationships of sedimentary rocks as defined by their mineral composition, texture, lithology, geometry, sedimentary structures, paleocurrent patterns,

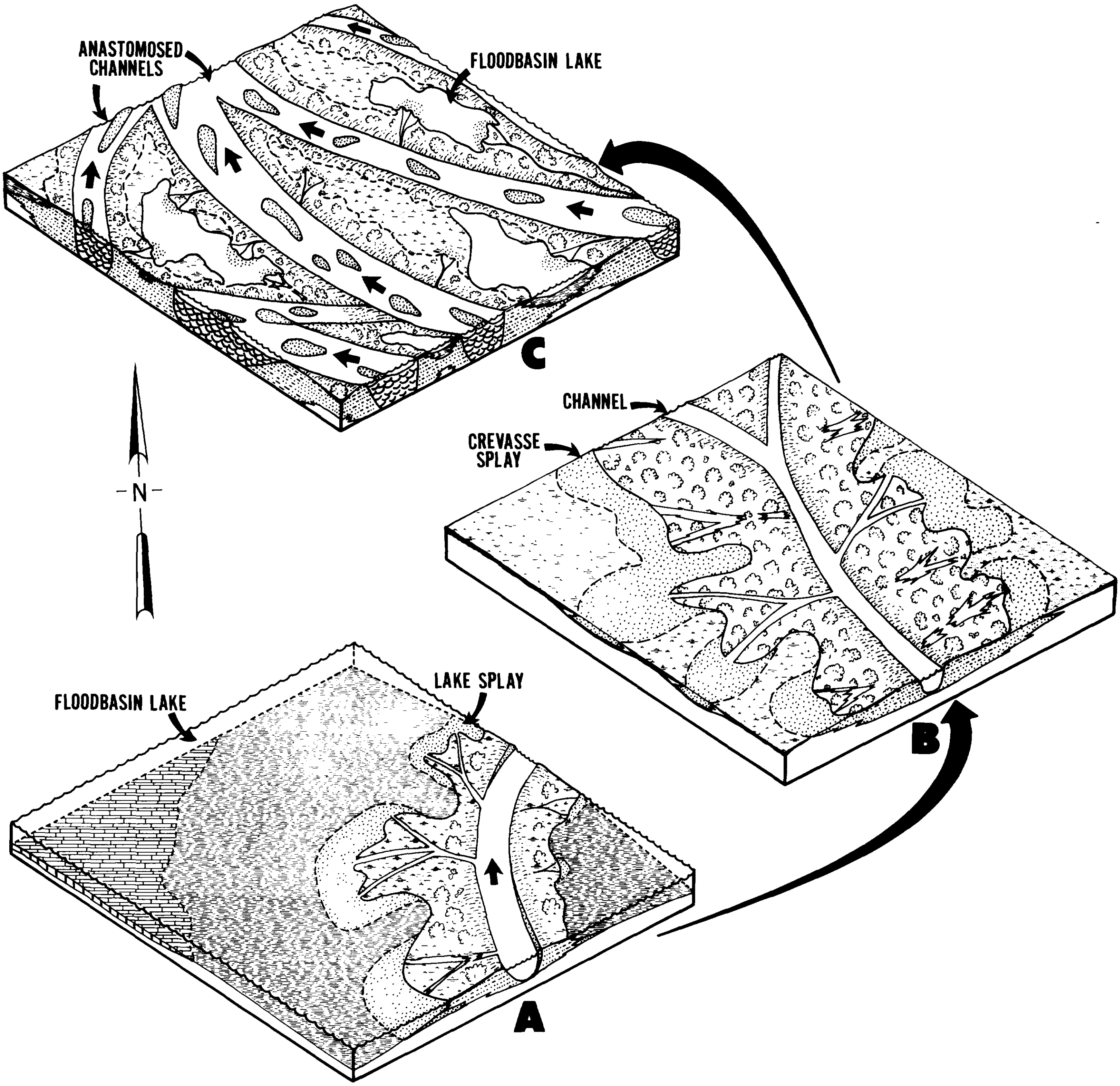

Figure 5.--Paleogeographic map (C) of the anastomosed fluvial deposits shown in figure 4. Adapted from Flores and Hanley (1983).

and fossil content. This basic approach to the analysis of ancient sedimentary environments is summarized in figure 6. Although these various methods were described for recognizing facies in the interpretation of their depositional environments, an alternative and more appropriate approach is to show specific case histories of ancient coal-bearing deposits, which are summarized in chapter III.

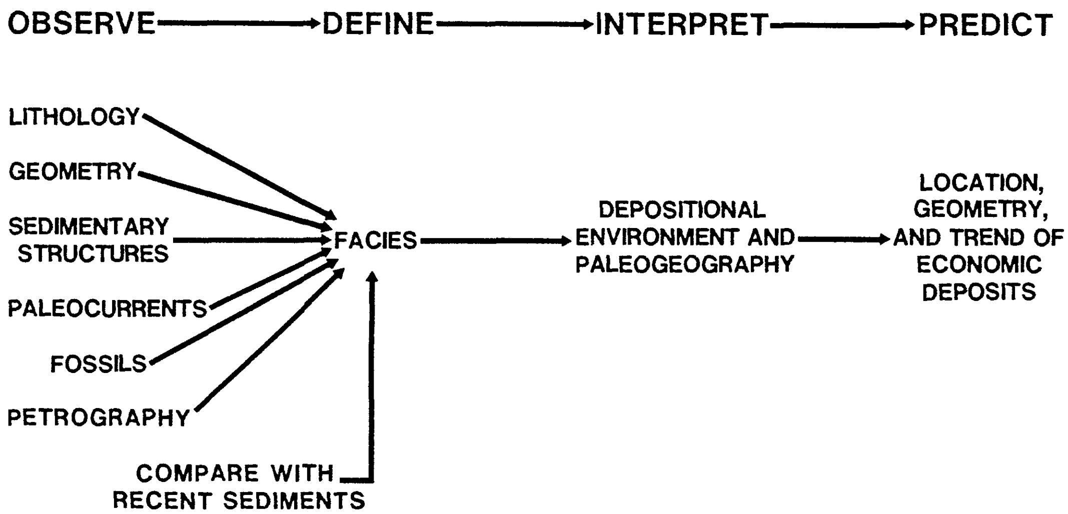

Figure 6.--Flow chart of basic approach to analysis of sedimentary environments. Modified from Selley (1980).

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CHAPTER II

COAL-ASSOCIATED FACIES AND THEIR SEDIMENTARY STRUCTURES

INTRODUCTION

Synthesis of sedimentary structures and their directional properties remains the most diagnostic criterion in interpreting depositional environments from facies analyses. This sedimentologic synthesis is particularly useful in fluvial-system deposits in which success or failure in determining facies and environments hinges on proper identification of sedimentary structures. That is, fluvial systems being clastic dispersal systems produce interrelated sets of sedimentary structures than can be used to reconstruct original conditions of sedimentation. The fluvial systems consist of alluvial, deltaic, and barrier-coastline dispersal systems. In the alluvial and deltaic systems, dispersal is chiefly along channels; whereas, in the barrier-coastline system, dispersal is mainly along foreshore and shoreface environments (see fig. 1, chap. I). These dispersal "conduits" also serve as major depositional areas. Mechanisms of dispersal vary from river-dominated in the alluvial system, river-, tide-, and wave-influenced in the deltaic system, to wave- and tide-dominated in the barrier-coastline system. Although these primary sediment dispersal areas are peripheral to peat-forming swamps, their current-formed deposits can serve as a key to recognition of depositional systems. Perhaps the most diagnostic detrital deposits for recognizing conditions of sedimentation in peat-forming floodbasins, which are peripheral to the primary dispersal areas, are those that accumulated alongside the swamps. These detrital deposits formed in dispersal subsystems that originated in the primary dispersal systems (e.g., channel, barrier). These dispersal subsystems include crevasse splay and crevasse delta in alluvial and deltaic systems, and washover, tidal inlet, flood-tidal delta, and bayhead delta in barrier-coastline system. Like detrital deposits of the primary dispersal systems, the deposits of these dispersal subsystems are also current-formed with imprints of flow conditions of transporting and depositing media.

This chapter strives to concentrate on physical description of current-produced sedimentary structures (see table 1, chap. I) of coal-associated facies deposited by the primary dispersal systems and subsystems. In addition, processes of deposition of the coal-associated facies will be discussed. The sedimentary structures and their sequences are the most easily recognized features of coal-associated facies and their field observation provides optimum benefit to geologists in understanding depositional environments related to coal sequences.

A large volume of detrital sediments is dispersed through the fluvial systems from drainage basins to depositional sinks. Within this depositional continuum, detritus are dispersed mainly intermittently along networks of channels in alluvial fans, alluvial plains, and delta plains. Continued coastwise dispersal of detritus ensues from the terminus of delta-plain channel networks forming delta fronts and barrier coastlines.

ALLUVIAL FAN

Alluvial fans are sediment dispersal systems primarily located in the headwaters or drainage basins of the fluvial system (see fig. 1, chap. I). Because of their relatively steep regional slope, highly variable discharge, and abundant coarse bedload sediments, alluvial fans tend to form low-sinuosity, multiple, diverging-converging channels called braided streams--a term also used for certain types of high-sinuosity, multiple-channel, bedload rivers. Alluvial fans tend to form when streams in confined mountain valleys emerge into an alluvial plain where the stream flow diffuses laterally, separating into more than one channel and resulting in deposition of coarse bedload sediments due to loss of competency. Thus, an alluvial-fan deposit may be recognized commonly by the presence of coarse detritus deposited near basin margins (fig. 1).

Miall (1977, 1978) defined six facies types of low sinuosity, multiple-channel, braided streams:

1. Trollhein facies type of proximal braided streams, which is predominantly in alluvial fans, is subject to debris flows.
2. Scott facies type of proximal braided streams, which includes alluvial fans that are characterized by stream flows.
3. Donjek facies type of distal gravelly braided streams.
4. South Saskatchewan facies type of sandy braided streams (cyclic).
5. Platte facies type of sandy braided streams (noncyclic).
6. Bijou Creek facies type of ephemeral braided streams subject to flash floods.

Although the Trollheim and Scott facies types are exemplified by proximal alluvial fans, the Scott, Donjeck, and South Saskatchewan facies types may occur in a gradational proximal-distal relationship on the same alluvial fan system. Miall (1980) suggested that the following numerical limits of gravel content be used to distinguish the facies types: Scott facies type >90 percent, Donjek facies type 10-90 percent, South Saskatchewan facies type <10 percent. The Donjek, South Saskatchewan, Platte, and Bijou Creek facies types may also occur in proximal reaches of high sinuosity, braided streams.

Trollheim Facies Type

This gravel-dominated alluvial fan deposit is named after the deposits of the Trollhein fan in arid or semiarid regions of California (Hooke, 1967). The vertical section, showing sequences of rock types and internal structures, is illustrated in figure 2A. The facies type corresponds to the proximal reaches of the alluvial-fan wedge shown in figure 1. This alluvial fan facies, which is dominated by debris-flow process, consists of poorly sorted, matrix-supported gravels with well defined flat, abrupt bases. These gravelly deposits are interbedded with finer-grained clast-supported gravels of streamflows representing incision of the fan. This facies contains minor amounts of crossbedded pebbly sands, silts, and muds occurring in the upper parts of fining-upward sequences.

Scott Facies Type

This gravel-dominated (>90 percent gravel) alluvial fan facies is named after the deposits of the Scott glacial outwash fan in southeastern Alaska (Boothroyd and Ashley, 1975). The vertical profile of this facies type, showing sequences of rock types and internal structures, is depicted in figure 2B. This facies type corresponds to the mid-portion of the alluvial fan model (fig. 1). It is found beyond the limit of debris flow or represents alluvial fan deposits that lack the influence of debris flow. It also occurs in proximal braided river deposits. The facies type consists chiefly of

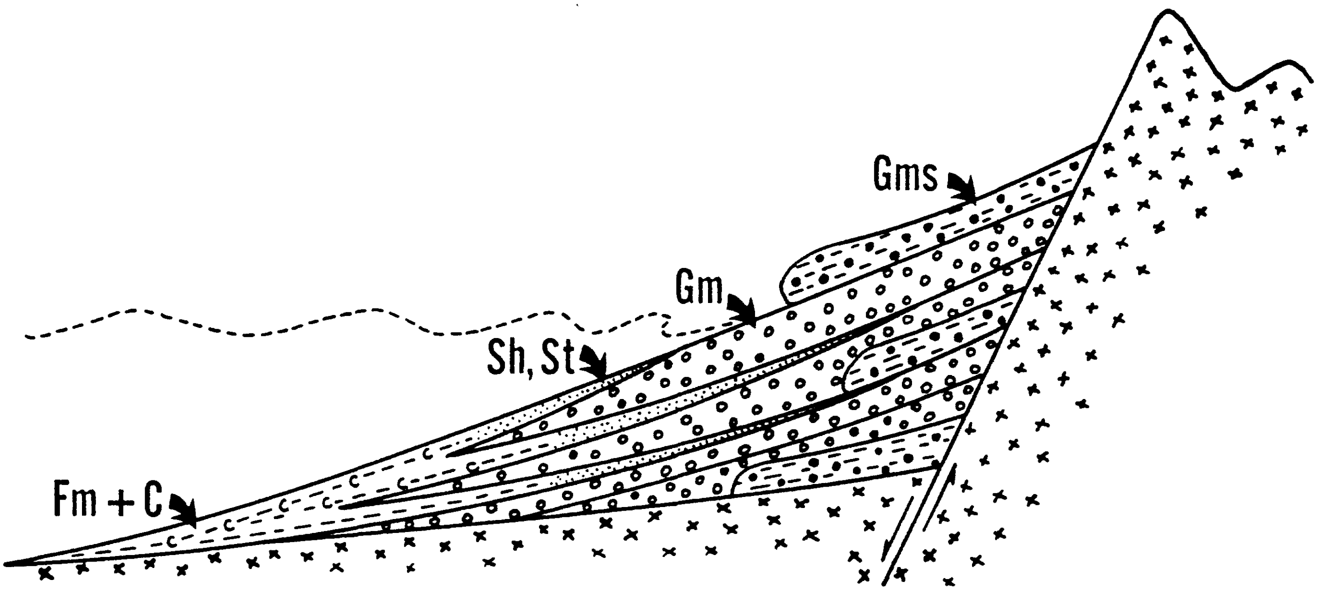

Figure 1.--An idealized longitudinal cross section and deposits of an alluvial fan formed along a basin margin. Gms = matrix-supported gravel; Gm = imbricate, framework gravel; St = trough crossbedded sandstone; Sh = horizontally stratified sandstone; Fm = massive sandy mud; C = coaly and carbonaceous mud. Adapted from Miall (1980).

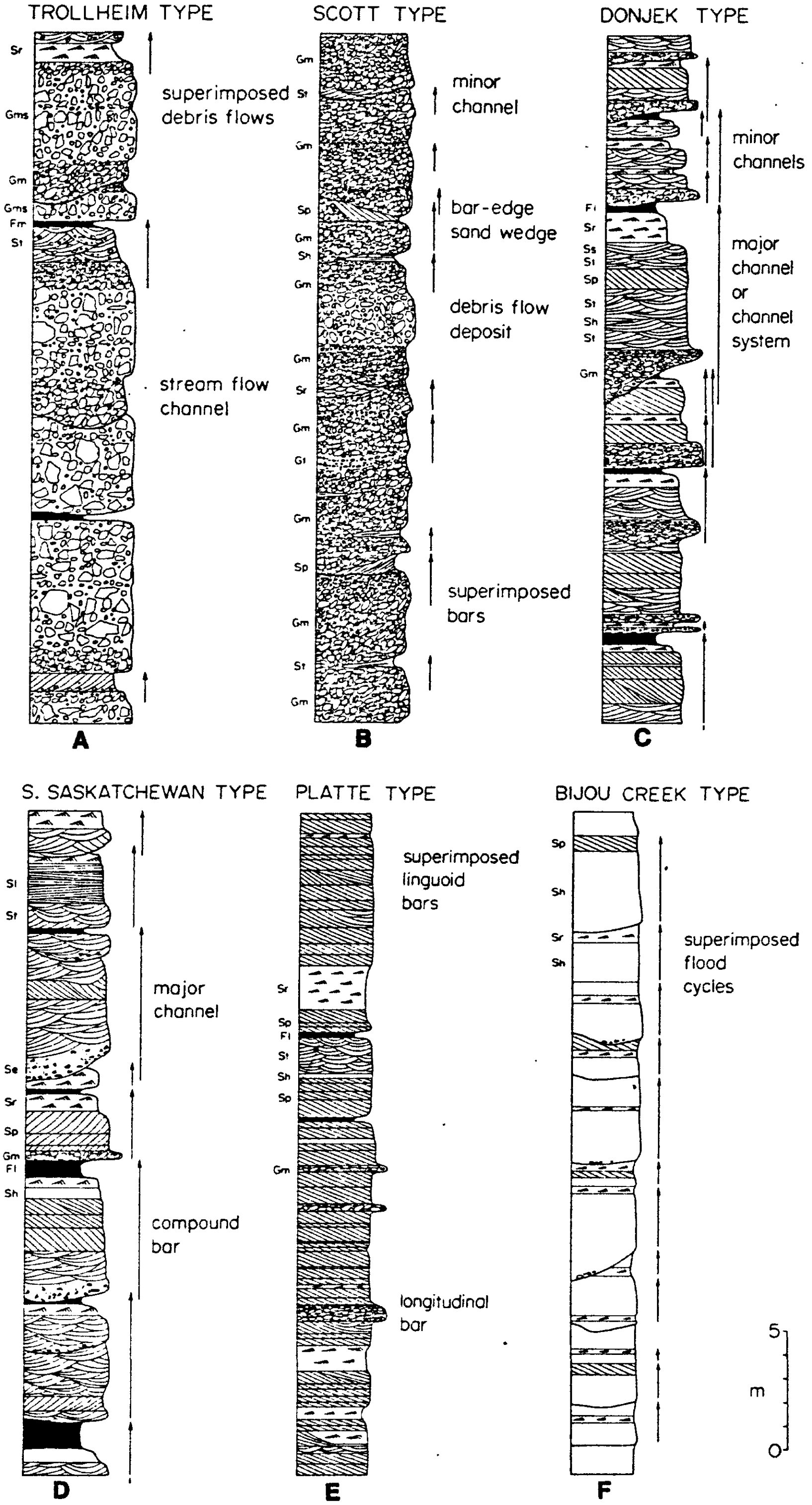

Figure 2.--Facies types of braided depositional profiles include Trollheim type, Scott type, Donjek type, South Saskatchewan type, Platte type, and Bijou Creek type. Symbols next to these facies are as follows: Gm = massive grain-supported gravel; Gms = massive matrix-supported gravel; ST = trough crossbedded sand; Sp = planar crossbedded gravel; Sr = rippled sand; Fm = mud and/or silt; Ss = sand with broad shallow scours; Fl = sand, silt, and mud; Sl = low-angle crossbedded sand (From Miall, 1978).

multistory, clast-supported gravel deposits of superimposed longitudinal bars and small-scale, gravel-sand cycles of low water episodes as in the waning stages of flood events. The sand units are deposited in abandoned channels or as microdelta wedges prograding from gravel bars as water level drops.

Donjek Facies Type

This facies type (fig. 2C) is named after the deposits of the Donjek River in Alaska (Rust, 1972). It is characterized by bedload sediments consisting of abundant gravel and sand. The gravels occur in deeper channels that may form longitudinal and transverse bars. Elevated, inactive channel reaches receive sand and gravel during flood stages. The highest part of the system may be covered by vegetation that traps the fine sediments from floodwater. Dominant stratifications include clast-supported gravels and trough and planar crossbedded sands. The clast-supported gravels are imbricated and are crudely horizontally (parallel) bedded. Ripple laminations in sand and parallel laminations in sand, silt, and mud are minor stratification types.

South Saskatchewan Facies Type

This facies type is named after deposits of the South Saskatchewan River, Alberta, Canada (Cant and Walker, 1978). The sequence of internal structures of this sand-dominated facies is shown in a vertical profile in figure 2D. Dominant bedforms are dunes and sandwaves that are translated into trough crossbeds as internal structures. Minor internal structures are planar crossbeds, ripple lamination, crudely bedded lag gravel and pebbly sand, and parallel laminated sands, silts, and muds.

Platte Facies Type

This facies type is named after deposits of the Platte River, Colorado and Nebraska (Smith, 1970). The vertical sequence of internal structures and rock types is shown in figure 2E. The facies is a variant of the South Saskatchewan facies type. The facies is deposited in a broad, shallow-braided river. Dominant bedforms are linguoid or transverse foreset bars and sandwaves, both translated in internal structures as trough and planar crossbeds, respectively. Planar crossbeds are usually more common than trough crossbeds. Minor types of stratification include parallel and ripple laminations.

Bijou Creek Facies Type

This facies type is named after deposits of Bijou Creek, eastern Colorado (McKee and others, 1967). The vertical section and the sequence of internal structures are shown in figure 2F. The facies is characteristic of sandy ephemeral braided streams in which deposition occurs only during violent flash floods. This results in dominantly parallel-laminated sands with minor planar crossbeds, and ripple laminations.

Coals In Alluvial Fan Settings

Alluvial fan facies and their association with coal deposits and/or coal-bearing deposits can be summarized in the following settings:

1. Intermontane basin setting where transverse alluvial fans emerge into and amalgamate with an alluvial plain that consists of basin axis or longitudinal channels (Hayes and others, 1973; Hite, 1976; Obernyer, 1978; Flores and Hanley, 1983).
2. Intermontane lacustrine setting where mountain front alluvial fans drain into a large lake that forms a temporary base level (Heward, 1978).
3. Piedmont-coastal setting where piedmont alluvial fans drain into coastal plain and marine water (Pedlow, 1979).
4. Glacial setting where outwash alluvial fans drain into coastal plain and marine water (Flores, 1982).

ALLUVIAL PLAIN

In the fluvial-system depositional continuum, the alluvial plain includes the system downslope to alluvial fans and upslope to deltas (see fig. 1, chap. I). The alluvial plain (fig. 3) can be divided morphologically and facies-wise into:

1. A belt of channel complexes.
2. Floodplain.

Channel Facies

Characterization of the alluvial plain depends largely on the patterns of the channel complex. Allen (1965) and Schumm (1971) recognized four channel patterns that developed in alluvial plains (fig. 4):

1. Braided channel.
2. Anastomosed channel.
3. Straight channel.
4. Meandering channel.

The most important parameters that control the patterns of the channel are sinuosity and braiding (Rust, 1978). Sinuosity is a ratio of the thalweg (line of the deepest channel) to valley length (straight line distance in that part of the valley over which the thalweg length is measured). High sinuosity is characterized by >1.5 ratio; whereas, low sinuosity is described by <1.5 ratio. Braiding measures the number of bars per meander wavelength of the channel; thus, this parameter defines the channel multiplicity. Single channel is defined by a braiding parameter of <1 and multiple channel is defined by a braiding parameter of >1. On the basis of the sinuosity and braiding parameters, the braided pattern is a low sinuosity, multiple channels; the anastomosed pattern is a high sinuosity, multiple channels; the straight pattern is a low sinuosity, single channel; and the meandering pattern is a high sinuosity, single channel. Because of the ambiguity in the use of "braided" and "anastomosed" patterns in which many workers use them as synonymous, Schumm (1971) suggested that "braided" channel pattern should refer to rivers in which flow diverges and rejoins around bars; whereas, "anastomosed" channel pattern should be used where the river divides into more than one permanent subchannel separated from each other by a floodplain. Each anastomosed channel displays its own pattern of braiding and sinuosity. Figure 5 shows a modern example of an anastomosed river.

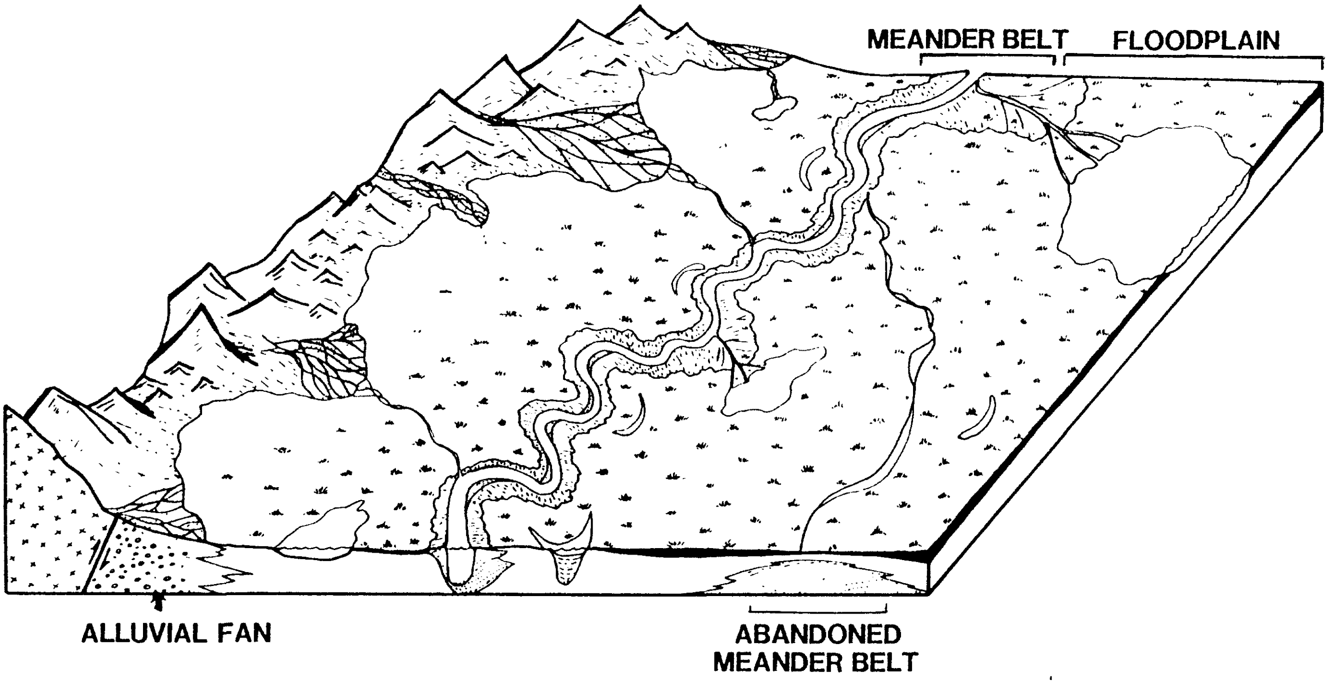

Figure 3.--Geomorphological subdivisions of the alluvial plain into channel complex (meander belt) and floodplain. Alluvial fans along the basin margin disperse sediments from a surrounding source area into the alluvial plain.

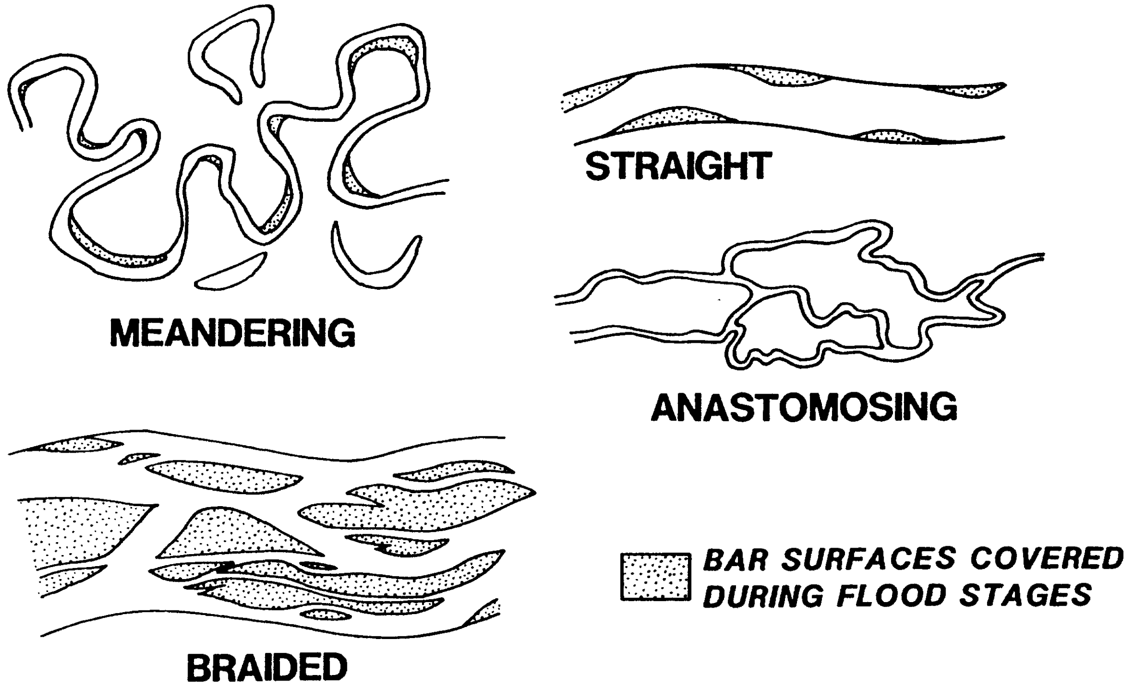

Figure 4.--Types of channel patterns or river types. Adapted from Miall (1980).

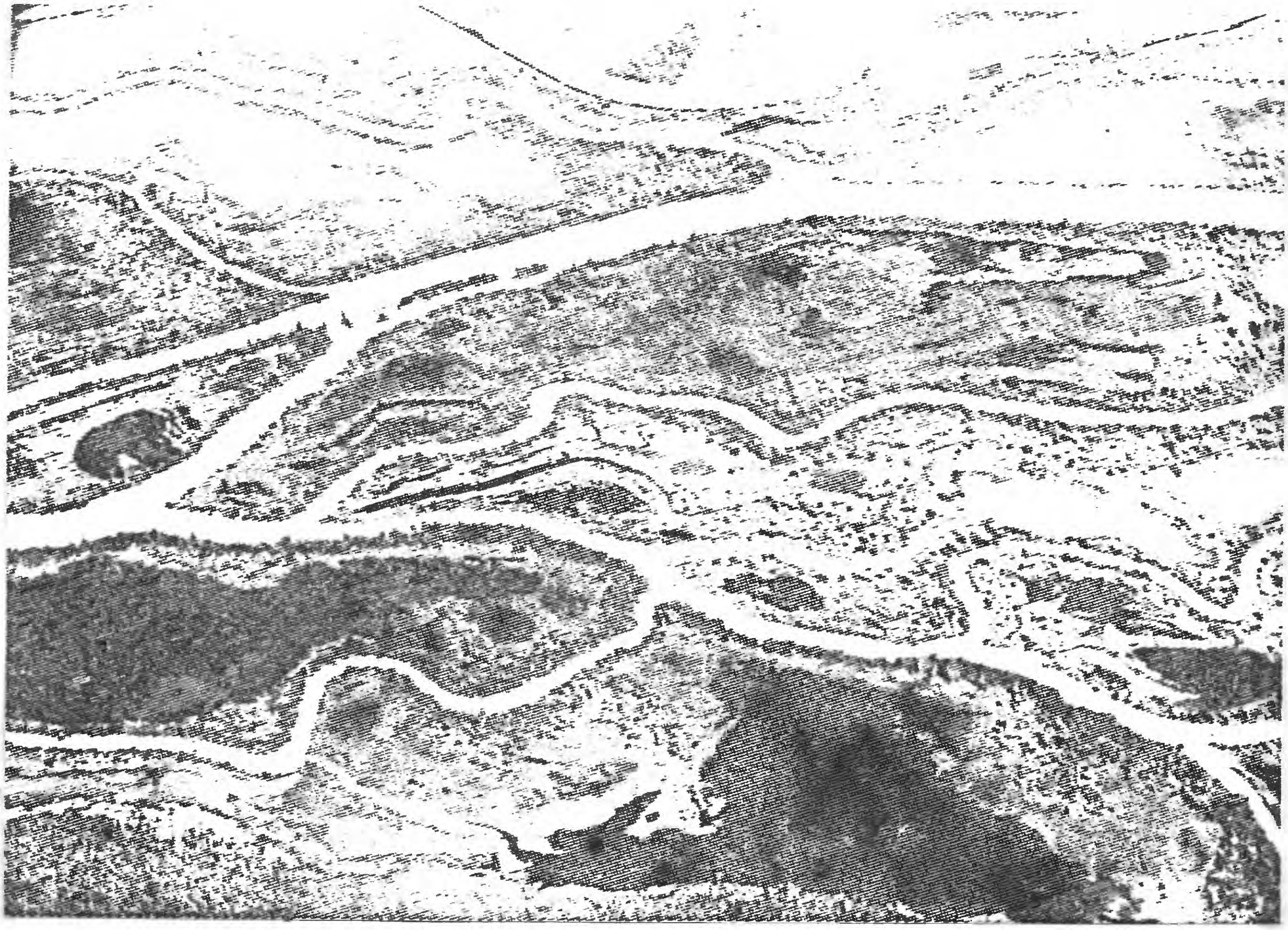

Figure 5.--Anastomosed channel pattern of the Upper Columbia River, Alberta, Canada. Photo courtesy of D.G. Smith, University of Calgary.

Schumm (1960, 1963) proposed a method of classifying channel patterns based on their sediment load. That is, meandering and anastomosed channel patterns consist of mixed load or suspended load; whereas, braided channel pattern contains bedload. Miall (1977) and Jackson (1978), however, pointed out some exemptions to this classification in which a high sinuosity or meandering channel pattern can also contain coarse bedload sediments.

In any single alluvial plain system, the braided or bedload channel is developed distal to alluvial fans and/or at the headwaters of the alluvial plain. The braided channel passes downstream into meandering or mixed load and suspended load channels. However, depending on the tectonic condition, which is usually a subsiding basin (Smith and Putnam, 1980), the braided channel may pass into an anastomosed or mixed load and suspended load channel. Such channel patterns constitute a spectrum of river types compared to those that are gradational with pebbly braided rivers to those downstream that are sandy, gently meandering rivers. These downstream changes in river types are often accompanied by a change in grain size and development of abundant longitudinal bar and transverse bar bedforms. Many sandy, gently meandering rivers show aspects of braided streams such as the development of mid-channel bars, while others show meandering of thalweg with the development of alternating, bank-attached bars. The low sinuosity of these rivers is probably caused by low levels of suspended load which suggests that the channels are rarely cut off. That is, under this condition a development of only a few clay plugs limits channel migration; otherwise, the channel freely sweeps across the floodplain in a continuous manner. The low sinuosity or gently meandering rivers are gradational into high sinuosity or highly meandering rivers farther downstream. The high sinuosity is influenced by low gradients and high suspended load (Leopold and Wolman, 1957; Schumm and Kahn, 1972). Under this condition, the position of meanders is stabilized by clay plugs propagated by channel cutoffs (chute and neck cutoffs) within a belt. This meander belt, which receives concentrated channel and overbank sediments, is built up as a "ridge" above the floodplain. The differential elevation between the meander belt and floodplain creates unstability that is relieved by breaching of overbank deposits or capturing a crevasse-splay channel during floods. The shift of the meander belt, which is known as the process of "avulsion," is accompanied by abandonment of the old meander belt. This abandonment process, which includes a long segment of the channel complex, is different from the abandonment of a meander bend that fills up with fine sediments.

The highly meandering rivers, which occur in the lower reaches of the alluvial plain, merge downstream with straight distributary channels of delta plains. These meandering rivers commonly display more regular and continuous discharge along a single channel that overflows during floods than do braided rivers. Thus, deposits of both meandering and braided rivers differ in characteristics as shown in table 1. Perhaps the most important deposit of meandering rivers is the point bar that is enclosed by the meander loop (fig. 6). Deposition of a point bar is at the convex side of the meander bend and it is accompanied by erosion at the "cutbank" or at the concave side of the meander bend. These simultaneous processes cause lateral migration of the point bar whose vertical grain size and bedform distributions are controlled by the helicoidal flow pattern that enters the meander bend (Allen, 1970; Bridge, 1975). The point bar surface, which is horizontal at about river

Table 1.--Difference between braided and meandering river deposits

[Modified from Ethridge, 1982]

Criterion	Braided rivers	Meandering rivers
1. Lateral accretion deposits	Linguoid bars, low water bar accretion, rare point bars	Fining-upward point bars (including epsilon cross-beds)
2. Vertical accretion deposits	Channel floor bedforms, sheet flood deposits, bar-top deposits, minor overbank deposits	Overbank deposits, channel-lag deposits, abandoned channel fill deposits
3. Type of scour surface	Channel erosion	Meander widening
4. Channel abandonment behavior	Progressive, as a result of aggradational fill	Sudden, as a result of meander chute or neck cutoff or course abandonment
5. Channel abandonment deposit	Fining-upward cycle	Fine-grained fill
6. Lateral relation of channel sand bodies	Wide, multilateral	Moderately wide, multilateral to narrow multistory
7. Stratification types		
a. Matrix-supported gravel	Rare to common	Absent
b. Clast-supported gravel (crude horizontal bedding and clast imbrication)	Common (longitudinal bar deposit)	Rare to common (generally as a thin lag deposit)
c. Trough to planar cross-bedded gravel	Rare to common	Absent
d. Trough cross-bedded sand	Common	Common
e. Planar cross-bedded sand	Common	Generally rare
f. Parallel laminated and rippled sand	Common	Common
g. Scour fill sand	Rare to common	Absent
h. Laminated sand, silt and mud and mud or silt drape; with bioturbation, roots, desiccation cracks, caliche nodules, etc.	Rare to common	Common

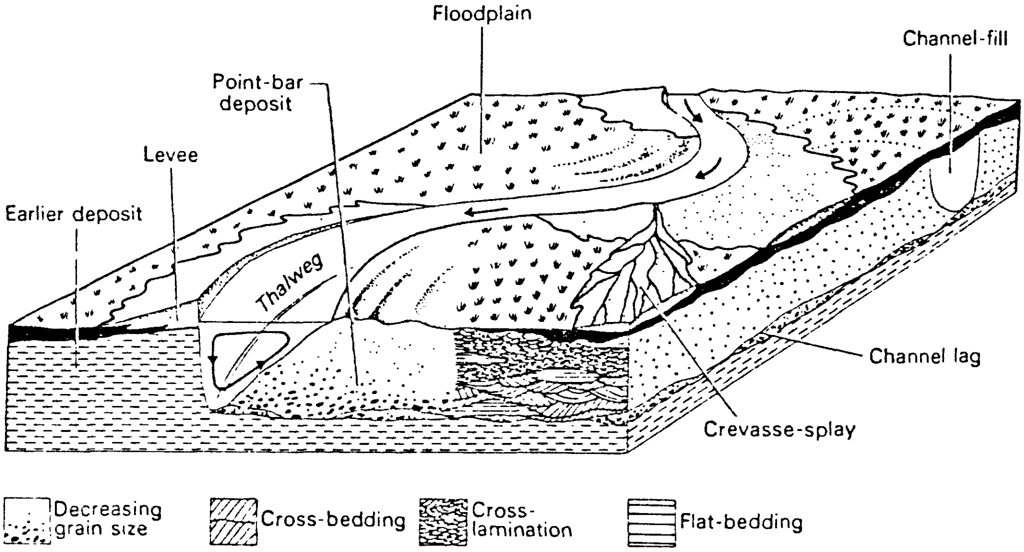

Figure 6.--Point-bar sequence, sedimentary structures, and water flow pattern at the inside of a meander bend. Floodplain facies (crevasse-splay, swamp, and levee deposits) is also illustrated. Adapted from Allen (1970).

level and slopes toward the thalweg, is a site of deposition where grain size distribution and bedforms develop. An idealized model of a point bar deposit consists of a fining-upward sequence that includes dune or sandwave bedforms in the deeper part of the channel and ripple and parallel bedforms in the shallower area (Sundborg, 1956; Frazier and Osanik, 1961; Bernard and others, 1962). Thus, the vertical sequence of internal structures of a generalized ancient point bar deposit from base to top consists of channel lag, large-scale trough cross lamination, parallel lamination, and ripple (current and climbing) lamination. Planar cross lamination in a few cases (Sundborg, 1956; McGowen and Garner, 1970; Jackson, 1975) is associated with the more common trough cross lamination. Because of variation in discharge, the upper (horizontal) surface of the point bar is capped by "clay drapes," which are deposits of falling flood stages. The lateral accretion surface (inclined towards the thalweg) is usually preserved as a large-scale, low-angle crossbed termed epsilon crossbed by Allen (1963). The dip of the epsilon crossbed varies from 1 for large rivers to 25 for small rivers. A couplet of this epsilon crossbed binds the trough crossbeds. Since the thickness of this crossbedded couplet may equal channel depth, it is suggested that the epsilon crossbed is a good indicator of the scale of the original river (Leeder, 1973). Figures 7, 8, and 9 show examples of ancient point bar deposits and their internal structures as well as the associated "clay plug" deposit.

Floodplain Facies

The floodplain occupies a much larger area in the alluvial plain than the channel complex. Floodplain facies plays the most important role in understanding the overall depositional environment of coals. Thus, associated detrital deposits, which are deposited by secondary disperal subsystems originating from the channel complex, are diagnostic indicators of the conditions of sedimentation of coals. The floodplain can be divided geomorphologically into levee and floodplain proper.

Levee.--This is a low-lying ridge that flanks the channel and is a result of overbank sedimentation during bankfull floods. During floods, fallout of coarse suspended sediments in the form of sands and silts occurs immediately along the channel while muds are deposited farther into the floodplain proper. These variable grain-size sediments are interbedded with each other, reflecting the fluctuating turbulence of floodwaters. Vegetation along the ridge serves as local barriers to floodwaters. The internal structures of levee deposits are ripple (current and climbing) lamination and parallel lamination locally disturbed by roots and trunks. Ironstone nodules may be common as a result of oxidation of organic matter. Figure 10 shows a sequence of sedimentary structures and lithology of an ancient levee deposit.

Floodplain Proper.--This area occupies the distal part of the floodplain away from the levee and channel complex. It consists of crevasse splays, swamps, and lakes. The crevasse splays are fan-shaped detrital deposits introduced into the floodplain by breaching of levee deposits by floodwaters. Sands are deposited proximal to the fan, and silts and mud distal to the fan. The crevasse splay builds out towards the floodplain proper imparting a coarsening-upward characteristic to the deposits, which from base to top consists of muds, silts, and sands. Crevasse splays are fed with sediments during recurrent floodings by the levee channel that bifurcates into organized subchannels as they expand areally into the floodplain. Since

Figure 7.--Ancient point-bar deposits (PB) in the Gallup Sandstone, Gallup sag, New Mexico.

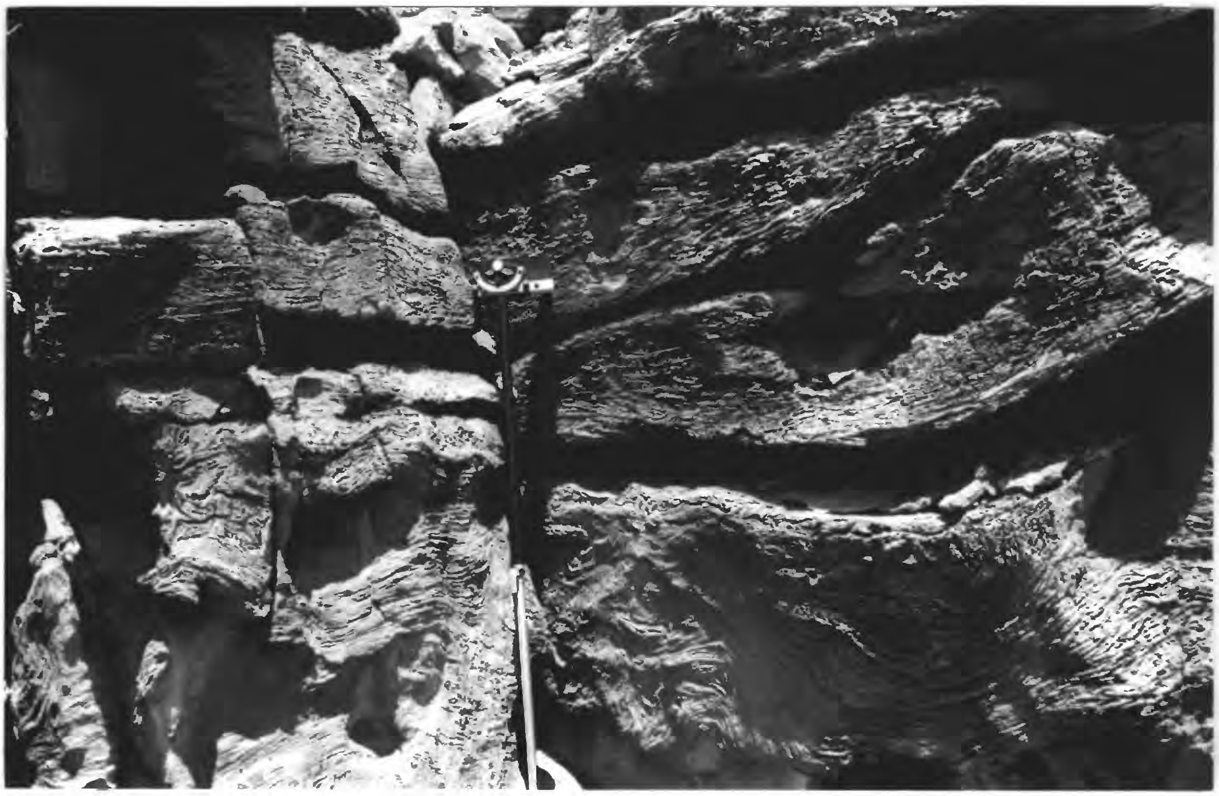

Figure 8.--A vertical sequence of internal structures of an ancient point-bar deposit. Trough cross lamination and convolute lamination.

Figure 9.--Ancient "clay plug" deposit of shale, siltstone, and carbonaceous shale, which may represent an abandoned channel deposit (above arrows) that resulted from a neck cutoff.

Figure 10.--Interbedded shale, siltstone, and sandstone of an ancient levee deposit. It displays ripple lamination and tree-trunk molds in the sandstone.

crevasse-splay sediments are mainly deposited during flood discharges through the feeder channels, rapid sedimentation by shooting (unidirectional) flow is reflected by abundant ripple (asymmetrical and climbing) lamination and subordinate trough crossbed and convolute lamination in the sands. The underlying muds and silts display vertical and horizontal burrows and lenticular bedding. Crevasse splays occur areally either as an isolated body or, as in the Brahmaputra (Coleman, 1969), a locally merged body forming an almost continuous sheet that is exemplified by tabular sands oriented parallel or subparallel to the channel complex. Furthermore, coalescing of crevasse splays produces a stack of coarsening-upward sequences. As long as feeder channels are open and continue to serve as conduits for floodwater discharges, crevasse splays will prograde into the distal floodplain. Here, they overrun peat-forming swamps as well as debouching into lakes. In the latter, the crevasse splay forms a crevasse delta or lake splay that consists of interbedded carbonates, muds, silts, and sands with enclosed freshwater faunas. The sands are ripple laminated, convolute laminated, and burrowed. Synsedimentary deformation structures are also common (Ethridge and others, 1981). The splays continue to fill the lakes until such time that the feeder channels become "healed" in which case the crevasse splay is abandoned. The abandoned crevasse splay is gradually invaded by vegetal growth from outlying, low-lying, poorly drained swamps. This elevated platform, which initially forms a well-drained swamp, may finally develop into a poorly drained swamp (Coleman, 1966). Examples of ancient alluvial-plain crevasse splay, lake splay, and swamp deposits, as well as their sedimentary structures, are shown in figures 11, 12, 13, and 14.

DELTA PLAIN

Deltas are the final resting places for most sediments that are dispersed by rivers of the alluvial plain as they enter oceans, semienclosed seas, lakes, or barrier-sheltered lagoons. A delta consists of a low-lying delta plain and a seaward-dipping delta front. The delta plain is the most important component of the delta since it is a coal-forming environment. The modern and ancient delta deposits are the most studied of the various deposits of the fluvial depositional systems. Perhaps, the works of Gilbert (1885, 1890) of the Pleistocene Lake Bonneville deltas and Barrell (1912, 1914) of the Upper Devonian deltas started the quest for understanding the processes and facies of deltas. However, an understanding of ancient deltaic deposits is not complete without their comparison to modern deltas, which was initially performed by Johnston (1921, 1922) in the Fraser River delta and later surpassed by classical work in the Mississippi River delta by Russell (1936), Fisk (1944), and Coleman and Gagliano (1964). Since these pioneering works, extensive publications on modern and ancient deltaic deposits and processes have appeared in the literature as stimulated by their economic importance (Bates, 1953; Allen, 1965; Van Andel, 1967; Fisher and others, 1969; Morgan, 1970; Wright and Coleman, 1973; Coleman and Wright, 1975; Galloway, 1975; Miall, 1980). Significant amounts of oil, gas, and coal resources are found in deltaic deposits in terrigenous basins of the world. The hydrocarbons are found mainly in the "skeletal framework" facies (e.g. distributary channel, delta-front sandstones), and coals are associated mainly with the "flesh nonframework" or interchannel facies. The characteristics of the framework and nonframework facies, which differ from one delta model to another, are controlled by several factors:

Figure 11.--A sequence of crevasse-splay (CS) and swamp (S) deposits in the Tongue River Member of the Fort Union Formation, Powder River Basin, Wyoming and Montana.

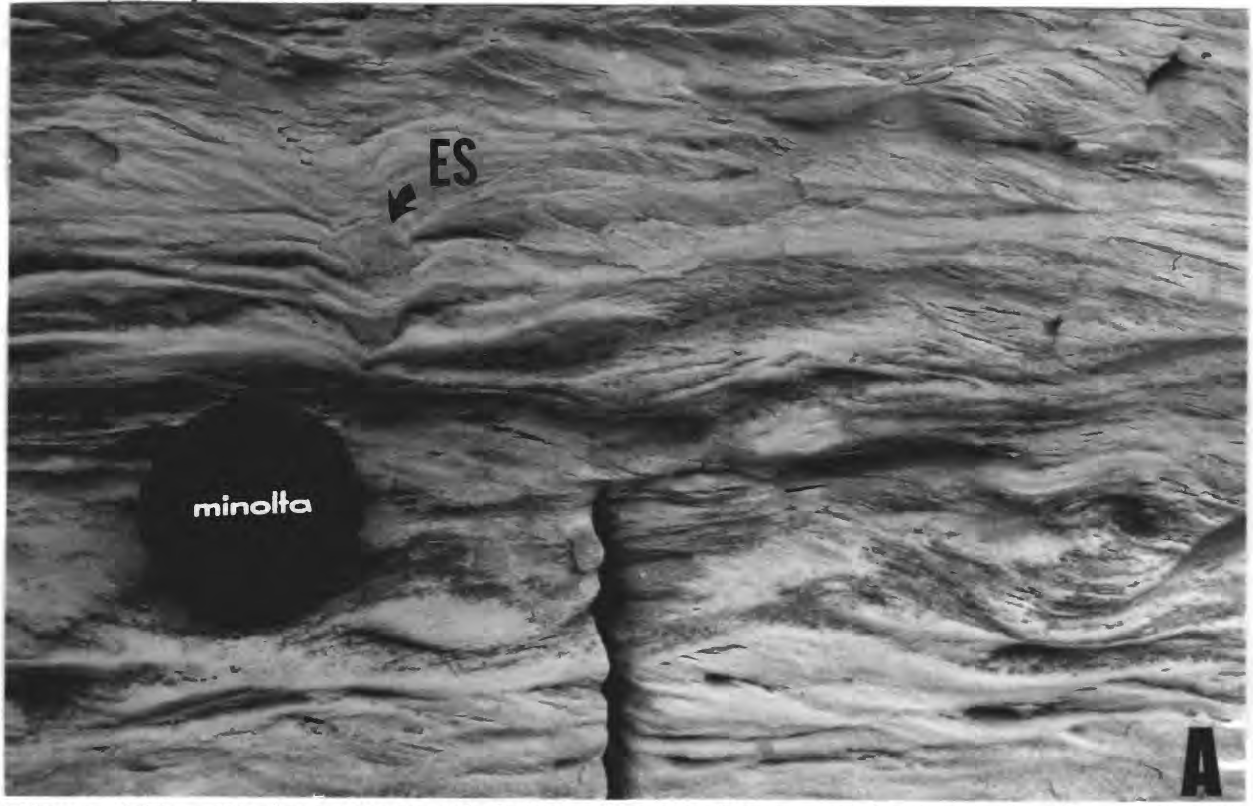

Figure 12A.--Various types of internal structures and burrows in the crevasse-splay sandstones shown in figure 11. A. Ripple laminations and a wide escape structure (ES) of a pelecypod.

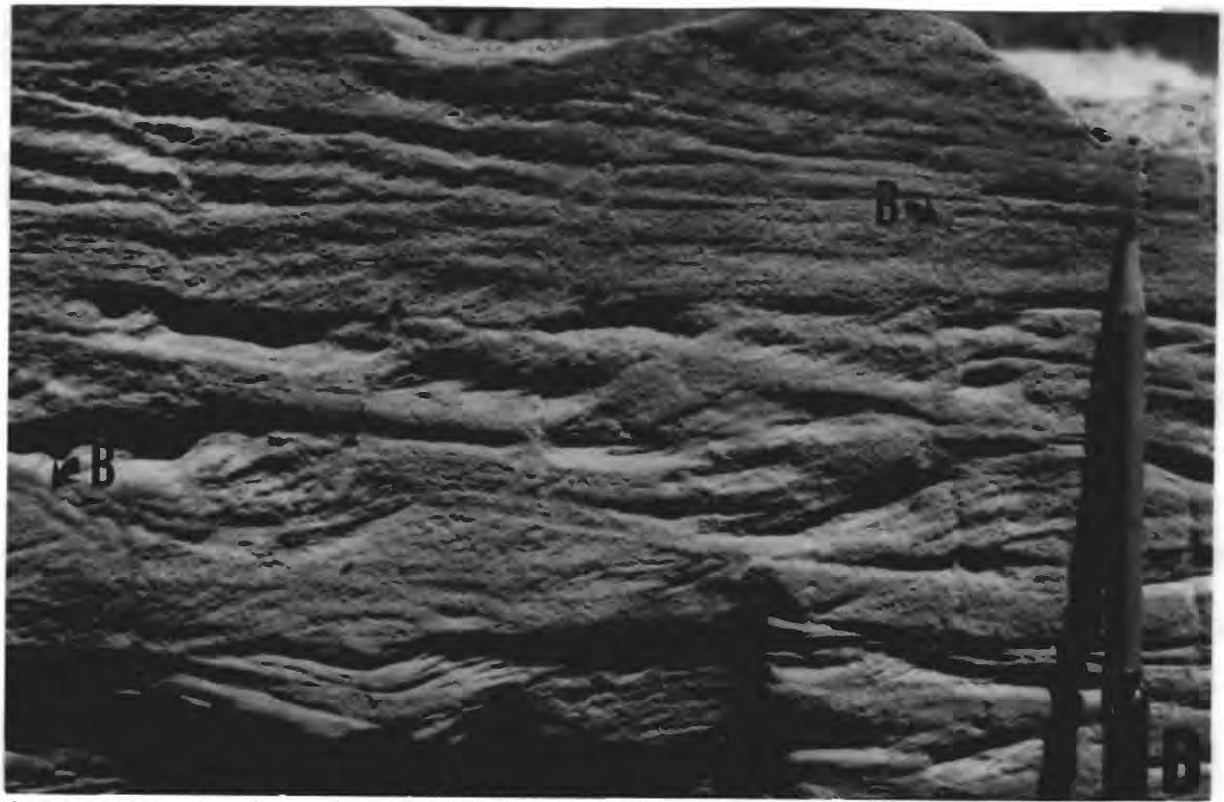

Figure 12B.--Various types of internal structures and burrows in the crevasse-splay sandstones shown in figure 11. B. Ripple lamination and narrow, straight to curvilinear burrows (B).

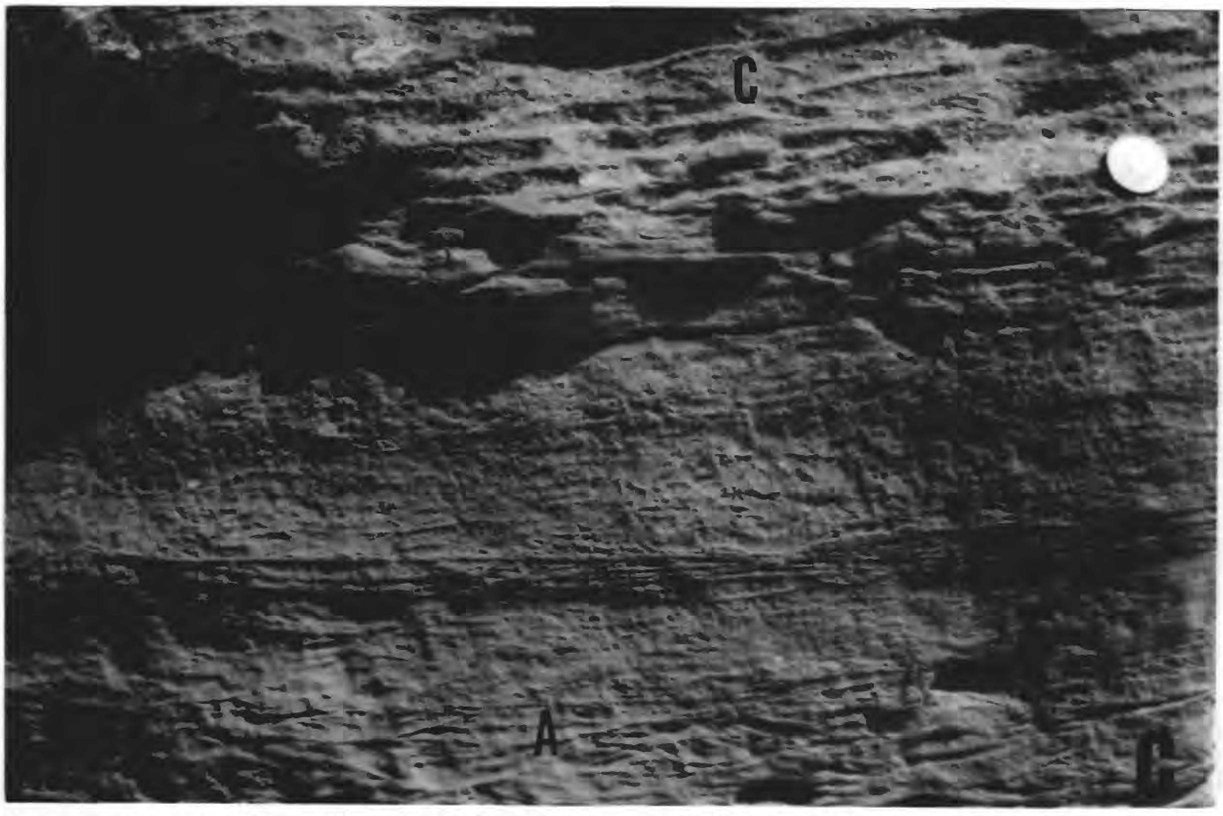

Figure 12C.--Various types of internal structures and burrows in the crevasse-splay sandstones shown in figure 11. C. Climbing or ripple drift (C) and asymmetrical ripple (A) laminations.

Figure 12D.--Various types of internal structures and burrows in the crevasse-splay sandstones shown in figure 11. D. Convolute laminations and a possible gas heave (G) structure.

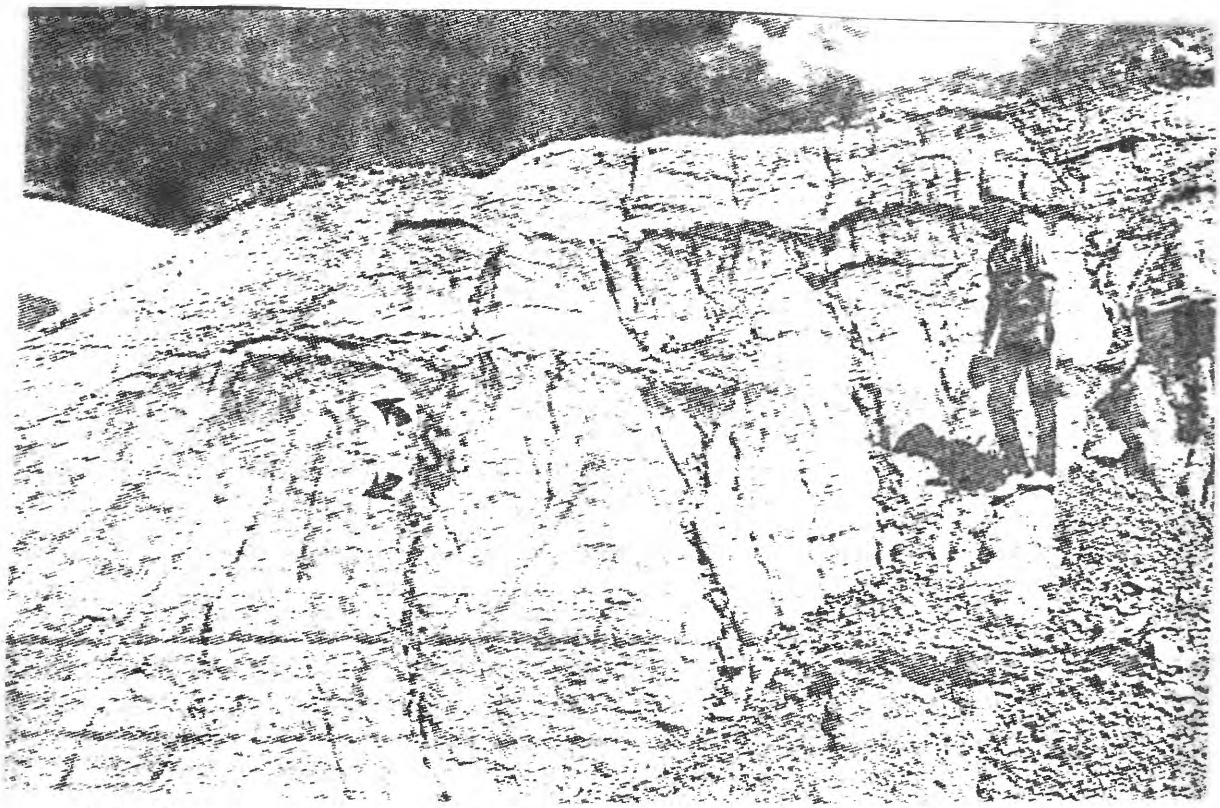

Figure 13.--Lake-splay deposit with synsedimentary deformation structures (S) in the Tongue River Member of the Fort Union Formation, Powder River Basin, Wyoming and Montana.

Figure 14.--Convolute lamination and diapiric structure in the lake-splay sandstone shown in figure 13.

1. Volume and rates of sediment input as well as the ratio of bedload/suspended load.
2. Water velocities, discharges, and densities for both rivers and basins.
3. Basin energy regimes, such as wave and tidal energy, as well as longshore currents.
4. Basin tectonic condition (e.g. subsidence, shape, and size).
5. Climate.

The relationship between the framework and nonframework facies are best demonstrated by a delta plain, which consists of active and inactive (abandoned) distributary channels separated by water-submerged and emergent interchannel areas. The interchannel areas, which consist of bays, lakes, tidal flats, marshes, swamps, and salinas, are sensitive to climate (Elliott, 1978a). In humid and subhumid climates, dense vegetation occurs over a large wetland of the interchannel areas in contrast to arid and semiarid climates in which the interchannel areas support sparse vegetation and accumulate salt deposits. Most delta plains are dominantly affected by fluvial or tidal process and rarely influenced by wave process. Thus, types of delta plains and their facies can be categorized according to the degree of effects of these processes following the similar schemes of classification of deltas and their delta plains (alluvial-dominated, tidal-dominated, and wave-dominated) presented by Galloway (1975) and Elliott (1978a).

Alluvial-Dominated Delta Plain Facies

This type of delta plain and its facies are exemplified by the Mississippi River delta plain (fig. 15). It is a product of active progradation and aggradation typical of a constructive delta system (Fisher and others, 1969). The associated distributary channels are characterized by unidirectional flow in commonly low sinuosity (straight to nearly straight) channels, although high sinuosity (meandering) channels develop in moderately high sand (bedload) systems, particularly in the upper delta plain. Braiding and anastomosing channels also occur, although rarely, in systems with a high proportion of bedload and suspended load. Facies and sequences of internal structures of distributary channels of alluvial-dominated delta plains resemble, to a large extent, those of the channel complex of the alluvial plain. However, the high frequency of bifurcation and channel "avulsion" or switching, which creates abundant abandoned distributary channel deposits, attests to the active progradation and aggradation of the system. In some alluvial-dominated delta plains, distal parts of abandoned distributary channel deposits are reworked by waves and sealed by beach deposits upon abandonment (e.g. Rhone and Ebro deltas; Kruit, 1955; Maldonado, 1975). The distributary channel deposits show an overall fining-upward sequence of muds, silts, and sands resulting from either channel abandonment or lateral migration. The deposits consist of basal lag grading upward into trough crossbeds, convolute lamination, and ripple lamination. Wave reworking of distal parts of these deposits results in well-sorted, parallel-laminated deposits with enclosed marine fauna. Like their alluvial-plain channel counterparts, distributary channels are flanked by rooted, rippled levee deposits consisting of interbedded muds, silts, and sands. Figures 16, 17, and 18 display a distributary channel sandstone, an abandoned distributary channel, and a sequence of internal structures in a distributary channel sandstone in ancient rocks.

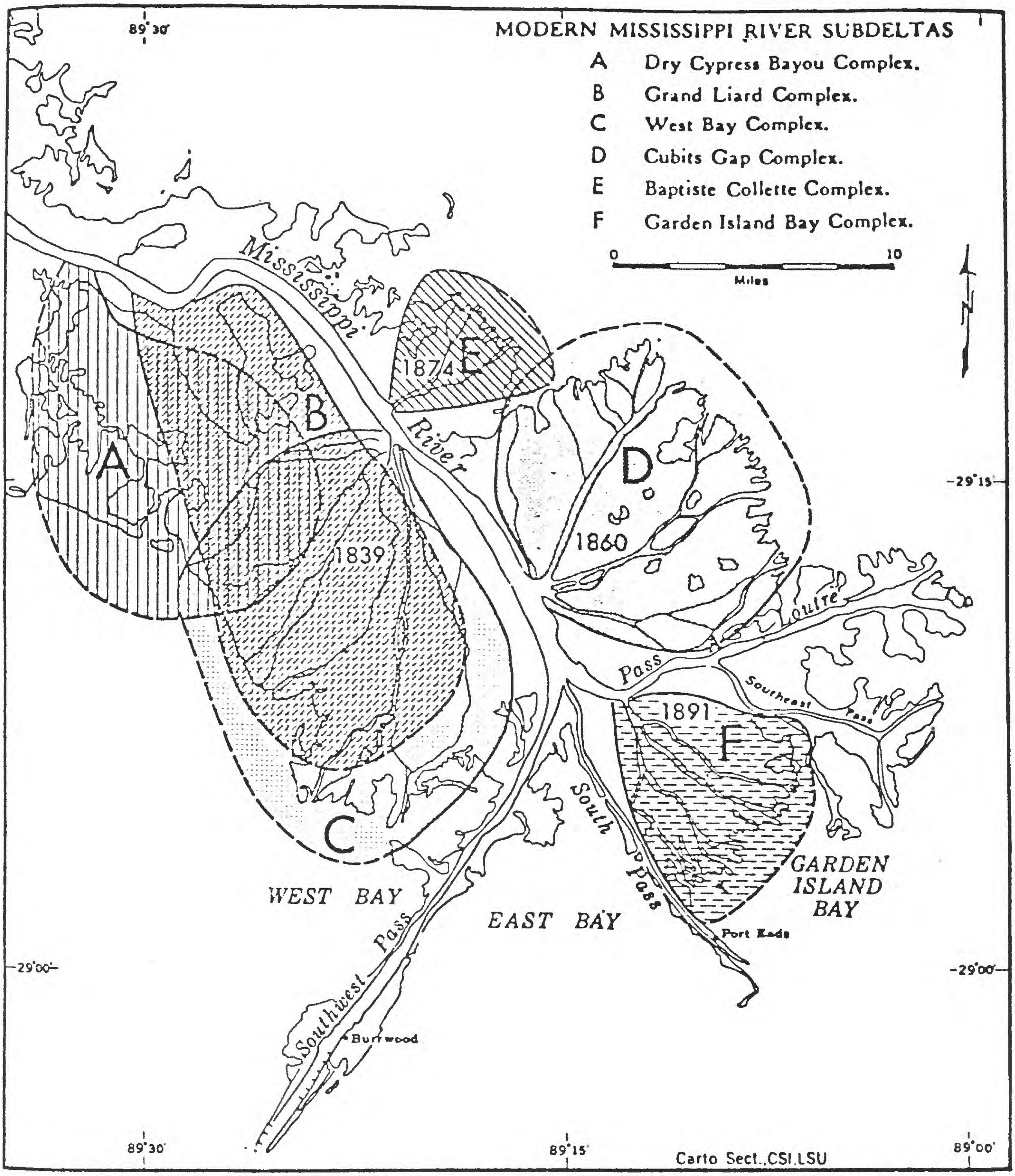

Figure 15.--Modern Mississippi River delta and its subdeltas or major crevasse splays. Dates indicate year of crevasse breakthrough. Adapted from Coleman and Gagliano (1964).

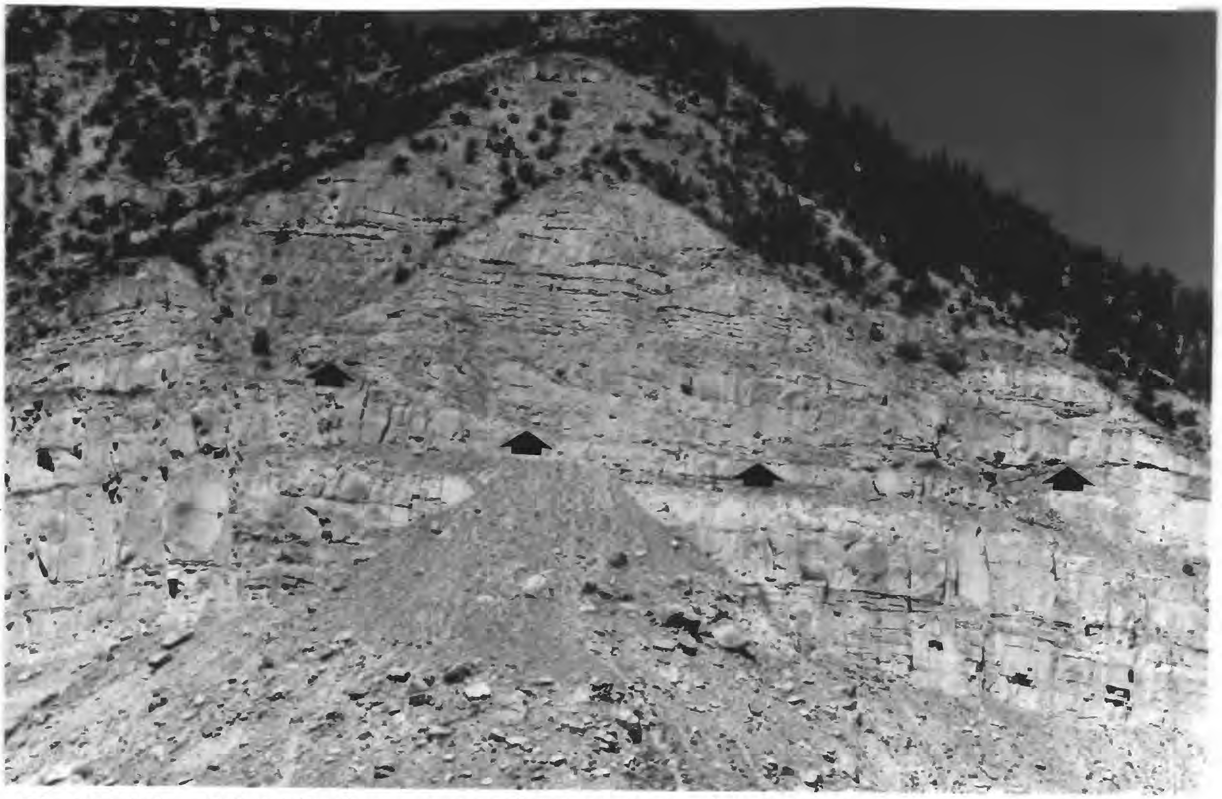

Figure 16.--Distributary channel sandstone (above arrows) scoured into a distributary mouth-bar sandstone in the Blackhawk Formation, Wasatch Plateau, Utah.

Figure 17.--A sequence of internal structures of distributary channel sandstone shown in figure 16. Trough cross beds in the lower part and ripple laminations in the upper part.

Figure 18.--An abandoned deposit of a distributary channel (A), which consists of shale and siltstone, in the Blackhawk Formation, Wasatch Plateau, Utah.

The interchannel areas of the alluvial-dominated delta plain are generally submerged by shallow water bays. Swamps and marshes are developed in emergent interchannel areas proximal to distributary channel-levee complexes. During floods, excess discharges are diverted into the bays from distributary channels by overbanking and/or crevassing processes much like their counterparts in the alluvial plain. Crevassing is the most important process of bay infilling as well as a dominant mode of deltaic progradation and enlargement by creating landforms that serve as marsh and/or swamp platforms. Incursion of floodwaters into the bays through crevasse channels may initially form a minor crevasse splay; however, continued diversion of floodwaters through crevasse channels leads to areal expansion. Conversion of the crevasse channels into principal conduits eventually transforms the splay into a subdelta with minor delta-plain and delta-front components. Both crevasse-splay and subdelta facies are coarsening-upward sequences of muds, silts, and sands. Like their alluvial plain crevasse-splay and lake-splay counterparts, the sands are dominated by ripple lamination (current and climbing) described in detail by Elliott (1974). In the subdelta, the sands may contain ripple lamination in the lower part and small- to large-scale trough cross lamination in the upper part. Synsedimentary deformation structures are also common in both crevasse-splay and subdelta deposits (Roberts and others, 1976). Unlike the lake splay, the muds and silts of the delta-plain crevasse splays and subdeltas contain brackish-marine fauna and burrows. In addition, upon healing of crevasse-splay channels, the crevasse splay-subdelta complex is abandoned and becomes a localized platform for development of oyster reefs (Gagliano and others, 1971), which eventually is overrun by peat-forming marsh and/or swamp. Thus, the areal and vertical distributions of peats are controlled by the mode of progradation of crevasse splay-subdelta complexes. Sedimentary structures of ancient delta-plain, crevasse-splay, and delta-front deposits are shown in figures 19, 20, and 21.

Tide-Dominated Delta Plain Facies

The tide-dominated delta plain and its facies are characterized by reworking and redistribution of fluvially-introduced sediments by flood and ebb tidal currents. This type of delta plain, which occurs in areas with moderate to high tidal range, is exemplified by the modern Irradaway, Mekong, and Ganges-Brahmaputra River deltas (fig. 22). During flood tidal conditions, tidal currents enter distributary channels and spill over banks innundating interchannel areas. The tidal waters recede during ebb tide. Thus, in the distributary channels, sediments are tidally reworked and redistributed in their lower reaches. Consequently, the distributary channels form linear sand ridges aligned parallel to the length of the channel (Coleman, 1969; Fisher and others, 1969; Coleman and Wright, 1975). The sand ridges, which are as much as several miles long and 65 ft high, yield braided channels that display low to moderately high sinuosity. The braided channel pattern is best developed in the lower reaches of the distributary channels where they also flare seaward. The tidally influenced distributary channel deposits show a fining-upward sequence of sands, silts, and muds. From base to top, they consist of a basal lag with marine shell fragments overlain by small- to medium-scale trough cross lamination that passes upward into burrows. The trough cross lamination shows bimodal and bipolar orientations. In contrast to the fluvial-dominated distributary channels, the tidally influenced distributary channels develop fewer abandoned fill deposits.

Figure 19.--A coarsening-upward crevasse splay in a delta plain from the Gallup Sandstone, Gallup coal field, New Mexico. Bioturbated lower part and rippled upper part.

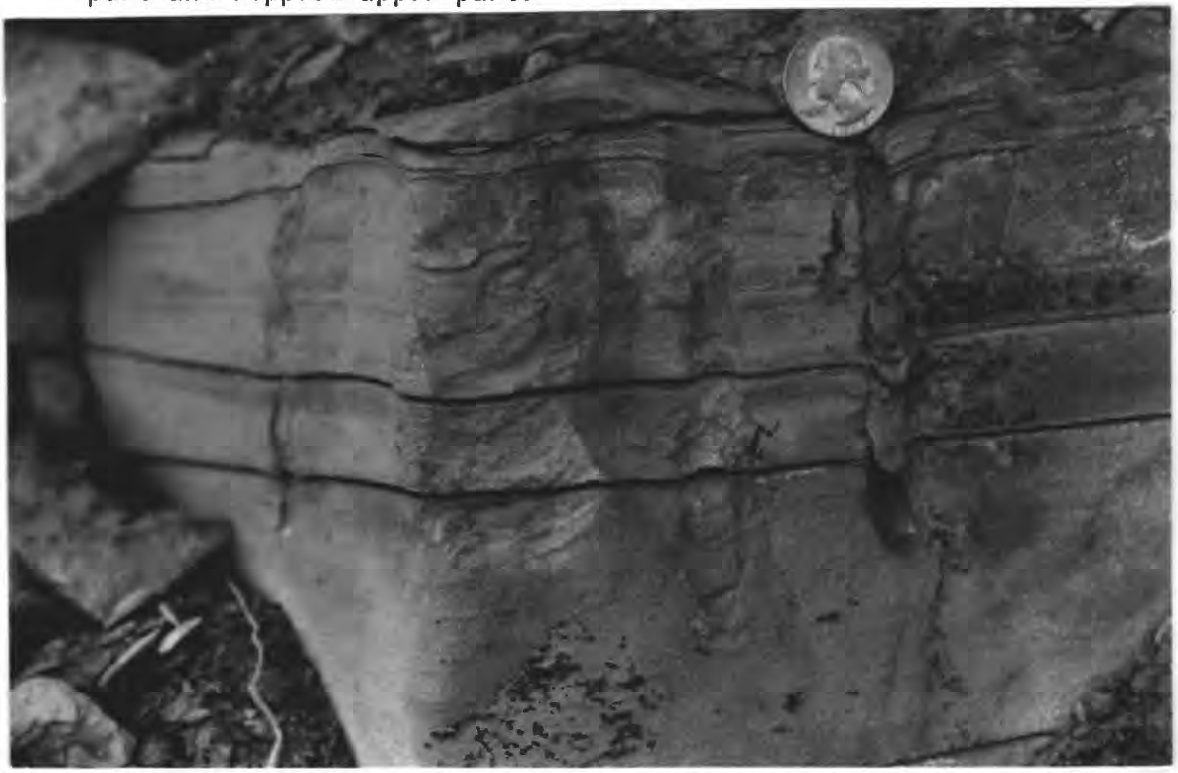

Figure 20.--Diplocriterion trace fossil and ripple lamination in the crevasse-splay sandstone in the upper part of the crevasse-splay deposit shown in figure 19.

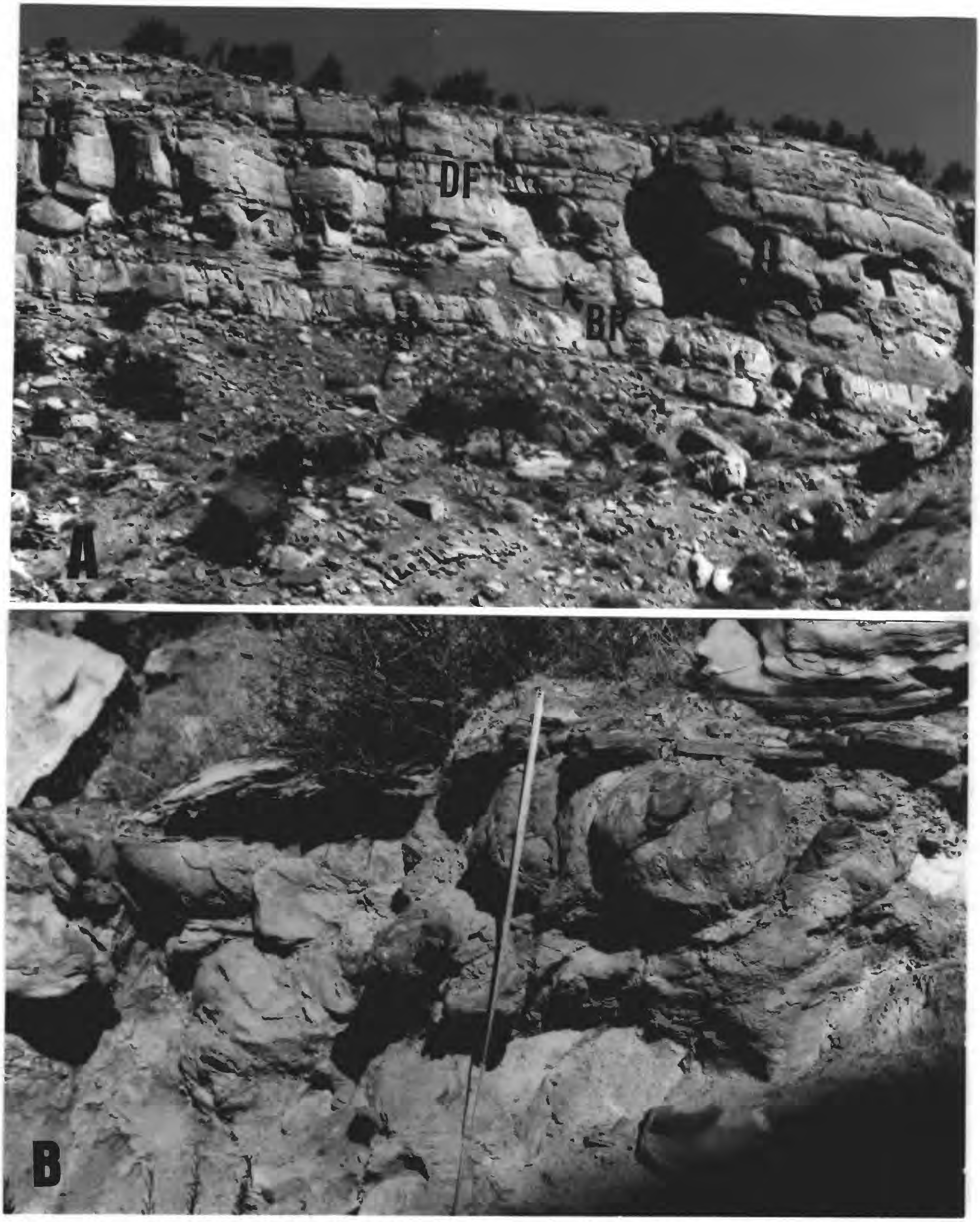

Figure 21.--A delta-front deposit (DF) consisting of coarsening-upward shale, siltstone, and sandstone in A. Lower part of the sequence includes synsedimentary ball-and-pillow structure (BP). A closer view of a ball-and-pillow structure is shown in B.

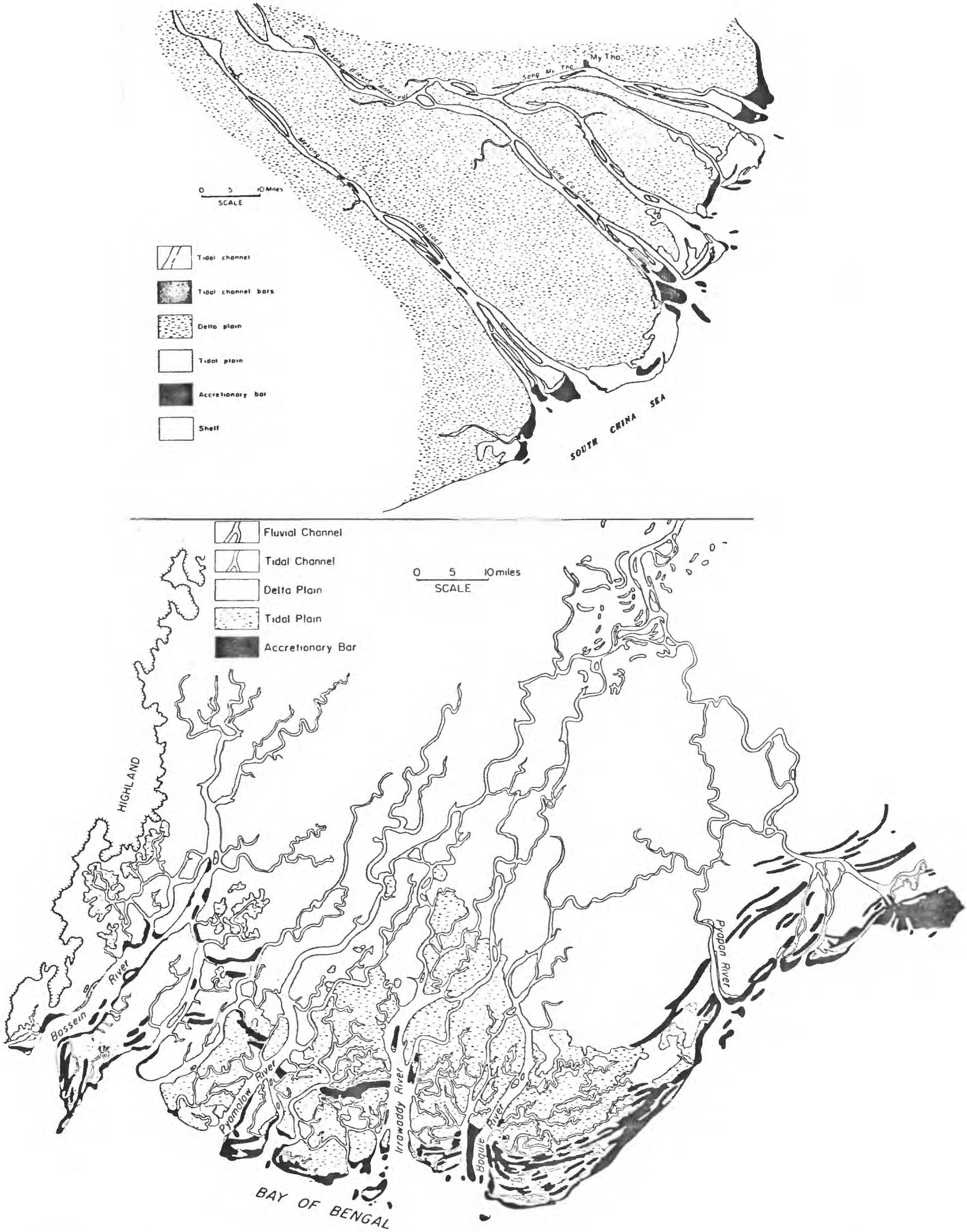

Figure 22.--Tide-dominated delta plains of the Mekong and Irradaway River deltas. Adapted from Fisher and others (1969).

The interchannel areas of the tidal-dominated delta plain consist mainly of intertidal flats, tidal creeks, and swamps. In the Niger delta (Allen, 1965), the tidal-influenced interchannel areas are dissected by densely distributed, dendritic tidal creeks that show meandering trunk channels. These tidal creeks funnel a large volume of tidal waters during flood tides, most of which spill over into the adjacent mangrove swamps. The meandering tidal creeks, which are laterally migrating, form point bar sands and clay plugs like their counterparts in the alluvial plain and alluvial-dominated delta plain. The point bar sands show bimodal trough cross laminations. However, unlike other depositional systems, the tidally dominated interchannel areas contain sparse crevasse splays and abundant intertidal flats; the latter is characterized by intensely bioturbated muds and silts that are occasionally lenticular and flaser bedded. Thus, the interchannel deposits of tidal-dominated delta plain are sheetlike facies consisting of channel sands and clay plugs separated by peat and intertidal sediments.

Wave-Influenced Delta Plain Facies

The wave-influenced delta plain and its facies owe their existence to massive wave reworking and redistribution of distributary mouth sediments into a sheetlike delta-front deposit. The result is a delta plain that is fronted seaward by a series of beach ridges. This type of a delta plain (fig. 23) is exemplified by those formed in Sao Francisco, Grijalva and Nile deltas (Psuty, 1966; Fisher and others, 1969; Coleman and Wright, 1975). The delta plain in these modern examples is supported by one or two distributary channels that show low to moderate and even high sinuosity. They are mainly sand filled. The high proportion of bedload may form braided channel patterns. The distributary channel deposits show classic fining-upward sequences that display dominantly unidirectional trough crossbedded sands in the lower part and rippled silty sand in the upper part. Although the distributary channel deposits resemble those of the alluvial plain and alluvial-dominated delta plain, they are more laterally continuous bodies (perpendicular to depositional strike). In addition, abandoned distributary channel deposits do not occur as commonly as those in the alluvial-dominated and tidal-dominated delta plains.

The interchannel areas of the wave-influenced delta plain consist mainly of swamps, a few lakes, and subordinate crevasse splays. Interchannel or interdistributary bays are absent as a result of the barrier effect of the subaerial beach ridges of the delta-front deposits. Under this condition, the interchannel swamps possess the potential of forming thick, extensive peat deposits. Furthermore, the sheetlike delta-front deposit provides a widespread platform on which delta-plain swamps develop as the depositional system progrades seaward. The crevasse-splay deposits are typically coarsening-upward sequences of muds, silts, and sands that display dominant ripple lamination. Unlike their counterparts in the alluvial plain and alluvial-dominated delta plain, the crevasse splays are isolated bodies that do not serve as precursors of subdeltas. Deltaic progradation involves the entire delta front rather than being concentrated at certain crevasse points.

BARRIER COASTLINE

The barrier coastline discussed in this chapter is a linear sand accumulation and related environments which occur within deltas and along

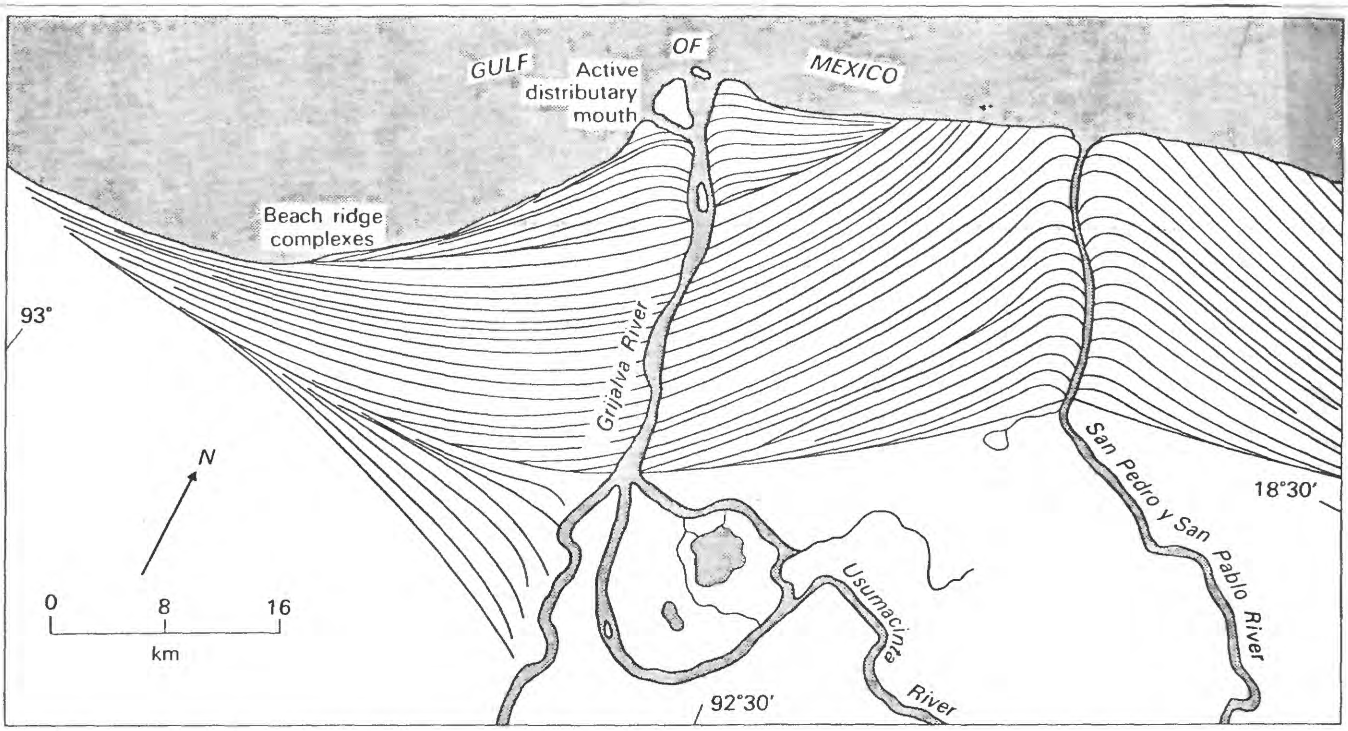

Figure 23.--The delta plain and delta front of the wave-dominated Grijalva River delta, Mexico (Psuty, 1966).

depositional strike from deltas (fig. 24). This shoreline sand is normally separated from land by shallow lagoons or bays and is dissected by tidal inlets. The coastwise linear sand is derived directly from adjoining deltas and is redistributed by longshore currents. Thus, within the depositional continuum in the fluvial systems, it is located at the end opposite the alluvial fan (see fig. 1, chap. I). A considerable number of publications are devoted to the processes and facies of barrier coastlines (LeBlanc and Hodgson, 1959; Bernard and others, 1962; Hoyt and Weimer, 1963; Van Straaten, 1965; Hails, 1968; Berryhill and others, 1969; de Jong, 1971; Hayes and Kana, 1978). Studies of modern barrier coastlines suggest four major environments (Elliott, 1978b; Selley, 1980):

1. Fluviatile coastal plain.
2. Lagoonal and tidal flat complex.
3. Barrier island.
4. Offshore marine shelf.

Fluviatile Coastal Plain Facies

The fluviatile coastal plain and its facies are made up of deposits of fluviodeltaic systems similar in characteristics to those deposits described for alluvial and deltaic plains. However, these systems are small in scale as typified by bayhead deltas that formed at the landward side of bays or lagoons. The rivers that feed the bayhead deltas are short headed. The lithological characteristics and sedimentary structures of the fluviatile coastal plain facies are very similar to those of their counterparts in the alluvial and delta plain. Like the alluvial and deltaic plain, these fluviodeltaic systems developed swamps that make up the most important peat-forming environment in the barrier coastlines.

Lagoonal and Tidal Flat Facies

The fluviatile coastal plain passes seaward into lagoons or bays and tidal flats. The characteristics of these lagoons and their facies are chiefly controlled by their tidal settings. That is, microtidal lagoons contain brackish to hypersaline waters resulting from limited interplay with the open sea due to sparse tidal inlets through the barrier islands. Consequently, lagoonal infilling from the barrier side is mainly by landward-flowing washover fans generated by storms (Hayes, 1967). Washover fans result from erosion of sediments from the seaward side of the barrier island and are transported through storm-breached, washover channels into the lagoons. Upon reaching the back barrier, the channelized washover flow transforms into sheet flow forming a lobe-shaped deposit. Coalesced deposits of these washover fans form the back-barrier flat that becomes a salt marsh and burrowed tidal flat as the washover channels are healed. Sedimentary structures of washover fan deposits include parallel to subparallel lamination at the proximal part and lagoonward-dipping foresets at the distal part, as well as trough crossbeds at the washover channels (Schwartz, 1975).

In mesotidal lagoons, the water is characterized by normal salinity due to continued exchange of lagoonal and open sea waters through numerous tidal inlets in the barrier islands. In this setting, flood-tidal deltas are important sites of deposition at the lagoonal inlet mouths. Bedforms produced in the flood-tidal delta range from sandwaves and megaripples to ripples that

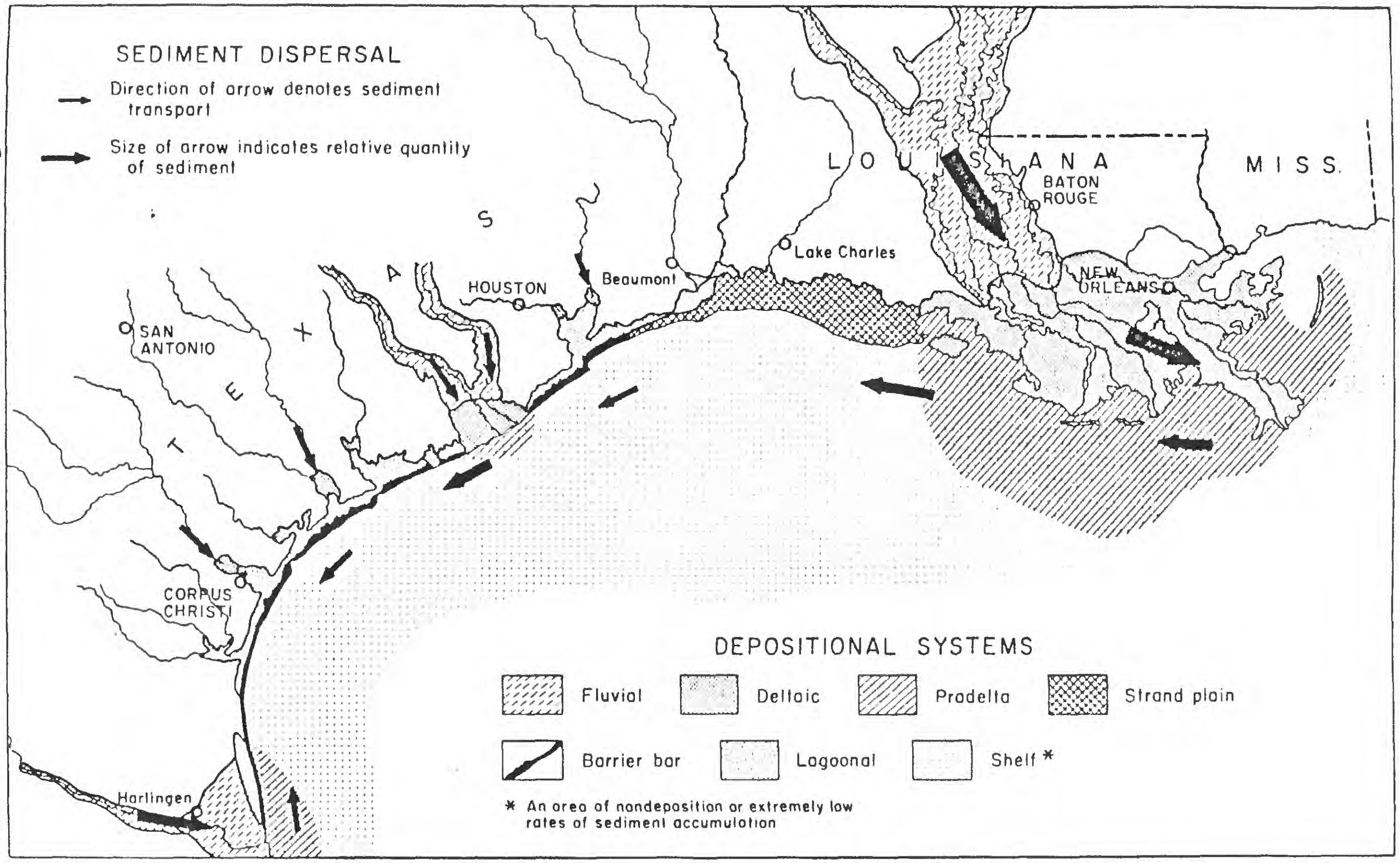

Figure 24.--Barrier coastlines and adjacent minor and major deltas in the northwest Gulf of Mexico. Adapted from Fisher and others (1969).

are represented in internal structures as large-scale trough and planar crossbeds, and ripple lamination, respectively. Thus, the lagoons are infilled along their margins by bayhead deltas at the landward side and by washover fans and flood-tidal deltas at the seaward side. Examples of a modern washover fan and an ancient flood tidal delta deposits are shown in figures 25 and 26.

The lagoon proper accumulates suspended fine-grained sediments from the marginal dispersal systems. These lagoonal sediments consist of muds and silts that are locally parallel laminated but generally structureless due to intense bioturbation. In addition, they contain organic-rich sediments with plant remains that are rafted from outlying swamps. The lagoonal sediments also contain enclosed brackish-water faunas. An example of an ancient lagoonal deposit is shown in figure 27. In ancient analogs, the filled surfaces of lagoons often served as broad swamp platforms on which thick coals accumulated. The coal-forming swamp also continued to develop over the barrier-island facies as the barrier coastline prograded seaward.

Barrier Island Facies

The lagoon is separated from the open sea by a barrier island that consists of facies that are controlled by sea-level fluctuations, rates of emergence and/or subsidence of depositional area, and variation in sediment supply. These factors may yield progradational and transgressive barrier deposits. The progradational barrier deposits, which are controlled by continued sediment supply, stable sea level, and low to moderate subsidence, consist of a coarsening-upward sequence from lower shoreface, foreshore, and aeolian dunes. These deposits vary from interbedded muds, silts, and sands in the lower shoreface grading upward into fine to medium sands in the upper shoreface, foreshore, and aeolian dunes. Sedimentary structures, from base to top, include intensely bioturbated and rippled lower shoreface deposits overlain by shelly, parallel-laminated, planar-crossbedded upper shoreface deposits that grade upward into seaward-dipping, subparallel lamination and trough crossbedded foreshore deposits, as well as trough crossbedded, rooted aeolian dune deposits (Bernard and others, 1962). In contrast to the progradational barrier deposits, the transgressive barrier deposits consist mainly of shoreface deposits that resemble those of the former deposits. This results from erosion of the overlying foreshore and aeolian dune deposits due to reduced sediment supply, rise in sea level, or high rate of subsidence of the depositional area. The progradational barrier deposits stand a better chance of being preserved than the transgressive barrier deposits in ancient rocks. Ancient progradational barrier deposits and their sedimentary structures are shown in figure 28.

The progradational and transgressive barrier deposits are locally replaced by tidal inlet and ebb tidal deposits. The tidal inlet deposits, from base to top, consist of gravel lags of shells and pebbles grading upward into bidirectional, planar crossbedded (produced from sandwave bedforms) and trough crossbedded sands with reactivation surfaces overlain, in turn, by parallel laminated, rippled, and planar- and trough-crossbedded sands (Kumar and Sanders, 1974; Hayes and Kana, 1978). Herringbone crossbeds, which show bipolar current directions, are common internal structures of tidal inlet sandstones. The ebb tidal deltas, located at the seaward mouths of tidal inlets, form broad sand accumulations on the beach face that display dune and

Figure 25.--An example of a modern washover fan-channel complex (W) along the Gulf Coast barrier coastline.

Figure 26.--Flood tidal delta sandstone showing sequence of internal structures in the Point Lookout Sandstone, Gallup coal field, New Mexico. The foresets of planar crossbeds dip landward.

Figure 27.--A lagoonal infill deposit consisting of burrowed, organic-rich shale in the lower part and rippled upper part from the Blackhawk Formation, Wasatch Plateau, Utah.

Figure 28.--Shoreface deposits (A) and their internal structures (B) of bimodal planar crossbeds in a progradational barrier facies in the Pictured Cliffs Sandstone, San Juan Basin, New Mexico.

sandwave bedforms (producing trough and planar crossbeds) with bidirectional foreset directions (Oertel, 1972, 1976; Hine, 1976). Ancient tidal inlet deposits and their sedimentary structures are shown in figures 29 and 30.

Offshore Marine Shelf Facies

The shoreface deposits of the barrier island grade laterally into deeper water sediments of the offshore marine shelf facies. These facies include mainly intensely bioturbated muds. Some silts occur as fine laminations or lenticular units with the muds. Horizontal bioturbation is a more common type of biogenic structure than vertical burrows.

SUMMARY

This chapter summarized the facies characteristics of alluvial fans, aluvial plains, deltaic plains, and barrier coastlines since they are directly related to peat- or coal-forming environments. These coal-associated facies are the most frequently encountered facies of the fluvial systems formed in ancient rocks. Thus, an understanding of the lithological variations and sedimentary structures of these facies is necessary in order to relate to the basin facies analyses of ancient coal-bearing rocks to be discussed in the next chapter.

Figure 29.--Tidal inlet sandstone (above arrows) in the Pictured Cliffs Sandstone, San Juan Basin, New Mexico.

Figure 30.--Herringbone crossbeds indicating flood and ebb tidal current fluctuations in the tidal inlet sandstone shown in figure 29.

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CHAPTER III

CASE HISTORIES OF ANCIENT FLUVIAL SYSTEMS AND THEIR COAL-FORMING ENVIRONMENTS

INTERMONTANE TRUNK-TRIBUTARY AND ANASTOMOSED FLUVIAL DEPOSITS

Recognition of trunk-tributary and anastomosed fluvial deposits in modern settings is simplified by the ability to visualize their geomorphologic architecture in large basins at any instant of time. Under this condition, components of the fluvial system, such as streams or tributaries feeding into the axially or centrally located principal channel or trunk are easily observed in their hierarchical arrangements. A similar situation exists in a modern anastomosed river system (Smith and Smith, 1980) where numerous coeval low-sinuosity channels diverge and converge within an alluvial plain. An important controlling factor in the development of intermontane trunk-tributary fluvial systems is their tectonic setting (Miall, 1981). In the Powder River Basin in Montana and Wyoming, the Paleocene-Eocene drainage system (fig. 1) was controlled by:

1. Uplifts that surrounded the basin, which consisted of the Bighorn Mountains to the west, Casper Arch-Laramie Mountains-Hartville uplift to the south, and Black Hills Mountains-Cedar Creek anticline to the east.
2. Basin lineaments or zones of fractures along which recurrent movements developed as a result of periodic reactivation of basement block boundaries throughout Phanerozoic time (fig. 2).

The Powder River Basin and surrounding uplifts were products of the Cretaceous-Tertiary Laramide orogeny. The uplifts around the basin margins were formed by the end of Cretaceous time (Curry, 1971; Seeland, 1976) but were topographically subdued by late Paleocene time. Lower Paleocene deposits (lower part of the Fort Union Formation) marked the first evidence for the development of the dominant structural feature of the Powder River Basin (Curry, 1971). Upper Paleocene deposits (upper part or Tongue River Member of the Fort Union Formation) marked the beginning of significant uplift in the surrounding area (Love, 1960) which continued into the Eocene as recorded by the Wasatch Formation.

The drainage network in the Powder River Basin during late Paleocene time (Tongue River Member) assumed characteristics of a trunk-tributary fluvial system (Galloway, 1979; Ethridge and others, 1981; Flores, 1980, 1983). Galloway's (1979) summary of the trunk-tributary drainage for the upper part of the Fort Union Formation in the southern part of the Powder River Basin is shown in figure 3. The fluvial facies consists of:

1. Fluvial axes facies consisting of mixed load (meandering) and suspended load (anastomosed) channel deposits.
2. Flanking facies of backswamps, lacustrine, and floodplain-tributary deposits.

The continuation of the drainage in the northern part of the Powder River Basin was defined by Flores (1980, 1983) and is shown in figure 4. The northerly oriented trunk channels in the southern part of the basin transformed into northeasterly oriented trunk channels in the northern part. However, as in the southern part of the basin, the trunk channel deposits are flanked by lacustrine-floodplain, backswamp, and tributary channel deposits. The trunk-tributary drainage network of the Powder River Basin was maintained through the Eocene time. However, in the intervening time (uppermost Tongue River Member), an anastomosed fluvial system was developed.

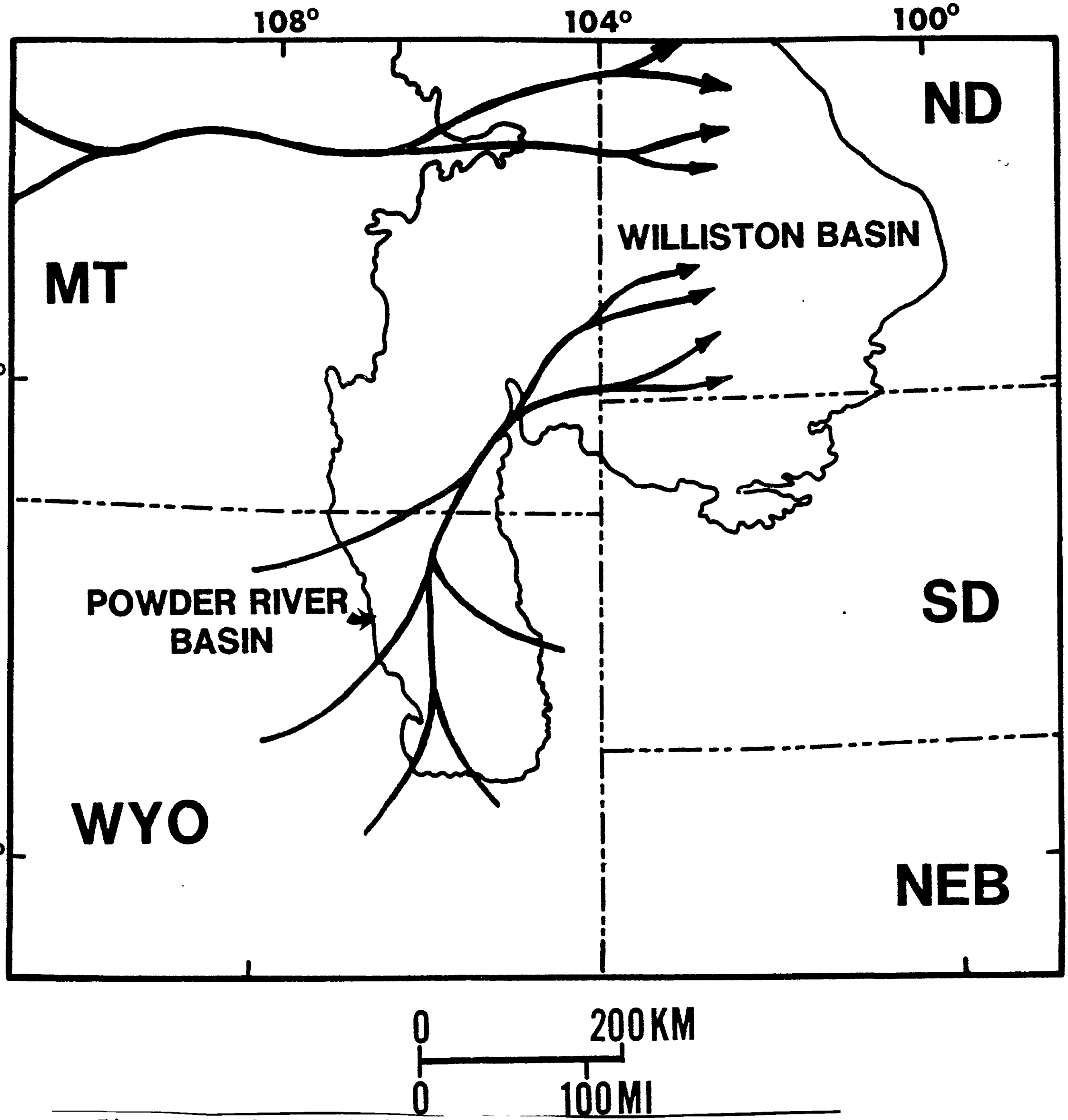


Figure 1.--Fluvial drainage patterns in the Powder River Basin and Williston Basin in Montana, Wyoming, and North Dakota.

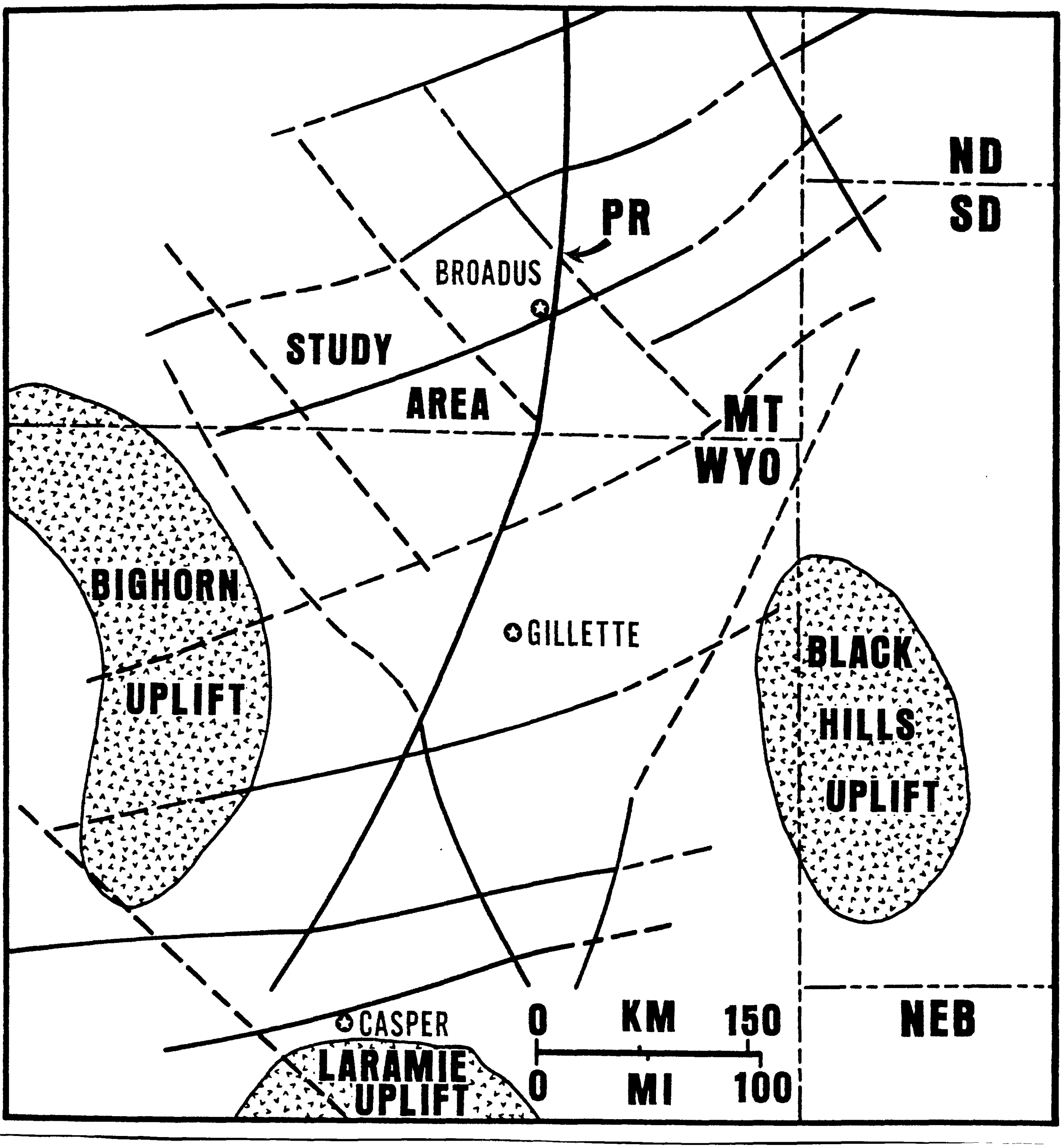


Figure 2.--Patterns of lineament structures in the Powder River Basin. Adapted from Flores (1983).

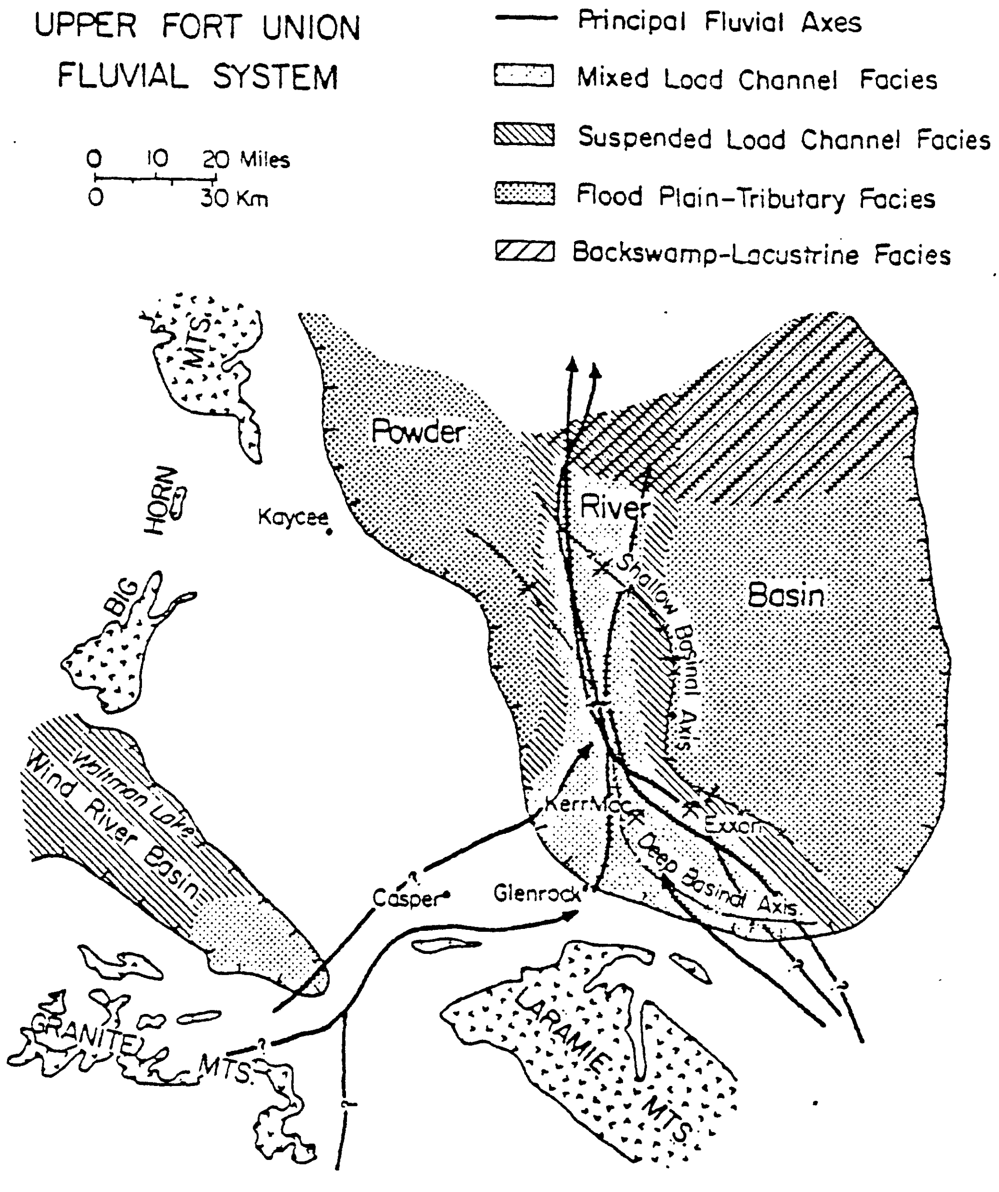


Figure 3.--Paleocene fluvial drainage system in the southern Powder River Basin. Adapted from Galloway (1979).

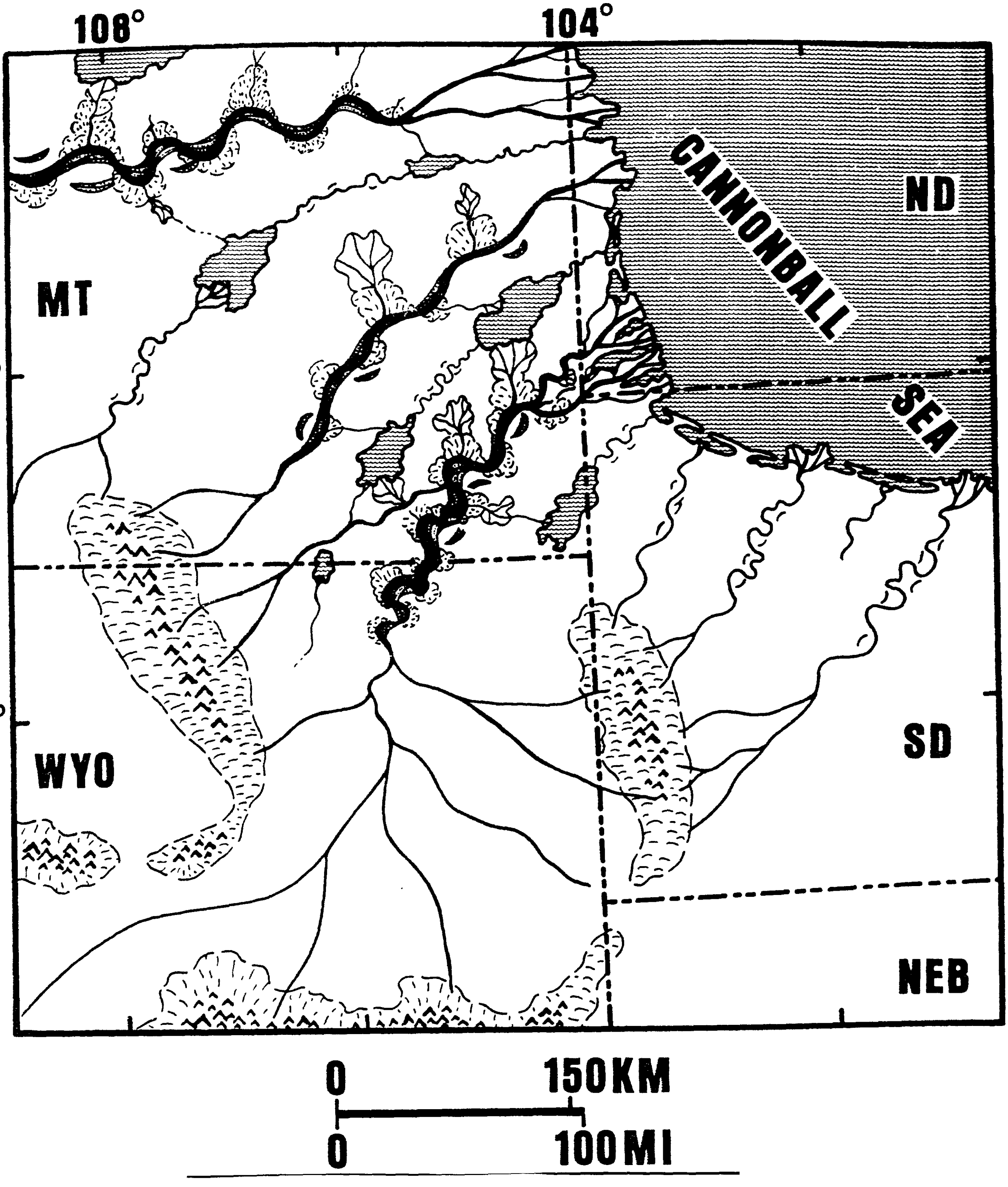


Figure 4.--Paleocene trunk-tributary and coastal plain fluvial systems in the Powder River Basin and Williston Basin, respectively. Adapted from Flores (1983).

Based on grain size, grain shape, and crossbed measurements of channel sandstones, Seeland (1976) suggested a trunk-tributary fluvial system (fig. 5) as the depositional setting of the Eocene Wasatch Formation, which overlies the Fort Union Formation. The isopleth map and vector mean crossbed map (fig. 6) constructed from the channel sandstones of the Wasatch Formation indicate areas of local sources and transport directions of sediments. Figure 6 shows three distinct source areas:

1. The Black Hills uplift drained by westward-flowing streams.
2. The Laramie Mountains-Hartville uplift drained by northward-flowing streams.
3. The Bighorn Mountains drained by eastward-flowing streams.

The streams apparently served as tributaries that converged into common northward-flowing trunk channels.

Tectonic Sedimentation

The Paleocene-Eocene fluvial sedimentation of the Powder River Basin was controlled by tectonism in three ways:

1. The orientation and location of the trunk and tributary channels.
2. The drainage pattern.
3. The grain size of the deposits.

The orientation of the tributaries during the Eocene and Paleocene is perhaps best related to the northeast and northwest orientations of the lineaments (figs. 2, 3, 4, and 5). The tributaries probably preferentially aggraded within lineament belts. The orientation of the trunk during the Paleocene may have been controlled by a north-northeast trending lineament (PR) that ran from west of Casper to west of Gillette, Wyo., continuing on to east of Broadus to east of Miles City, Mont. (figs. 2 and 4). The north-northeast trending trunk and lineament may have been coincident to the basin axis during the Paleocene. During Eocene time, the trunk or longitudinal river shifted to the west (fig. 5) following the present axis of the basin.

In response to tectonic setting, the drainage pattern of the trunk river transformed from meandering to anastomosed during the Paleocene. During the Eocene, the tributaries or traverse streams are traceable at their headwaters to alluvial fans at the western part of the basin flanking the Bighorn Mountains (fig. 5). At this time, the bounding Bighorn uplift was tectonically active as indicated by deposition of the Eocene Kingsbury and Moncrief Conglomerate Members of the Wasatch Formation in alluvial fans and braidplains. These deposits may have been preceded by older alluvial fan deposits of the Upper Fort Union Formation, which consists of as much as 300-ft-thick conglomerates that are distributed along the northeast flank of the Bighorn Mountains from south of Sheridan to south of Buffalo, Wyo. (Hose, 1955; Mapel, 1959). A reconstruction of the Wasatch fluvial setting at the northeast flank of the Bighorn Mountains is shown in figure 7 (Obernyer, 1978). Growth faults along the front of the uplift also may have influenced overthickening of Eocene coals.

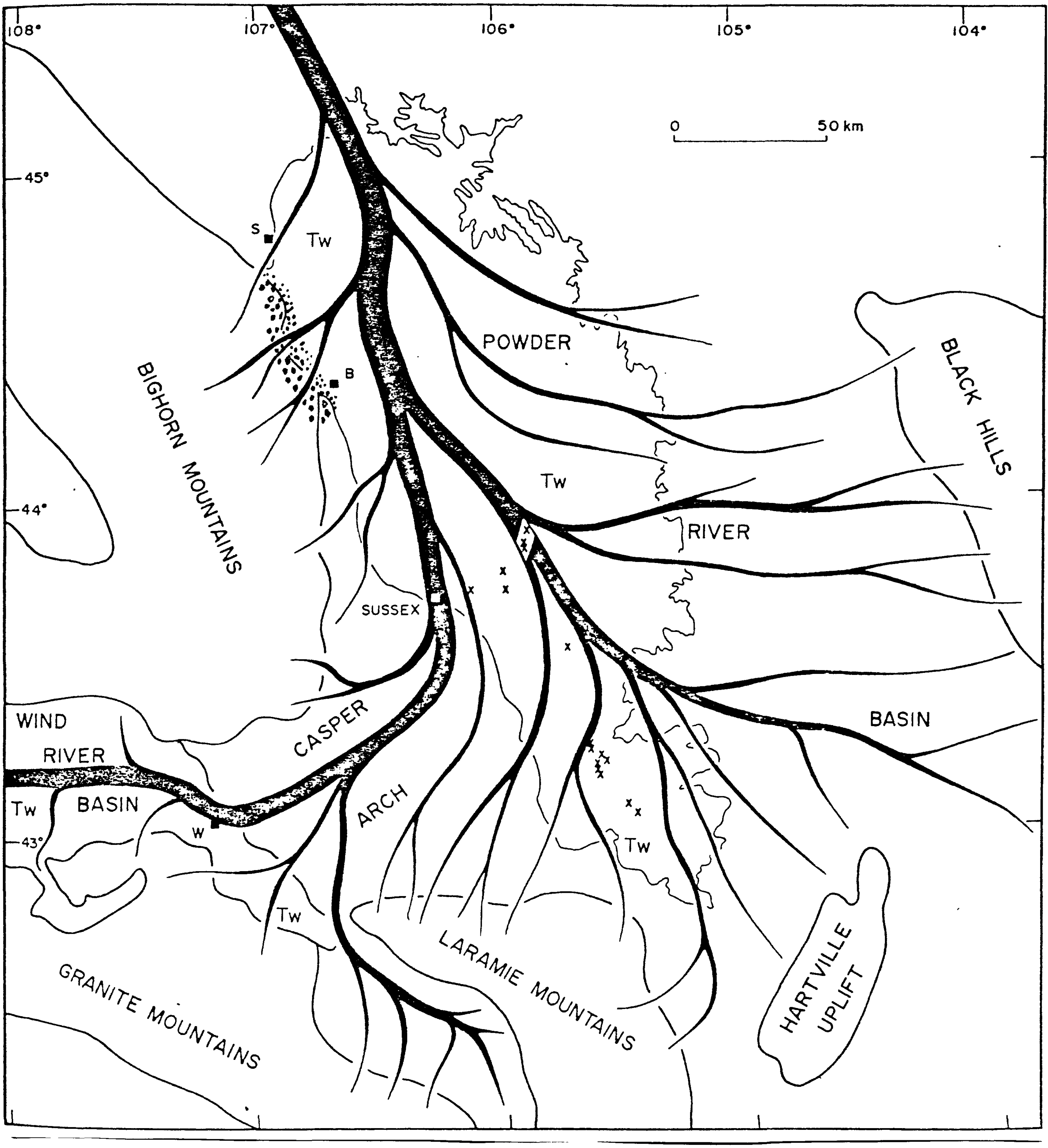


Figure 5.--Eocene trunk-tributary fluvial setting in the Powder River Basin. Modified from Seeland (1976).

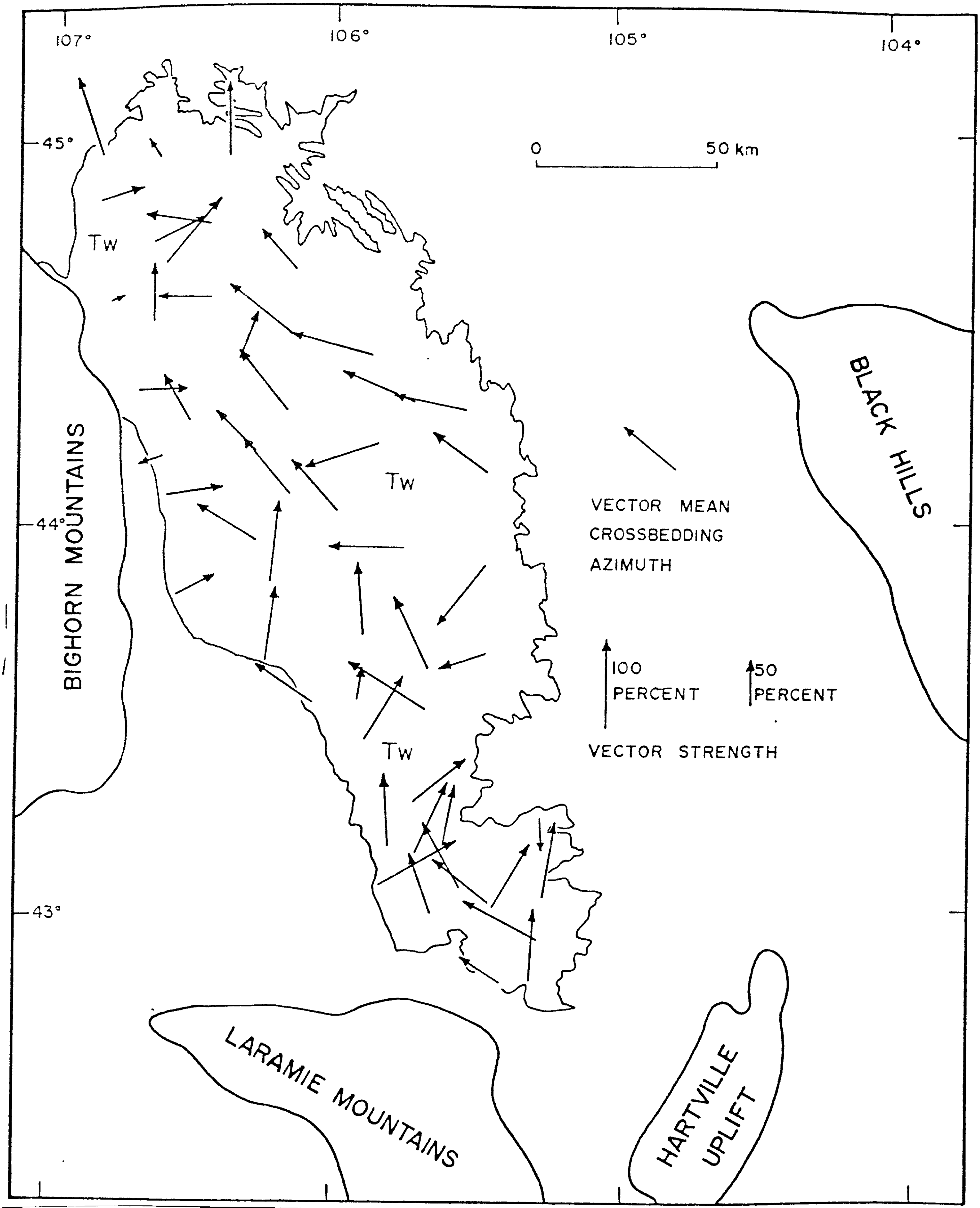


Figure 6.--Crossbed orientations and distributions of the Wasatch (Tw) channel sandstones in the Powder River Basin. Adapted from Seeland (1976).

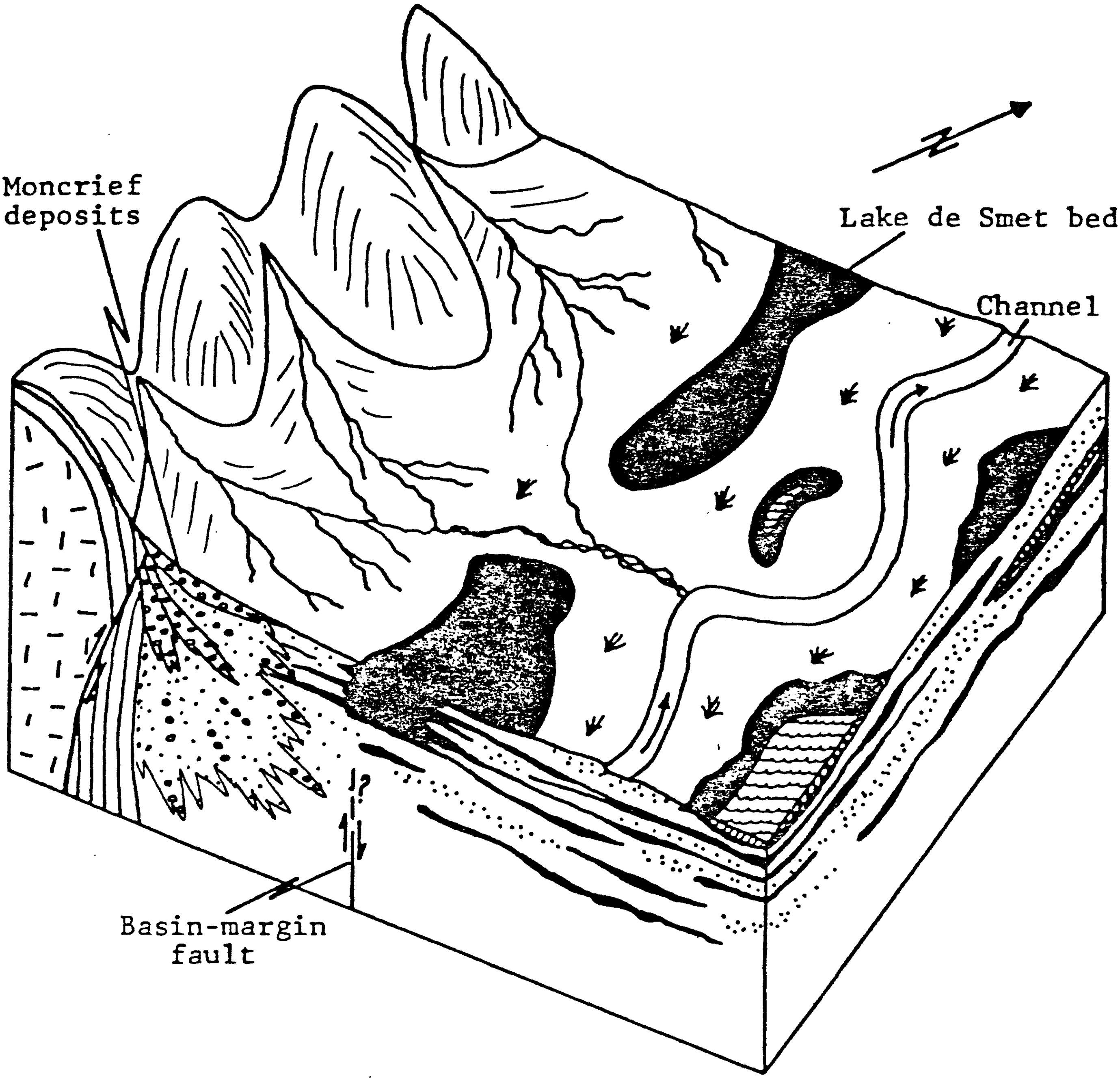


Figure 7.--Block diagram showing relationships between transverse alluvial fans, alluvial plain, and swamp in which the Lake de Smet coal bed accumulated in the Wasatch Formation in the vicinity of Buffalo, Wyo. Modified from Obernyer (1978).

FACIES OF TRUNK-TRIBUTARY AND ANASTOMOSED SYSTEMS

The lower part of the Tongue River Member of the Fort Union Formation exposed in the northern Powder River Basin contains three major facies:

1. Channel-levee facies of the trunk, tributary, and anastomosed rivers.
2. Floodplain-lacustrine facies of the intratrunk basins.
3. Backswamp facies that cover both of the above environments.

Channel-Levee Facies

The channel-levee facies consists of depositional elements that include channel sandstone bounded by levee deposit of silty sandstone, siltstone, and shale. This facies occurs either as isolated or clustered deposits. In parts of the stratigraphic interval where this facies comprises the bulk of the deposits, the facies consists of closely spaced, high-density channel sandstones bounded by levee deposits. The common occurrence of this facies in the interval, as indicated by their clustering, imparts an overall channel-dominated characteristic to the facies.

The channel sandstones are basally erosional, fine to coarse grained, and form a fining-upward sequence. They are as much as 150 ft thick and 8 mi wide. The channel sandstones may consist of more than one body that include accretion-bar deposits separated by epsilon crossbeds (Allen, 1963), an evidence of lateral migration of the channels. The lower part of the channel sandstones contains large-scale trough and tabular crossbeds locally interrupted by large convolute laminations; the upper part contains small-scale trough crossbeds, convolute lamination, and ripple lamination. The large-scale crossbedded lower part probably represents dune and bar migration in the main channel fill. The small-scale crossbedded upper part probably represents channel fill in shallower water during abandonment. The channel sandstones contain lag deposits of rounded sandstone, siltstone, shale, limestone, ironstone, carbonaceous shale, coal spar, and mollusk fragments that probably represent reworked deposits of the floodplain-lacustrine facies. Rare channel sandstones that consist of about equal amounts of sandstone, siltstone, and shale probably represent abandoned deposits.

The channel sandstones consist of three types: trunk, tributary, and anastomosed channel sandstones. Although the channel sandstone types have similar internal structures, they differ from each other in their grain-size variation, thickness, and width. The tributary channel sandstones are narrower, generally thinner, and a few contain more siltstone, shale, and silty sandstone beds as drapes between accretion units than do the trunk channel sandstones. In addition, the tributary channel sandstones are encased in abundant overbank siltstone and shale. The anastomosed channel sandstones are characterized by vertical stacks of channelet sandstones and an absence of point bars as well as shale-siltstone drape deposits.

Thick sequences of alternating beds of siltstone, shale, carbonaceous shale, ironstone, and fine-grained sandstone commonly border channel sandstones, and also are repeated vertically and laterally in overlapping arrangements. The sandstone and siltstone show ripple drift and lenticular ripple laminations. This lithologic sequence commonly is rooted and contains tree stumps in growth position; such stumps were recognized by Allen (1965) as important features of a levee deposit. The levees resulted from settling of suspended load during repeated overbank floodings. The interbedded sequence of different size particles reflects the fluctuating floodwater stages, and the overlapping of these deposits suggests localized overtopping of low-lying areas of the levees.

Floodplain-Lacustrine Facies

The floodplain-lacustrine facies grades laterally into the channel-overbank facies. The floodplain-lacustrine facies is characterized by abundant crevasse-splay sandstone, siltstone, shale, limestone, and ironstone. These lithic types commonly contain freshwater mollusk fossils (Flores, 1981). The facies consists of two major deposits: crevasse splay and lake splay.

The crevasse-splay deposits are tabular in shape and have an average thickness of 15 ft. They are either a single coarsening-upward sequence of shale, siltstone, and sandstone or a stack of coarsening-upward sequences. The sandstone occurs as a sheetlike deposit that is as much as 25 ft thick and as much as 9 mi wide. The sandstone can be divided into crevasse-channel and crevasse-splay types. The crevasse-channel sandstone, which is a fining-upward sequence, is basally erosional, very fine to medium grained, and contains small-scale trough crossbeds and ripple laminations. The crevasse-channel sandstone probably represents secondary drainage conduits of the crevasse splay that served to transmit sediments to the distal area. The crevasse-splay sandstone is a very fine grained to medium-grained, coarsening-upward deposit and has gradational or sharp basal contacts. Internal structures of the crevasse-splay sandstone are dominated by various combinations of climbing ripple (ripple drift) lamination, asymmetrical ripple lamination, parallel lamination, small-scale cross lamination, and convolute lamination. A few degassing and pelecypod escape structures are present and smooth-walled vertical burrows and root marks are common. Rippled silty sandstones commonly are intercalated in the crevasse-splay sandstone. The abundant ripple laminations suggest deposition by slow-moving currents on margins of bodies of standing water such as lakes. That the influx of detritus into the crevasse splay was episodic and triggered by flood stages is indicated by the silty intercalations and escape structures of pelecypods.

The crevasse-splay sandstone is underlain by coarsening-upward shale and siltstone, interbedded with ironstone layers, which contain abundant vertical and horizontal burrows. The shale and siltstone in the sequence thin or thicken depending upon whether these deposits are at the distal or proximal ends of the crevasse splay, respectively. The overall crevasse-splay deposit probably represents deposition during flood conditions in which the fluvial channels at bankfull stage breached levees and deposited fan-shaped sediments into the floodplain. Vertical stacking of crevasse-splay deposits resulted from overlapped progradation and lateral shifts of the splay lobes. Continuous progradation of the crevasse splay into distal parts of the floodplain led to their debouching into ephemeral lakes. This process of deposition resulted in the accumulation of lake-splay or crevasse-delta deposits.

The lake-splay or crevasse-delta deposits consist of sequences of shale, siltstone, and silty sandstone interbedded with limestones. These sequences contain freshwater mollusk fossils. The shale, siltstone, and silty sandstone are ripple laminated and burrowed. Some vertical burrows represent escape structures of pelecypods. Pulses of sediment that discharged into ephemeral lakes from crevasse splays during flood events probably disturbed bottom fauna and resulted in their catastrophic burial or rapid escape. Generally, the limestone is mud-supported, thin, lenticular, and contains freshwater mollusk fossils. Limestone that has a high detrital content is commonly unfossiliferous. The limestone probably represents carbonate precipitation in ephemeral lakes during cessation of detrital influx or nonflood conditions. It also reflects deposition in distal parts of lakes not subjected to detrital sedimentation. Localized as well as widespread distribution of mollusk fossils occur in the floodplain-lacustrine facies. These fossiliferous zones mark the sites of lakes and suggest that lacustrine processes in the floodplain were active to various degrees, perhaps controlled by local or regional subsidence and compaction of underlying sediments. Biostratigraphic and biofabric studies of the freshwater mollusk fossils by John Hanley (oral commun., 1981) can document the evolutionary development of the lakes in terms of their depth, size, and relationship to external sedimentation.

Backswamp Facies

The backswamp facies, which formed in the floodplain, is closely related to the floodplain-lacustrine facies. The major deposits of the backswamp facies are coal and carbonaceous shale that consist of clay- and silt-size particles mixed with varying proportions of organic material. The coal beds are underlain, overlain, and grade laterally into and interbed with the carbonaceous shale beds. A minor amount of organic-rich siltstone and sandy siltstone occurs as partings in the coal beds. The coal beds range in thickness from a few inches to as much as 60 ft. The coal beds associated with the fluvial deposits of the intermontane basin fluvial system are usually thicker than those associated with deposits of the coastal plain fluvial system. The coals associated with the fluvial deposits range from lignite to subbituminous. Individual coal beds, which can be traced laterally along outcrop for as much as 20 mi prior to splitting or merging, are more laterally extensive in the intermontane basin fluvial deposits than in the coastal plain fluvial deposits. The discontinuity of coal beds associated with the fluvial facies resulted from their erosion by channels, splitting by overbank detrital sediments, and thinning over the channel-overbank deposits. In a few places, thick coal beds are immediately above channel-overbank deposits. This situation suggests that abandoned channel-overbank deposits served as topographically high alluvial ridges that frequently were encroached upon by adjoining backswamp.

In contrast, coal beds associated with the floodplain-lacustrine facies are more laterally continuous than those associated with the channel-dominated facies, as demonstrated in subsurface coal-bed correlations by Culbertson (1980). However, the coal beds associated with the floodplain-lacustrine facies are prone to splitting by crevasse-splay deposits. Thick coal beds occur at the proximal and distal ends of the floodplain-lacustrine facies; however, the coals are thickest at locations distant from the channel-overbank deposits.

Perhaps the most significant difference between coal beds associated with the intermontane basin fluvial deposits and coastal plain fluvial deposits is their thickness. The unusually thick coal beds of the backswamp facies of the intermontane basin fluvial depositional setting can be explained by the following interrelated factors:

1. Entrenched aggradation of fluvial channels.
2. Subsidence due to basement tectonic control.
3. Differential compaction of the various suites of alluvial plain facies.
4. Length of time of peat accumulation.
5. Nature of the backswamp's paleoflora.
6. Paleoclimate.

Comparison of regional three-dimensional relationships of facies based on cross sections and paleogeographic maps of the trunk-tributary and anastomosed fluvial deposits illustrates the changing facies associations of coal occurrence.

1. The thick, minable coal deposits in the trunk-tributary fluvial system occur as zones mainly associated with the intratrunk-tributary facies.
2. The zones of thick coal beds are separated by floodplain-lacustrine facies.
3. Coals associated with facies of anastomosed streams are thin.
4. Coals related to facies of anastomosed streams contain abundant carbonaceous shale and are associated with floodplain-lacustrine facies.

Facies Stratigraphy and Deposition

The lateral and vertical relationships of the three major facies of the trunk-tributary and the anastomosed fluvial deposits are shown in cross sections (figs. 8, 9, and 10).

Trunk-Tributary System

Figure 8 shows, in the lower part of the Tongue River Member, two local areas of vertically stacked channel-levee facies that grade laterally into floodplain-lacustrine and backswamp facies. These facies grade upward into en-echelon arranged channel-levee facies, which in turn grade laterally into floodplain-lacustrine and backswamp facies. The vertically stacked channel facies represent localized aggradation of trunk rivers probably influenced by tectonic subsidence related to lineament structures. Prolonged aggradation at any one place in the alluvial plain, if maintained for a long period of time, caused floodplain-lacustrine conditions at the flanks and development of poorly drained backswamps in the distal parts where thick coals accumulated. These locally aggrading trunk river channels were succeeded by deposits of laterally shifting channels that were also flanked by floodplain-lacustrine facies separated by thick backswamp coals.

Figure 9 shows the three-dimensional variations of the deposits of the laterally shifting channels and fringing floodplain-lacustrine-backswamp facies. The laterally equivalent intratrunk and tributary facies to the western part of the Powder River Basin is shown in figure 10. In this part of the Powder River Basin, based on subsurface data, the intratrunk-tributary facies are associated with very thick coals (fig. 11). On the whole, the unusually thick characteristics of the coals in this part of the basin may be due to:

1. Depositional environment.
2. Tectonic setting.

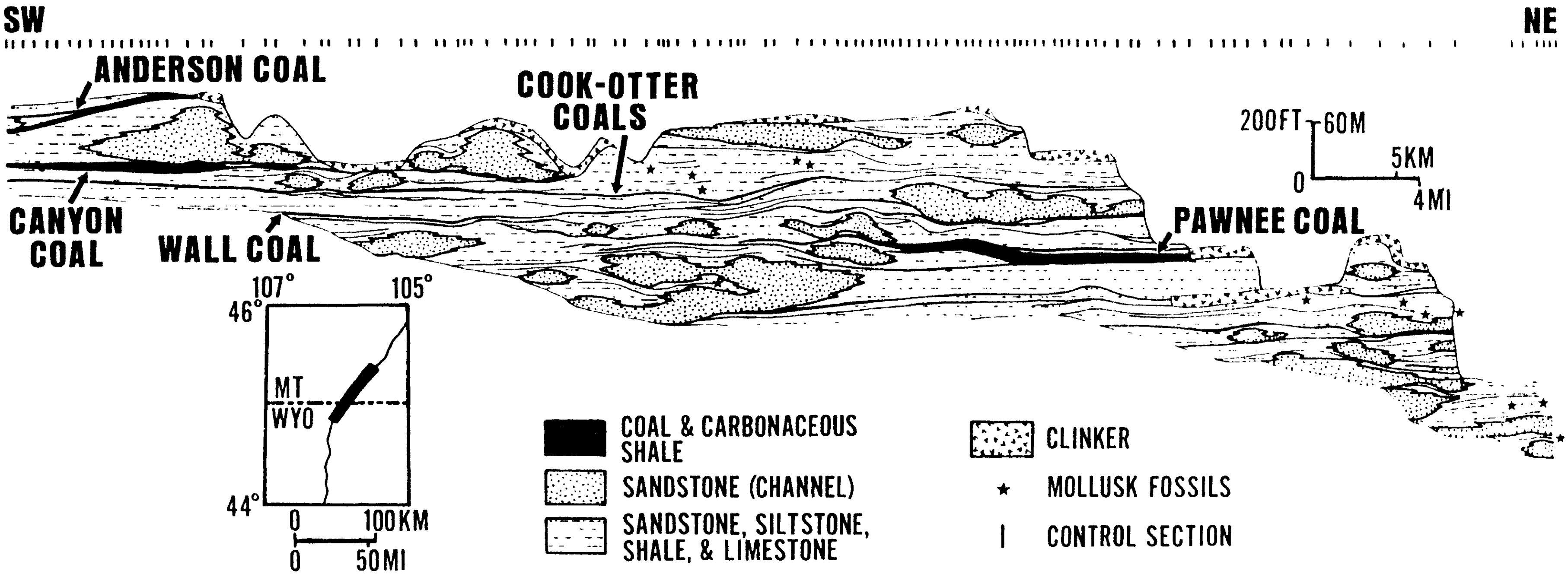


Figure 8.--Stratigraphic cross section showing facies variation of the lower part of the Tongue River Member of the Fort Union Formation. Adapted from Flores (1983).

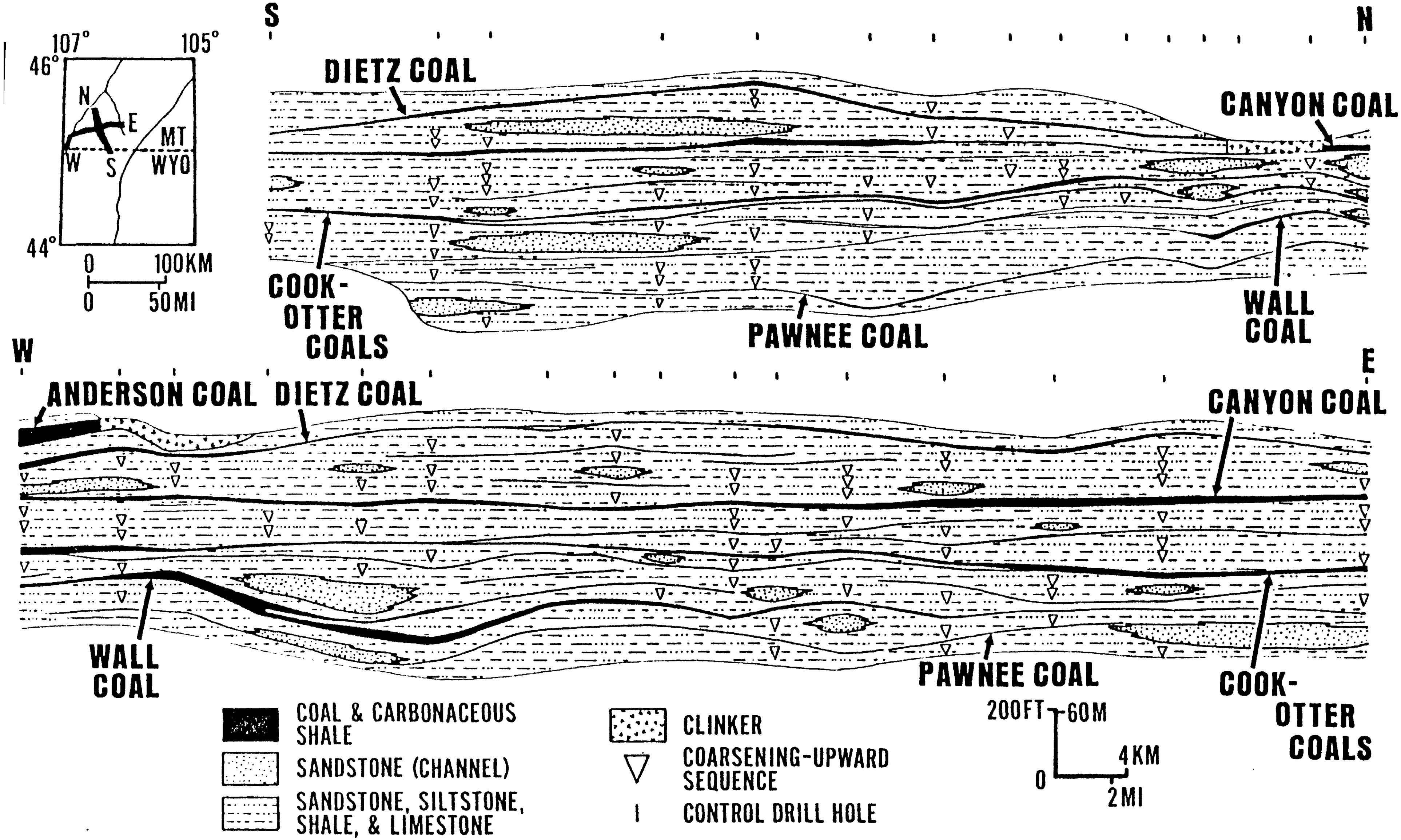


Figure 9.--Stratigraphic cross section showing three-dimensional facies variation of the lower part of the Tongue River Member. Adapted from Flores (1983).

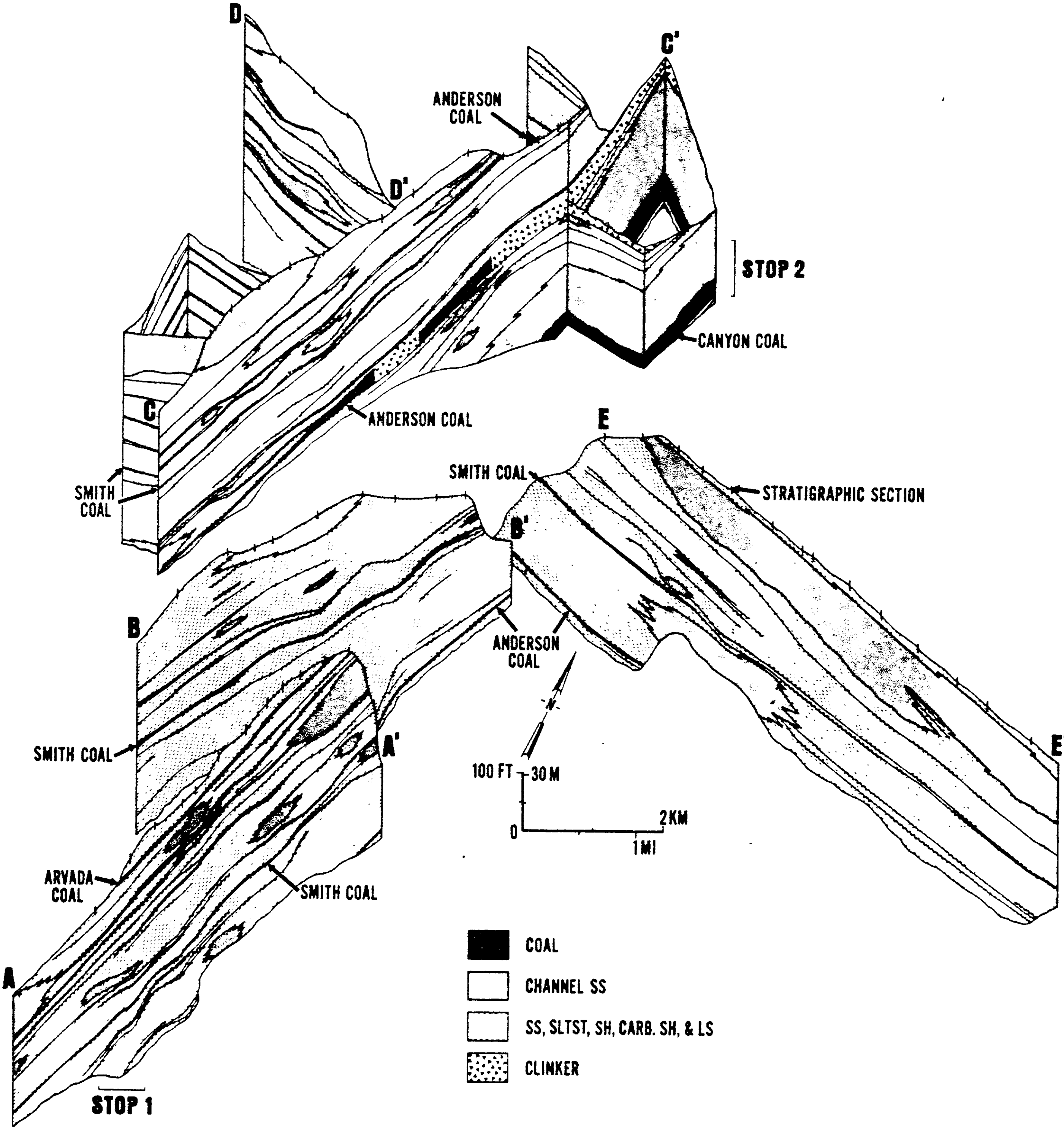


Figure 10.--Stratigraphic cross section showing facies variation of the anastomosed fluvial deposits of the upper part of the Tongue River Member. Adapted from Flores (1980).

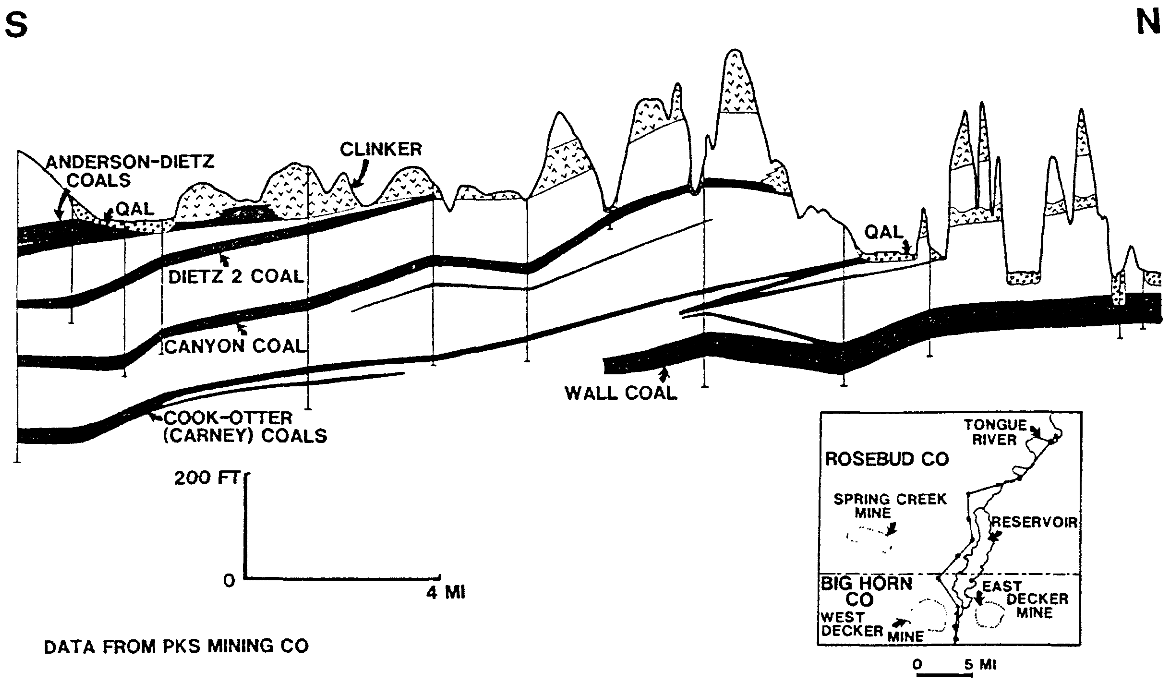


Figure 11.--Stratigraphic cross section showing thick coal deposits in the lower part of the Tongue River Member.

The effects of these controlling factors are shown in figure 12. The 20-ft isopach lines of the thick economic coals that formed west of the trunk rivers show directions of elongation of northwest-southeast, northeast-southwest, and north-south. The direction of elongation of uniformly thick coals may have been controlled by lineament structures, which probably developed in response to recurrent crustal movements. These movements resulted in subsidence affecting prolonged constructions of backswamps.

Anastomosed System

The lower part of the Tongue River Member of the Fort Union Formation is succeeded upward by facies of the anastomosed fluvial system. The facies relationships of the anastomosed fluvial deposits are shown in figure 10. The facies relationships are characterized by:

1. Coeval existence of closely-spaced channel-levee facies separated by floodplain-lacustrine facies and backswamp facies.
2. The channel sandstones consisting of multistory cut-and-fill festooned units.
3. Abundant freshwater fossiliferous zone (lacustrine facies) interdispersed with crevasse splay deposits.
4. Backswamp coal facies rich in carbonaceous shale.

A detailed description of the vertical variation of the channel-levee, floodplain-lacustrine, and backswamp facies of the anastomosed fluvial system is shown in figure 13. Figure 10, which shows the three-dimensional relationship of these facies, specifically displays the coeval nature of the channel sandstones in the uppermost part of the interval, a characteristic similar to those determined from recent anastomosed fluvial deposits of the Saskatchewan River by Smith and Smith (1980). A block diagram (fig. 14) shows a hypothetical texture and bed geometry of typical "Saskatchewan-type" anastomosed fluvial deposits as constructed by Smith and Smith (1980). The characteristics of the subenvironments and their deposits can be summarized as:

1. Straight channels aggrade vertically producing stringerlike deposits of coarse sediments.
2. Curved channels that slightly migrate may produce point-bar deposits.
3. The above types of deposits are unlikely to be laterally extensive.
4. Interchannel areas consist of organic and suspended sediments with peat accounting for as much as 30 percent of the sediments.
5. On the whole, overbank sediments in the anastomosed Saskatchewan River comprise 80-90 percent of the total accumulation of fluvial deposits.
6. The abundance of fine-grained overbank deposits and rooted plants are factors that have stabilized the overbank, which in turn, inhibited channel migration.
7. The Saskatchewan anastomosed river, located in a temperate well-vegetated environment, is in contrast to the Cooper Creek (Central Australia) which is an anastomosed fluvial system in an arid climate where mud-dominated, nonvegetated overbank succession played an important role in inhibiting channel migration (Rust, 1981).

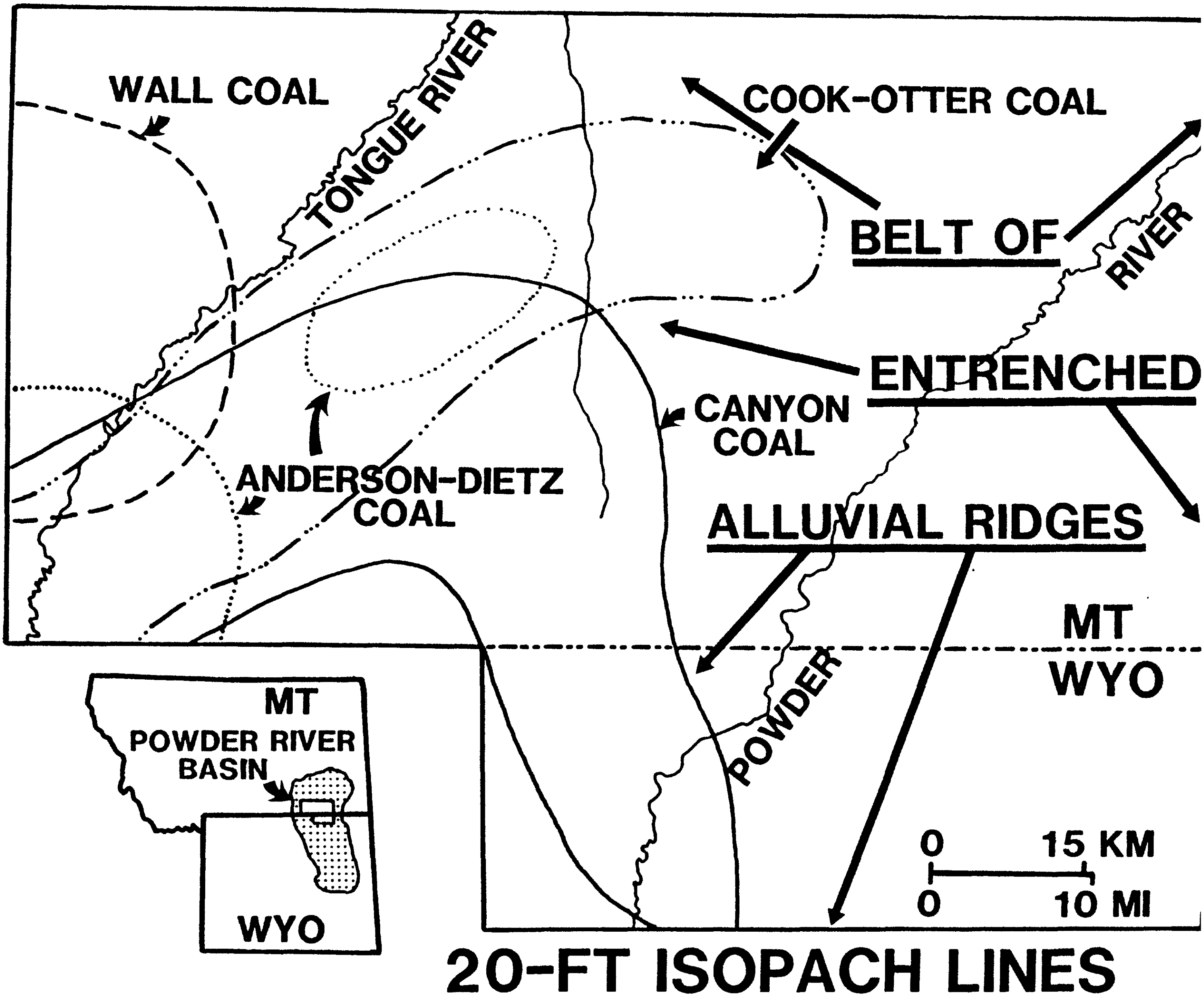


Figure 12.--Isopach map of the thick coals west of the alluvial meander belt.

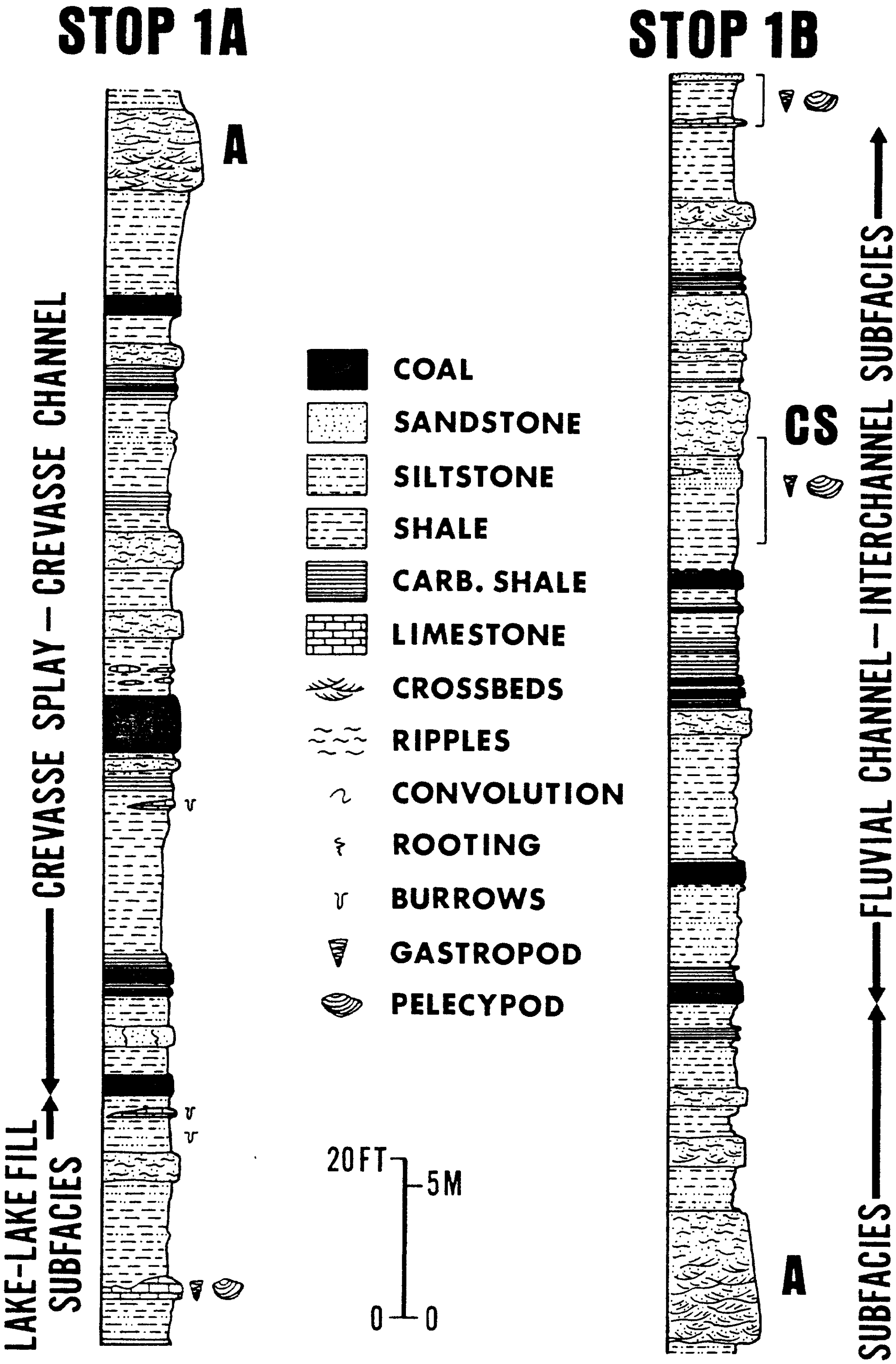


Figure 13.--Vertical facies variation of anastomosed fluvial deposits in the upper part of the Tongue River Member. Adapted from Flores (1980).

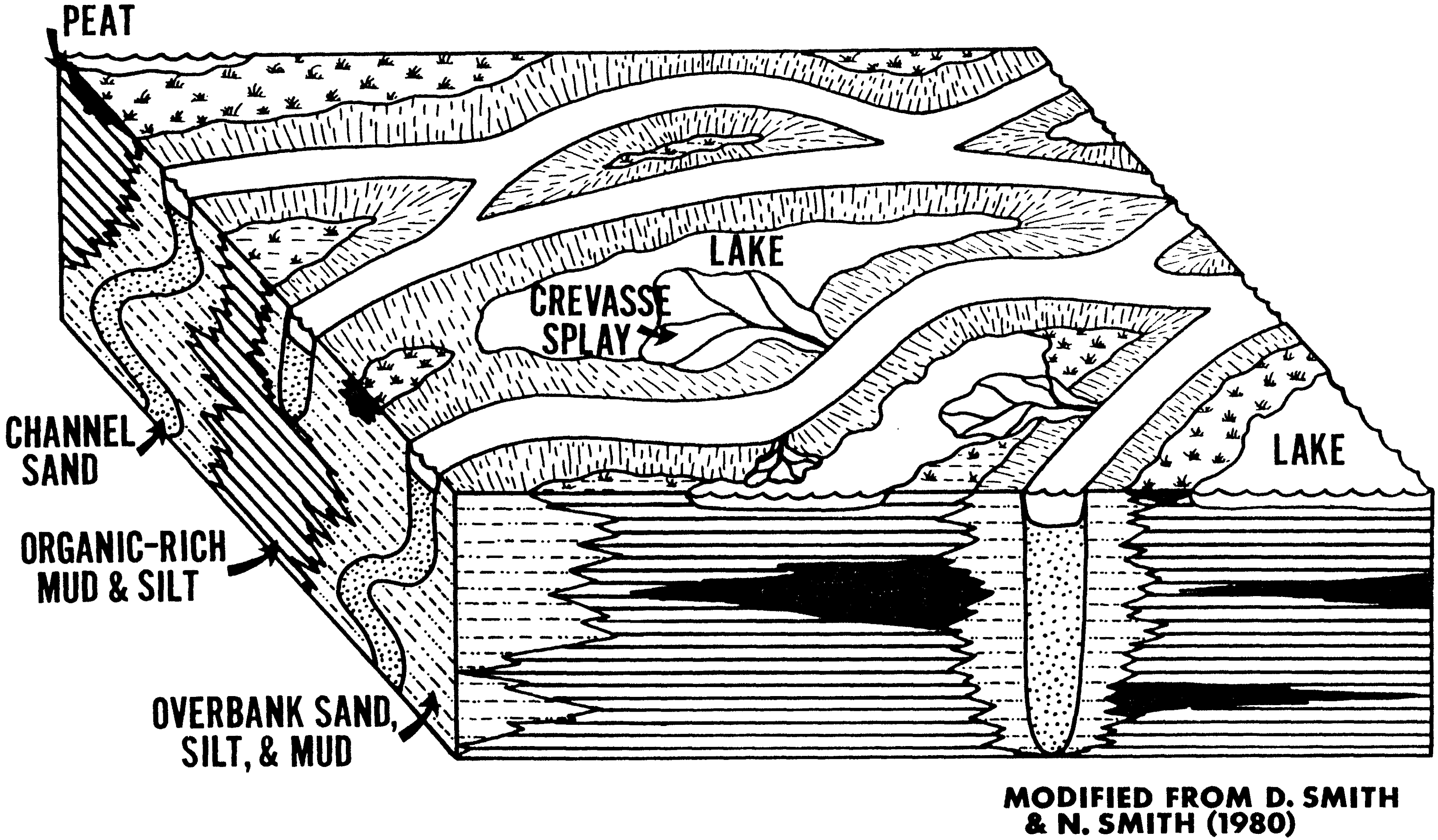


Figure 14.--Hypothetical vertical and lateral facies variation of the "Saskatchewan-type" anastomosed fluvial deposits. Adapted from Smith and Smith (1980).

So far, ancient deposits of anastomosed fluvial deposits have been generally unrecognized. However, isolated coarse channel deposits are associated with thin coals and organic-rich fine sediments such as those in the Permian-Carboniferous coal-bearing fluvioglacial and glaciolacustrine deltaic deposits in northeastern Karoo Basin (Miall, 1981). Recently, Smith and Putnam (1980), Putnam (1982), and Smith and Putnam (1982) described ancient anastomosed fluvial deposits with examples in the Lower Cretaceous upper Manville Group. The channel sandstones in this rock unit served as oil and gas reservoirs. Other anastomosed fluvial deposits in the coal-bearing Cretaceous basal Mesaverde Group near Rangely Dome in northwestern Colorado were described by Payne and Scott (1982).

COASTAL PLAIN-PIEDMONT FLUVIAL DEPOSITS

The Rocky Mountain Region contains numerous Tertiary and Cretaceous fluvial coal-bearing deposits that were formed in coastal plain and piedmont fluvial settings between epeiric and epicontinental seas at one end and Laramide uplifts at the other end. These fluvial settings are in contrast to those of the intermontane trunk-tributary fluvial setting in the Powder River Basin. These coastal plain-piedmont fluvial settings were drained primarily by short-headed streams that were less than 100 mi long. The short-headed stream drainages typified coastal plain-piedmont settings in the Tertiary-Cretaceous deposits in the Wasatch Plateau, Utah, Gallup sag, New Mexico, and Raton Basin, Colorado and New Mexico. These fluvial settings are in contrast to the coastal plain fluvial setting of Tertiary deposits landward of the epicontinental Cannonball sea southwest of the Williston Basin, which connected upstream to the trunk-tributary fluvial setting in the Powder River Basin. The proximity of these fluvial settings to the Laramide uplifts may have partly influenced the distribution of the coal deposits.

Wasatch Plateau

The economic coal deposits in the Wasatch Plateau are primarily produced from the Cretaceous Blackhawk Formation. The Blackhawk Formation grades upward into the Castlegate Sandstone, which consists of pebbly to granule-sized sandstones characterized by festooned and tabular crossbeds. The Castlegate probably represents deposits of braided streams in alluvial fan and distal braidplains (Fouch and others, 1983). The coarse-grained Castlegate Sandstone represents synorogenic deposits related to thrust belts along the eastern part of the Sevier orogenic belt to the west (Fouch and others, 1983). The Star Point Sandstone, which conformably underlies the Blackhawk Formation, is interpreted as coastal sand deposits of distributary mouth bar-channels that grade laterally in barrier-shoreface deposits (Marley and others, 1979; Flores and others, 1982). The Star Point Sandstone grades downward into the prodelta-offshore silty shale deposits of the Blue Gate Member of the Mancos Shale. The environmental settings arranged in vertical stratigraphic associations are shown in figure 15.

The most important economic coal deposit of the Blackhawk Formation is the Hiawatha coal which formed in two swamp settings in the marginal coastal environments:

1. Swamps formed on filled surfaces of lagoon and back-barrier areas.
2. Delta plain that is supported by major distributary channels that crosscut the lagoons.

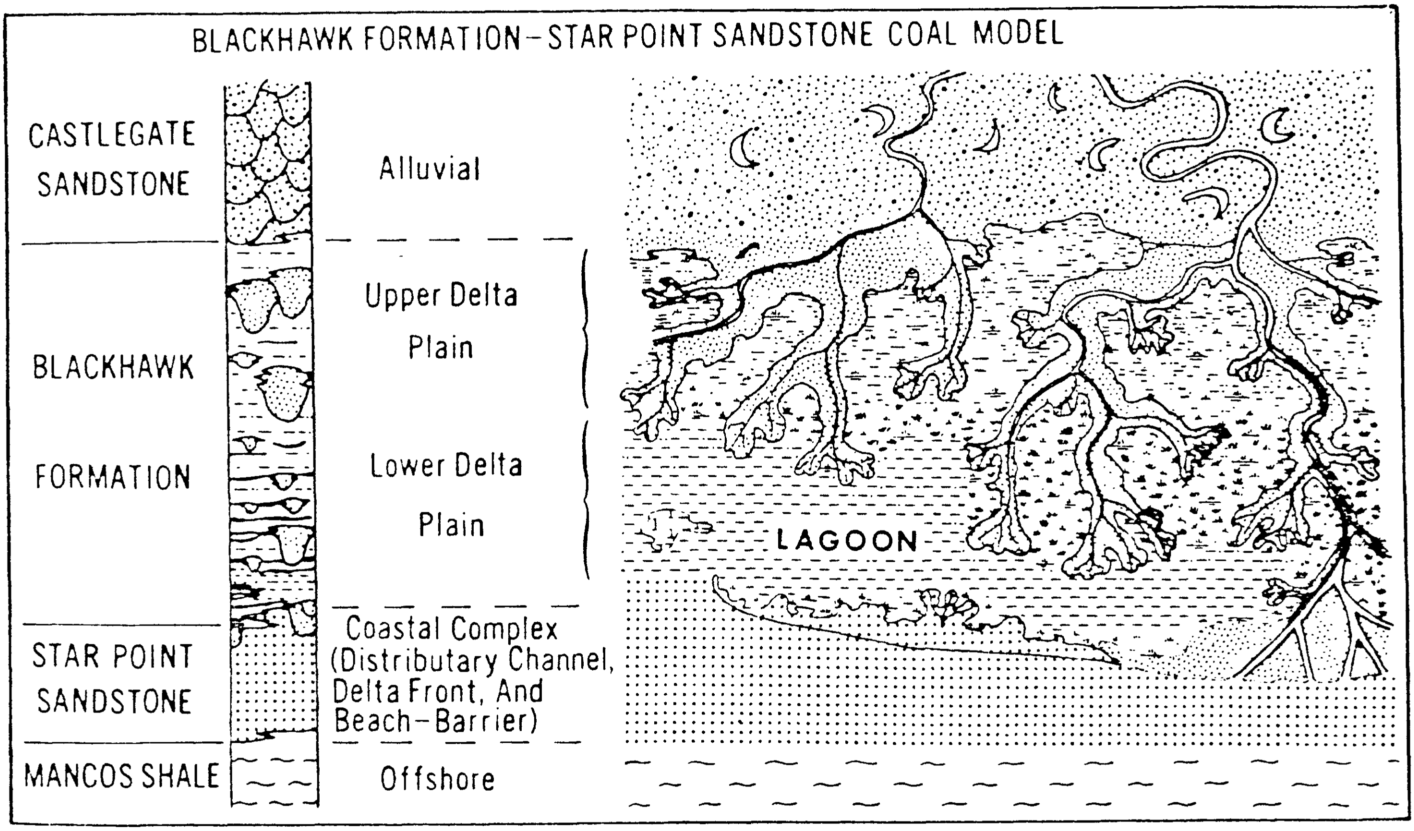


Figure 15.--Vertical stratigraphic-environmental variations of the Star Point Sandstone, Blackhawk Formation, and Castlegate Sandstone. Adapted from Marley and others (1979).

The Hiawatha coal accumulated in these settings is as much as 15 ft in thickness and elongate in shape, with uniformly thick coal parallel to subparallel to the paleoshoreline or length of the barrier.

Landward of the barrier and major delta-plain settings are coastal plain fluvial settings that grade farther landward into piedmont fluvial environments. In the coastal plain fluvial settings, coals primarily formed in floodbasins between major fluvial systems and short-headed streams that drained the environment behind lagoons and in interchannel areas of short-headed streams. The distribution of these floodbasin and interchannel coals is shown in figure 16.

The fluvial floodbasins in the more landward setting appear to lose their importance as coal-forming environments. The presence of thin coals with abundant interbedded carbonaceous shale, as well as predominance of carbonaceous shale over other organic deposits, suggest that the floodbasins may have received abundant detritus that choked the swamps. This condition is probably prevalent in areas near the piedmont setting close to the uplifts, which formed alluvial fans and braidplains (Castlegate settings).

Raton Basin

Upper Cretaceous and Paleocene strata in the Raton Basin, New Mexico, represent a major episode of clastic progradation into the epeiric sea of the Pierre Shale. Major depositional settings sequentially include: regressive sandy coast (Trinidad Sandstone), backstrand peat swamps with small channels (Vermejo Formation), peat-rich interfluvial swamps and large fluvial channels (Raton Formation), and well-drained braidplains (Poison Canyon Formation). The fluvial coal deposits in the Raton Basin are mainly contained in the upper part of the Vermejo Formation and the overlying Raton Formation. The Raton Formation is overlain by the Poison Canyon Formation, which is barren of coal deposits and is mainly composed of thin, narrow, coarse-grained channel sandstone and reddish-green, fine-grained overbank shale, siltstone, and silty sandstone. The Poison Canyon Formation probably represents anastomosed, low-sinuosity channel and interchannel deposits. The Raton Formation is dominated by sandstones consisting of channel, abandoned channel fill, levee, and splay feeder deposits. Adjacent floodbasins consist of crevasse-splay sandstones and siltstones, and thick sequences of fine-grained deposits. The latter include bioturbated mudstones and siltstones of well-drained floodbasins, and coals and carbonaceous shales of poorly drained swamps. Thick coals ($\geq$5 ft) appear to be associated with and split by splay deposits marginal to channels. These coals have channel sandstone as roof rock. The basal Raton Formation is locally gradational into the underlying Vermejo Formation. Less developed floodbasin deposits, thin coals (<1 ft), and extensive fluvial channel sandstones characterize the upper 200 ft of the Raton Formation which conformably underlies the Poison Canyon Formation. The Vermejo Formation consists mainly of coal-bearing deltaic deposits (river-dominated) underlain by delta-front deposits (distributary channel, mouth bar, and abandonment facies) of the Trinidad Sandstone, which in turn grade downward into the prodelta deposits of the Pierre Shale.

In comparison with the Cretaceous Blackhawk Formation in the Wasatch Plateau, the fluvial coals in the Raton Formation in the Raton Basin are thick (as much as 11 ft). Coal swamps in the Raton Formation, like those in the Blackhawk Formation in the Wasatch Plateau, were developed in fluvial

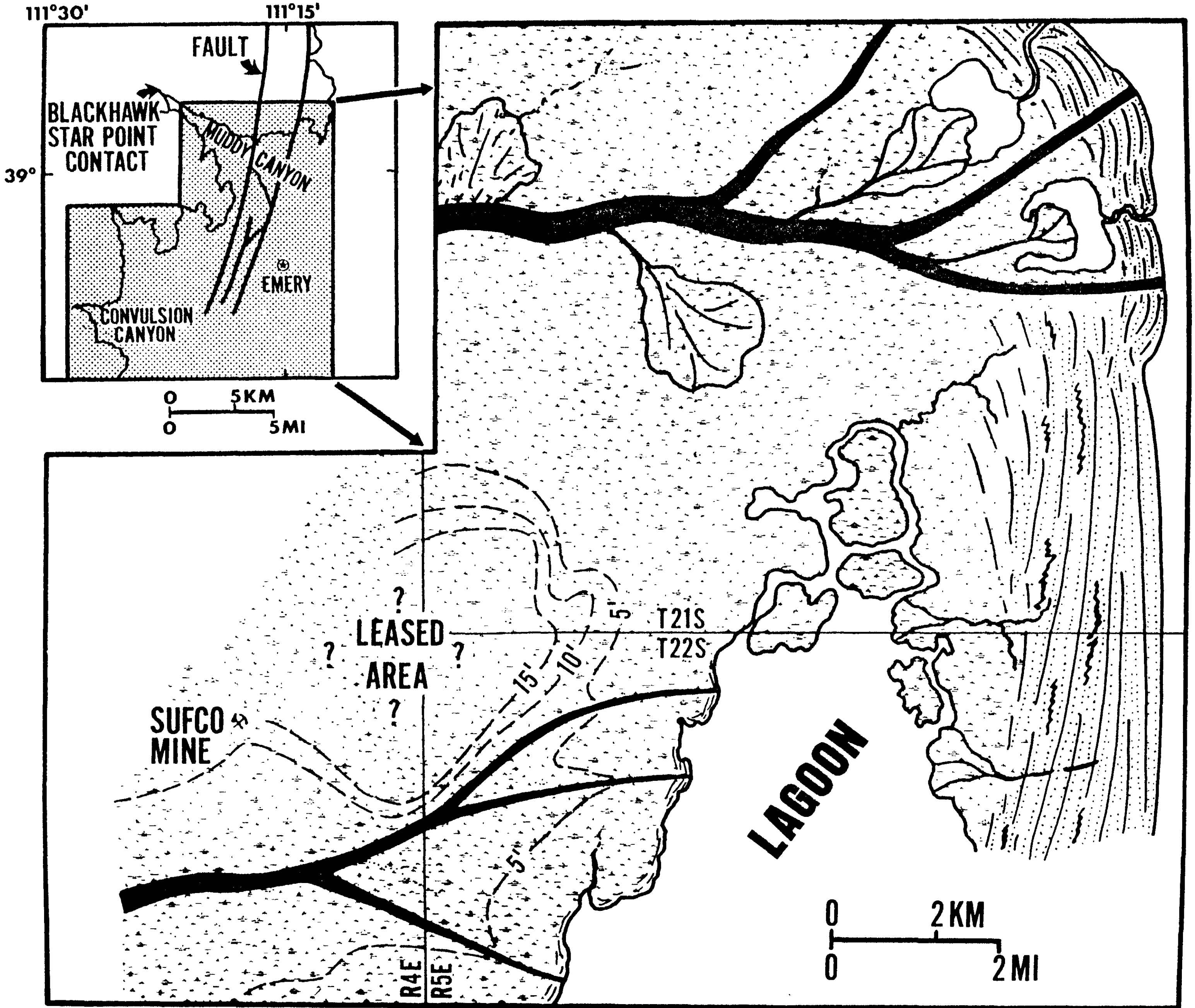


Figure 16.--Floodbasin swamp that accumulated coals between a short-headed stream and a major distributary channel that crosscut a lagoon in the Blackhawk-Star Point environmental setting.

floodbasins associated with major meandering fluvial channels and short-headed streams. However, in the Raton Formation the floodbasin swamps accumulated thick coals and were far removed from detrital influx. This condition developed despite the close proximity of the fluvial system to the Laramide Sangre Cristo uplift to the west-southwest. Thus, in the Raton Basin, the coal-forming environments are best developed in:

1. Deltaic plain swamps (related to river influenced, high destructive delta system).
2. Alluvial plain swamps (related to large meander belts).

The deltaic plain swamps formed in the Vermejo Formation were developed in two major marginal delta-plain settings (Flores and Tur, 1982):

1. Interdistributary swamps.
2. Swamps over abandoned subdeltas.

The interdistributary coal swamps formed thick organic deposits that are elongated subparallel to distributary channels. The swamps formed on abandoned subdeltas also formed thick coal deposits but are more laterally extensive compared to the interdistributary coals.

Gallup Sag

In the Gallup sag, the coal-bearing strata of the Mesaverde Group consists of the Gallup Sandstone, Crevasse Canyon Formation, and Menefee Formation. They were deposited following earlier marine transgressions reflected by tongues of the Mancos Shale (Sears and others, 1941). The Gallup Sandstone, which locally contains economic coals, grades upward into the coal-bearing Dilco Member of the Crevasse Canyon Formation. These strata are overlain in the Gallup coal field by the barren Bartlett Member of the Crevasse Canyon Formation, which in turn grades upward and laterally into the undifferentiated coal-bearing interval of the Gibson Coal Member of the Crevasse Canyon Formation and the Cleary Coal Member of the Menefee Formation. The Cleary Coal Member grades upward into the barren Allison Member of the Menefee Formation. The undifferentiated Gibson and Cleary Coal Members are split to the northeast by the pinchout of the merged deposits of the Point Lookout Sandstone and its Hosta Tongue. Beaumont and others (1956) identified the coal-bearing deposits below and above the Point Lookout Sandstone as the Gibson Coal Member of the Crevasse Canyon Formation and the Cleary Coal Member of the Menefee Formation, respectively. Most of the coal mining occurs southwest of the Point Lookout Sandstone pinchout and in the undifferentiated interval that crops out northwest, north, and northeast of Gallup, N. Mex.

Our study of Upper Cretaceous coals was concentrated in the Gallup coal field in the Gallup sag in the southwest San Juan Basin. The environments of deposition of coal in these coal fields range from fluvial, wave-reworked delta, and barrier systems in the Gallup Sandstone (Hohman and Flores, 1981) to alluvial, delta-plain, lagoon, and barrier-back barrier systems in the Crevasse Canyon and Menefee Formations (Cavaroc and Flores, 1980, 1981).

Gallup Sandstone

The Gallup Sandstone exposed in the western part of the Gallup coal field is interpreted as a vertical continuum of delta, barrier, and fluvial deposits. The delta plain flanked by barrier complexes (fig. 17) served as sites of accumulation of economic coal beds. In contrast, coal beds formed in the fluvial environment are thin, lenticular, and of no economic consequence.

The Gallup Sandstone can be divided vertically into three major facies:

1. Delta-front barrier shoreface.
2. Delta-plain back barrier.
3. Fluvial facies.

The delta-front barrier shoreface facies, in the lower part of the Gallup Sandstone, is a coarsening-upward sequence that consists of interbedded sandstone, siltstone, and shale grading upward into a thick sandstone. The delta-front facies is characterized by lenticular-shaped sandstones with abundant festoon crossbeds in the upper part. In addition, bioturbation in the delta-front facies is rhythmically occurring and is concentrated in the upper parts of the sandstone beds. Ball-and-pillow structures are also common in the delta-front facies. The barrier shoreface facies is recognized by its sheetlike shape, common presence of horizontal and ripple laminations, vertically stacked planar crossbeds, and the nature of the bioturbation of the sandstone beds, which are intensely burrowed throughout. The delta-front facies is locally scoured by fluvially influenced distributary channels and the barrier shoreface is locally dissected by flood-and-ebb influenced tidal inlets. The delta-front facies grades laterally imperceptibly into the barrier shoreface facies as a result of reworking by waves.

The delta-front and barrier shoreface facies are overlain by delta-plain and back-barrier facies, respectively, in the middle part of the Gallup Sandstone. The back-barrier facies consists of heavily bioturbated lagoonal and washover sandstone, siltstone, shale, carbonaceous shale, and coal in the lower part. These deposits are replaced upward and laterally by delta-plain sandstone, siltstone, shale, carbonaceous shale, and coal. Although both facies are coal bearing, the delta-plain facies appears to be more important in controlling in the occurrence of coal than the back-barrier facies. That is, the delta-plain facies contains more numerous coal beds, although more discontinuous than in the back-barrier facies. The coal beds are as much as 3.5 ft thick and laterally traceable in outcrops to as far as 4 mi. Coal settings in the back-barrier area were in swamps marginal to lagoons. These lagoons infilled by deposits of washover, flood tidal deltas, and tidal channels formed platforms that became sites for the expansion of the swamps. In contrast, the delta-plain coals formed in swamps marginal to interdistributary bays that were filled by crevasse splays.

The upper part of the Gallup Sandstone includes deposits of meandering to low-sinuosity streams. These fluvial deposits consist mainly of numerous stacked, overlapped, varicolored, pebbly channel sandstones. The channel sandstones in the lower part of the fluvial facies are overlapped, lenticular bodies that grade laterally and vertically into overbank-floodplain sandstone, siltstone, shale, carbonaceous shale, and a few discontinuous dirty coal beds. These interchannel deposits are isolated, indicating deposition in the

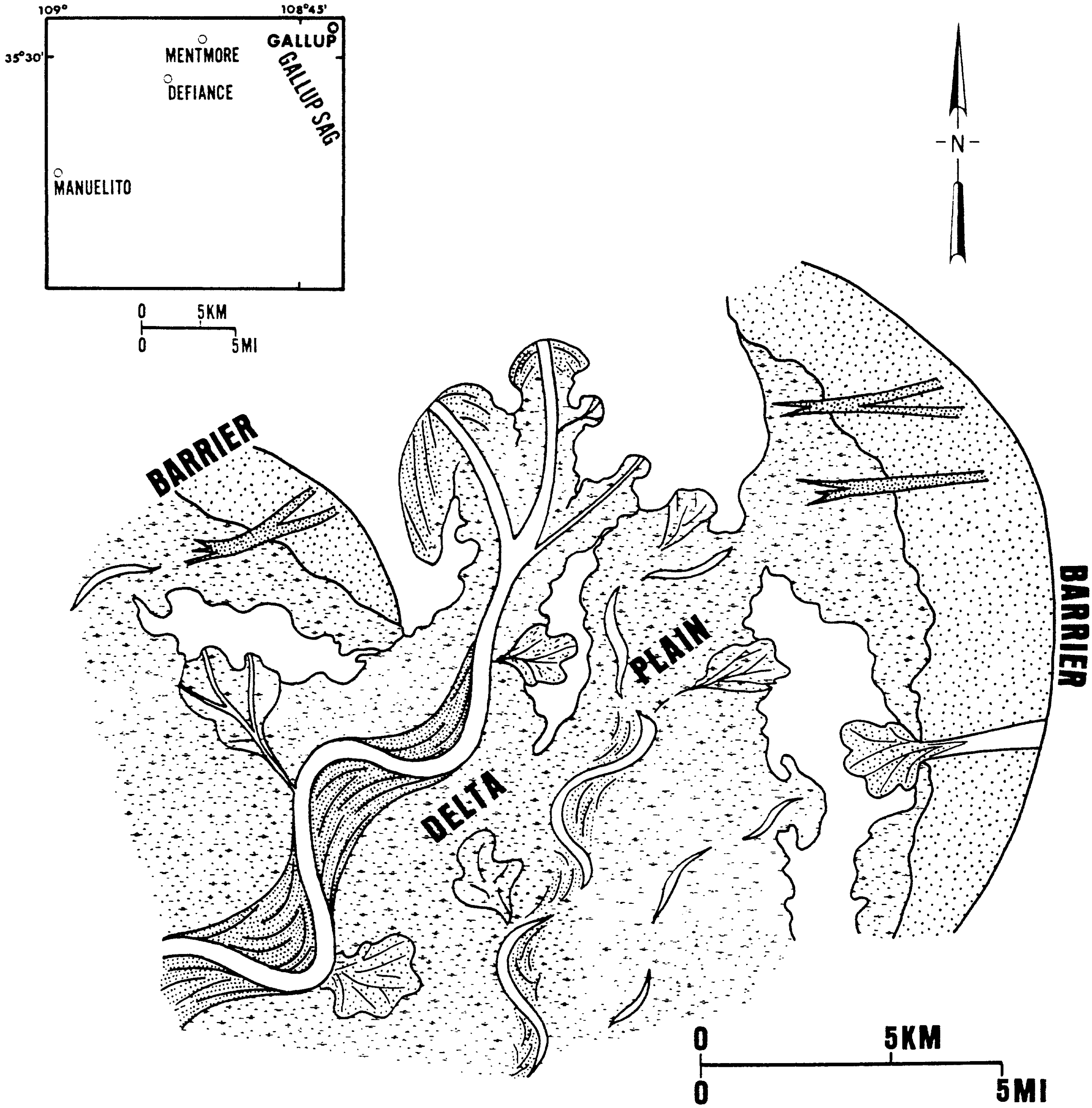


Figure 17.--Reconstruction of paleogeographic setting of the Gallup Sandstone in the Gallup sag. Adapted from Hohman and Flores (1981).

alluvial plain by high-sinuosity streams. The channel sandstones in the upper part of the fluvial facies comprise laterally merged and multiple stacked deposits that display blanketlike shape. The channel sandstones so dominate this part of the fluvial facies that little or no interchannel deposits are preserved. This is probably a result of deposition by low sinuosity or braided streams. The multiple stacking of the channel sandstones may indicate scour and fill within nonmigrating channels.

Crevasse Canyon and Menefee Formations

The most important coal-bearing interval of the Crevasse Canyon and Menefee Formations is the undifferentiated Gibson (Crevasse Canyon) and Cleary (Menefee) Coal Members. These coal members contain economic coals whose occurrence is controlled by their environments of deposition. Major coal depositional settings change from well-drained alluvial, through delta plain and open lagoon, into reworked barrierlike complex of the Point Lookout Sandstone and its Hosta Tongue within a 25-mi distance (fig. 18). These depositional changes are recorded in a southwest to northeast direction from the Mentmore area (northwest of Gallup), through the Gibson Canyon area (north of Gallup), to the Coal Mine Canyon area (northeast of Gallup), respectively.

Mentmore Area

The undifferentiated Gibson and Cleary Coal Members in the Mentmore area grade laterally into the Bartlett Barren Member of the Crevasse Canyon Formation; the latter member is characterized by scarcity of coal and a high proportion of channel sandstones to fine-grained detritus.

The Bartlett Barren Member contains about 70 percent sandstone, the bulk of which is lenticular channel sandstones. These sandstones are mainly multiple stacked and basally conglomeratic, and grade laterally and vertically into interbedded sandstone, siltstone, shale, carbonaceous shale, and a few thin coals. The Bartlett Barren Member probably represents active predominantly well-drained alluvial floodplain deposits. The channel sandstones were not deposited in freely meandering channels but rather in channels that shifted due to avulsion. Small channel sandstones represent transient crevasse feeder channel deposits that originated from contemporaneous major channels. The fine-grained detritus juxtaposed with the channel sandstones represent well-drained, overbank-floodplain deposits. Reducing environments within the alluvial floodplain appear to have continually received detritus from adjacent streams, resulting in the accumulation of carbonaceous shale and dirty coal beds.

The coal-bearing Gibson-Cleary interval in the Mentmore area, locally merging with the Bartlett Barren Member, consists of interbedded channel sandstone and overbank-floodplain sandstone, siltstone, shale, carbonaceous shale, and coal. Carbonaceous shale and coal beds commonly associated with the fine-grained deposits are as much as 7.5 ft thick and are laterally traceable in outcrops as far as 2 mi. The channel sandstones maintain the lenticular geometry and sequence of internal structures of those in the Bartlett Barren Member; however, they are not as common as in the Bartlett Barren Member. The depositional setting of the undifferentiated Gibson and Cleary Coal Members is interpreted as a portion of the Bartlett alluvial plain partly abandoned by major channel activity. Waning of fluvial activity,

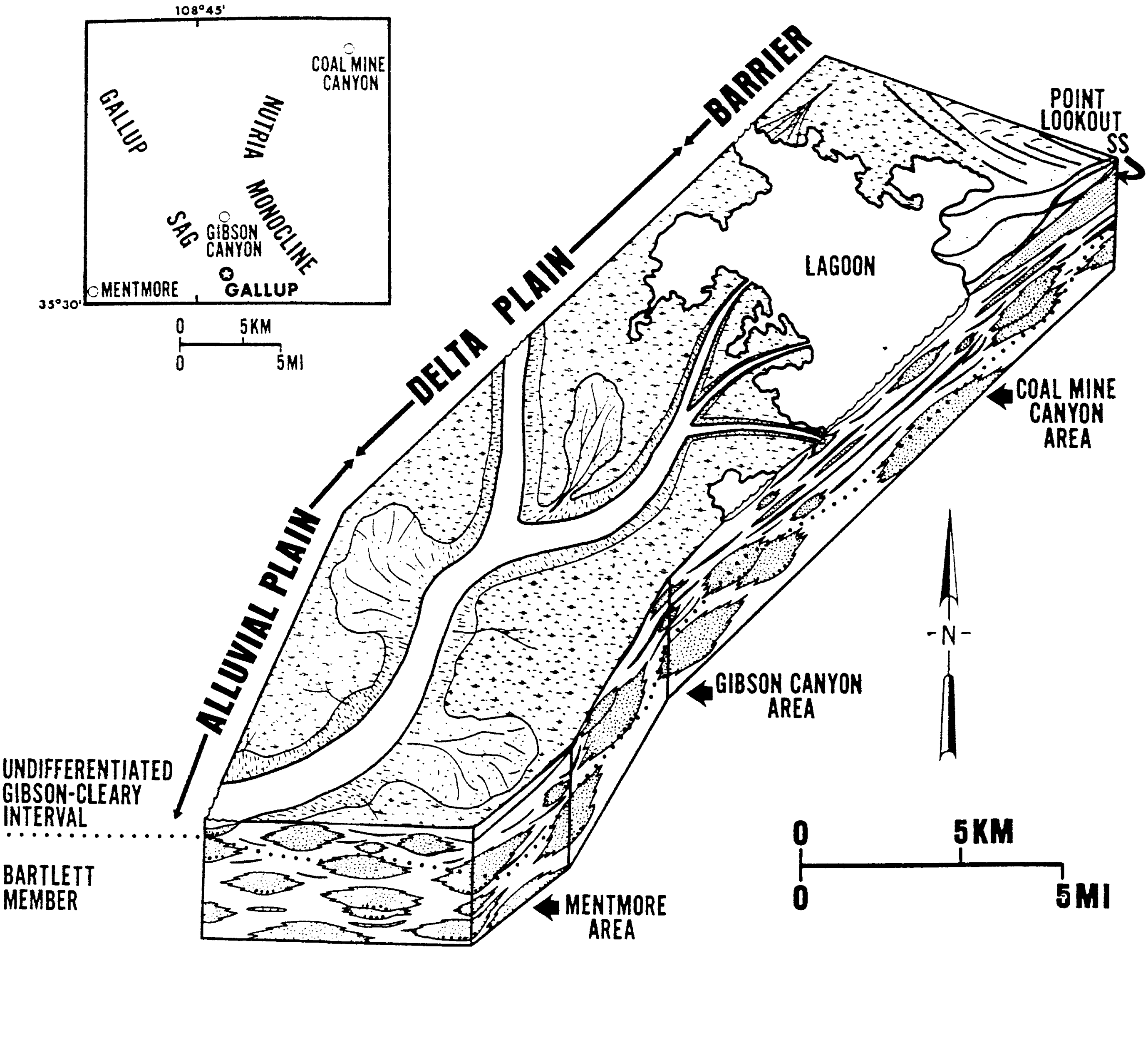


Figure 18.--Paleogeographic reconstruction of coal-forming depositional environments in the Crevasse Canyon, Menefee, and Point Lookout Formations. Adapted from Cavaroc and Flores (1981).

characterized by minimal detrital influx, directly influenced the accumulation of thick peat in the poorly drained backswamps that followed shift and abandonment of channel activity.

Gibson Canyon Area

The undifferentiated Gibson and Cleary Coal Members in the Gibson Canyon area are underlain by the Bartlett Barren Member and, in general, have similar rock types as those in the Mentmore area. However, some pronounced differences in lithogenetic types exist. The crevasse-splay sandstones are more abundant in the Gibson Canyon area. This sequence is also commonly overridden by small channel sandstones. The second major difference is the presence of burrowed zones, a feature not found in the Mentmore area. Burrows are intense and are found in clean sandstones interbedded with carbonaceous siltstones as well as in the crevasse-splay sandstones. As at Mentmore, the lowest coal beds of the Gibson and Cleary Coal Members pass laterally into the Bartlett Barren Member. However, the major coal beds of the undifferentiated coal-bearing interval are thick, laterally extensive (on the order of several miles), and associated with widespread carbonaceous shale zones.

The thick, laterally continuous coals of the undifferentiated Gibson and Cleary Coal Members in the Gibson Canyon area are interpreted as accumulations over embayed, drowned surfaces of earlier alluvial deposits. The bifurcation of the primary fluvial channels into distributaries suggests that the embayed areas represent interdistributaries on a delta plain. That these bays were built up into poorly drained swamp platforms is indicated by the associated intensely bioturbated sequences.

Coal Mine Canyon Area

In the Coal Mine Canyon area, the Gibson and Cleary Coal Members are split by the landward pinchout of the merged deposit of the Point Lookout Sandstone and its Hosta Tongue.

Deposits of the Cleary Coal Member, which overlie the Point Lookout Sandstone, consist of thin, rippled, and burrowed sheet sandstones associated with carbonaceous shales. These deposits give way to small lenticular channel sandstones associated with coarsening-upward detritus. Coals associated with these deposits are thin (less than 1 ft) and discontinuous. The coal beds are as much as 6.5 ft in thickness and laterally are more continuous in the upper part of the interval.

The Gibson Coal Member beneath the Point Lookout Sandstone contains sheetlike rippled and burrowed sandstones that are interbedded with small lenticular channel sandstone bodies. Carbonaceous shales contain intense bioturbation and are interbedded with clean silty sandstone and carbonaceous siltstone. Coal beds are as much as 13 ft in thickness and are laterally discontinuous. The coal beds in the lower part of the Gibson Coal Member pinch out and become discontinuous in the Bartlett Barren Member.

The merged deposit of the Point Lookout Sandstone and its Hosta Tongue in the Coal Mine Canyon area provided a physical barrier that isolated carbonaceous lagoonal deposits of the Gibson Coal Member, which in turn formed across abandoned fluvial facies of the Bartlett Barren Member. The barrier sandstone complex was oriented from northwest to southeast. The abundant southwesterly dips of the crossbeds and intense bioturbation of tops of the sandstones indicate deposition in back-barrier washover and/or flood tidal delta environments. Lagoonal conditions at the back of the barrier complex were subsequently destroyed by northerly influx of detritus beginning in the Cleary time. Reintroduction of deltaic to alluvial activities seaward behind the regressive Point Lookout Sandstone permitted optimum conditions for accumulation of thick coal in the late Cleary time.

Williston Basin

The coastal plain fluvial deposits in the southwest Williston Basin are mainly contained in the Paleocene Tongue River Member of the Fort Union Formation. This rock unit grades downward into the coal-bearing Ludlow Member of the Fort Union Formation that contains tongues of the marine Cannonball Member of the Fort Union Formation. The deltaic deposits of the Ludlow Member were described by Belt and Flores (1982) and the fluvial deposits of the Tongue River Member were described by Warwick, Flores, and Cavaroc (1982).

The Ludlow Member represents deltaic deposits marginal to the epicontinental Cannonball sea. These deposits were accumulated in a high constructive delta which is characterized by temporal distributary channel and crevasse splay-interdistributary swamp facies. The interdistributary swamps accumulated as much as 3-ft-thick coal deposits that are laterally discontinuous. The distributary channels, which averaged 2 mi wide, switched laterally within a 5-mi-wide belt. Subdelta environments were developed and avulsion occurred within this belt. Thus, in this setting coals accumulate as thick and laterally extensive beds. Coal deposits as much as 12 ft thick accumulated under this condition. The abandoned subdelta swamps eventually were destroyed by brackish-marine incursions due to compaction and autocyclic shifts. The brackish-marine incursions mark the limit of Cannonball sea transgression into the southwestern Williston Basin.

The delta-plain deposits of the Ludlow Member are succeeded upward by alluvial-plain deposits of the Tongue River Member. The alluvial channels, which averaged 4.5 mi wide, formed 15-mi-wide meanderbelts. Avulsion of meanderbelts created floodbasins, 10-20 mi across, in which crevasse splays interacted with freshwater lakes and swamps. The floodbasin swamps, where not choked by crevasse splays or drowned by lakes, accumulated as much as 28-ft thick lignite, which can be traced laterally in outcrops for 15 mi prior to going into the subsurface. Subsurface data suggest that these coal swamps are as much as 25 mi wide and more than 90 mi long (Rehbein, 1977; Lewis, 1979).

The fluviodeltaic deposits of the Fort Union Formation in the southwest Williston Basin probably represent the downflow continuum of the trunk-tributary fluvial system in the Powder River Basin (Flores, 1980, 1983; Winczewski and Groenewold, 1982). The fluvial setting is similar to that suggested by Galloway (1979) in which the intermontane riverine plain continued downstream into a fluvial coastal and deltaic plain (figs. 4 and 19).

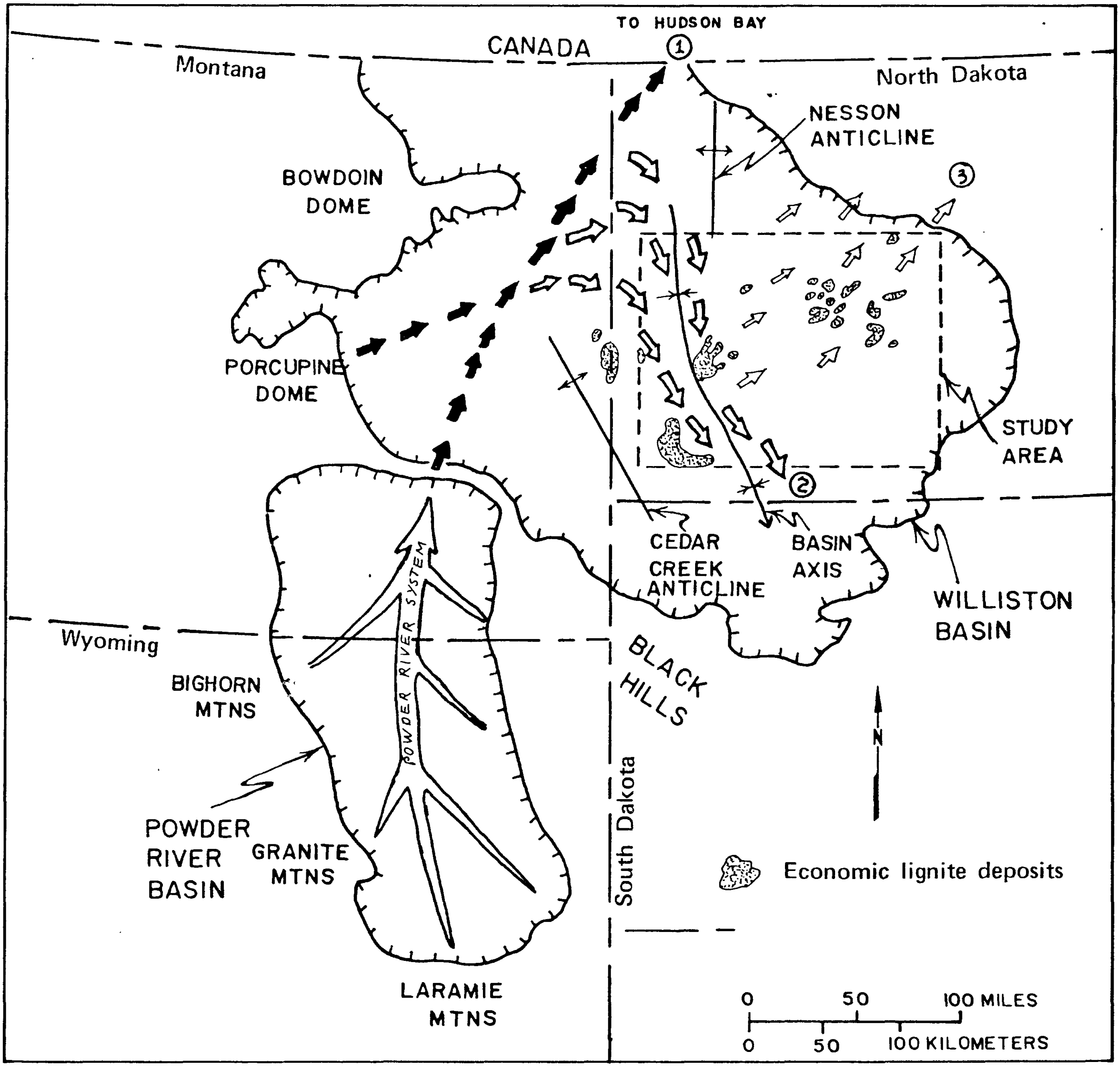


Figure 19.--Regional patterns of fluvial systems during the Paleocene and their relationship to structural features in the Powder River Basin and Williston Basin. Adapted from Winczewski and Groenewold (1982).

GLACIOFLUVIO-DELTAIC DEPOSITS

The Permo-Carboniferous Itarare Formation in the northeastern part of the Parana Basin, State of Sao Paulo, Brazil, exemplifies coals deposited in continental ice-influenced environments. The Itarare Formation is exposed along a belt 185 mi long and 85 mi wide and is bounded by Precambrian and Devonian rocks on the southeast and by Permian and Mesozoic rocks on the north (fig. 20). The coal deposits in the Itarare Formation occur in the Buri, Cirquilho, and Monte-Mor coal fields. The Itarare Formation, which is as much as 4,265 ft in thickness, grades laterally into the Aquidauana Formation (fig. 21) that is exposed to the northeast of the outcrop belt (northeast of the crosshatch in fig. 20). The Aquidauana Formation consists of periglacial deposits and the Itarare Formation contains proglacial deposits. The periglacial deposits of the Aquidauana Formation are composed of coarse detritus of alluvial fan-flow till. The proglacial deposits of the Itarare Formation consist of generally coarsening-upward sequences of glaciomarine, glaciodeltaic, and glaciofluvial deposits of shales, siltstones, sandstones, coals, and conglomerates (fig. 22). The proglacial interval includes glaciomarine deposits in the lower part and glaciofluvial deposits in the upper part, and average 650 m in thickness. The glaciomarine deposits comprise prodeltaic, brackish-marine, brachiopod-bearing shales, siltstones, and "dropstones." The "dropstones" are ice-rafted granitic, metamorphic, and diabasic rock fragments that range from pebble to boulder size. The prodeltaic deposits grade upward into the glaciodeltaic deposits that are composed of delta-front shales, siltstones, sandstones, and "dropstones." The delta-front deposits are characterized by abundant synsedimentary deformation structures in the form of large scale diapirs, slumps, and microfaults. These synsedimentary deformation structures attest to the high fluidity of sediments and high density sedimentation in glacial settings. The high density sedimentation in the delta front resulted in turbidite and submarine fan deposits.

The coal deposits in the proglacial sequence are formed mainly in swamps in the glaciodelta-plain and glaciofluvial environments. However, the swamps in the glaciodelta plain are better developed than those in the glaciofluvial environment. The coal beds are thin, as much as 2.5 ft in thickness, and are discontinuous, lenticular-shaped bodies. These Gondwanan coals typically contain Glossopteris and Gangamopteris flora, high ash content varying from 21 to 75 percent, and high sulfur content varying from 0.3 to 10.5 percent (analysis from outcrop samples). They range from subbituminous to bituminous with heating values varying from 3,000 to 5,700 cal/gm. The coals were principally accumulated in interdistributary swamps that were frequently suffocated by crevasse splays. The crevasse-splay deposits are coarsening-upward sequences of shales, siltstones, and sandstones. The lower part of the sequence is highly bioturbated and the upper part is ripple laminated. However, unlike their nonglacial counterparts, these crevasse-splay deposits contain "dropstones" that probably resulted in deposition from ice rafting into the crevasse splays during floods. The interdistributary swamps were probably drowned by local subsidence due to compaction of underlying sediments and/or eustatic rise of sea level due to deglaciation. This condition may have developed interdistributary bays in which crevasse splays prograded and infilled, which in turn, served as a swamp platform upon abandonment. The interdistributary deposits grade laterally into interdistributary channel sandstones that contain lag deposits of granitic, metamorphic, and diabasic

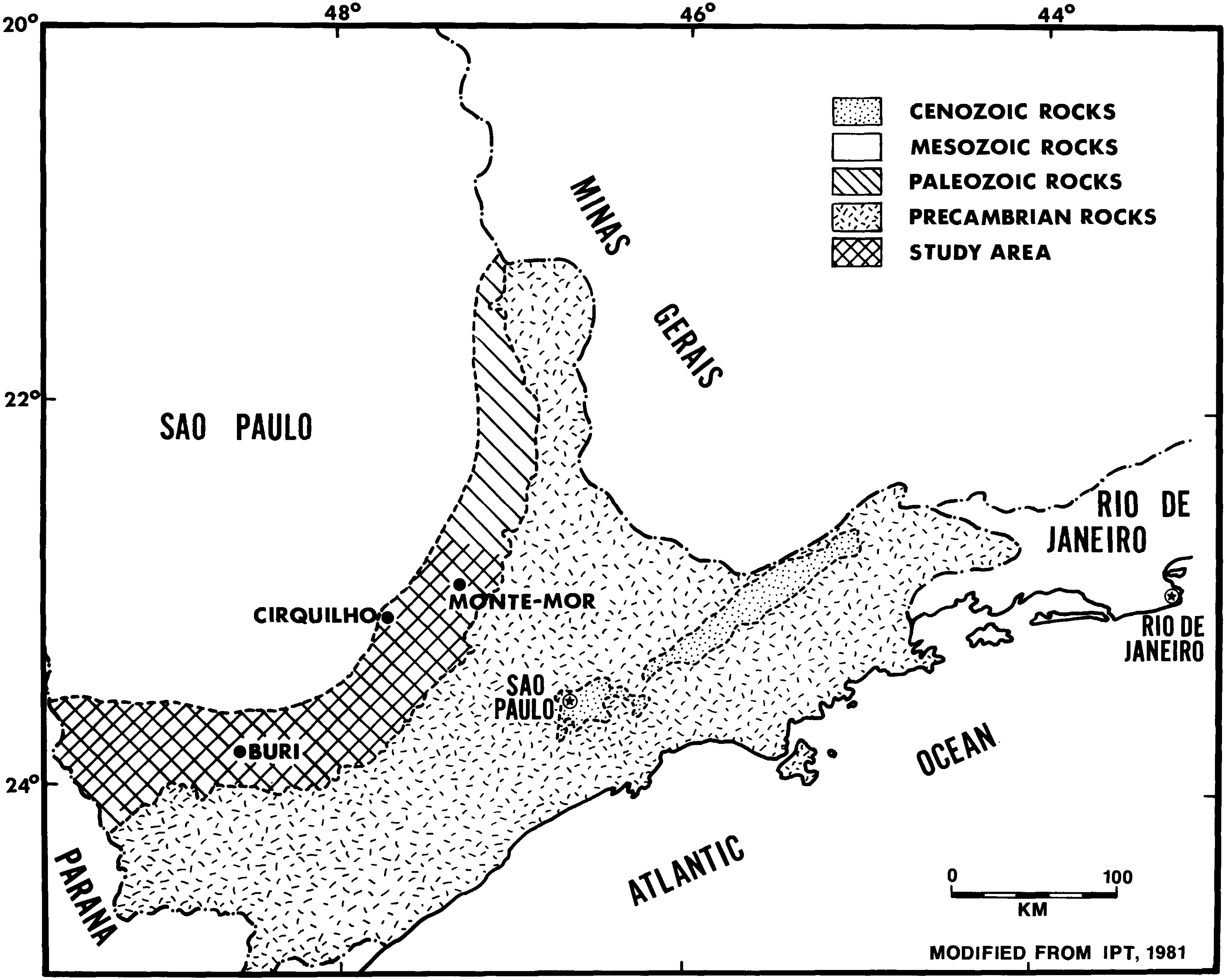


Figure 20.--A generalized geologic map of the State of Sao Paulo, Brazil, and locality map of the study area. Adapted from Instituto de Pesquisa Tecnologica (1981).

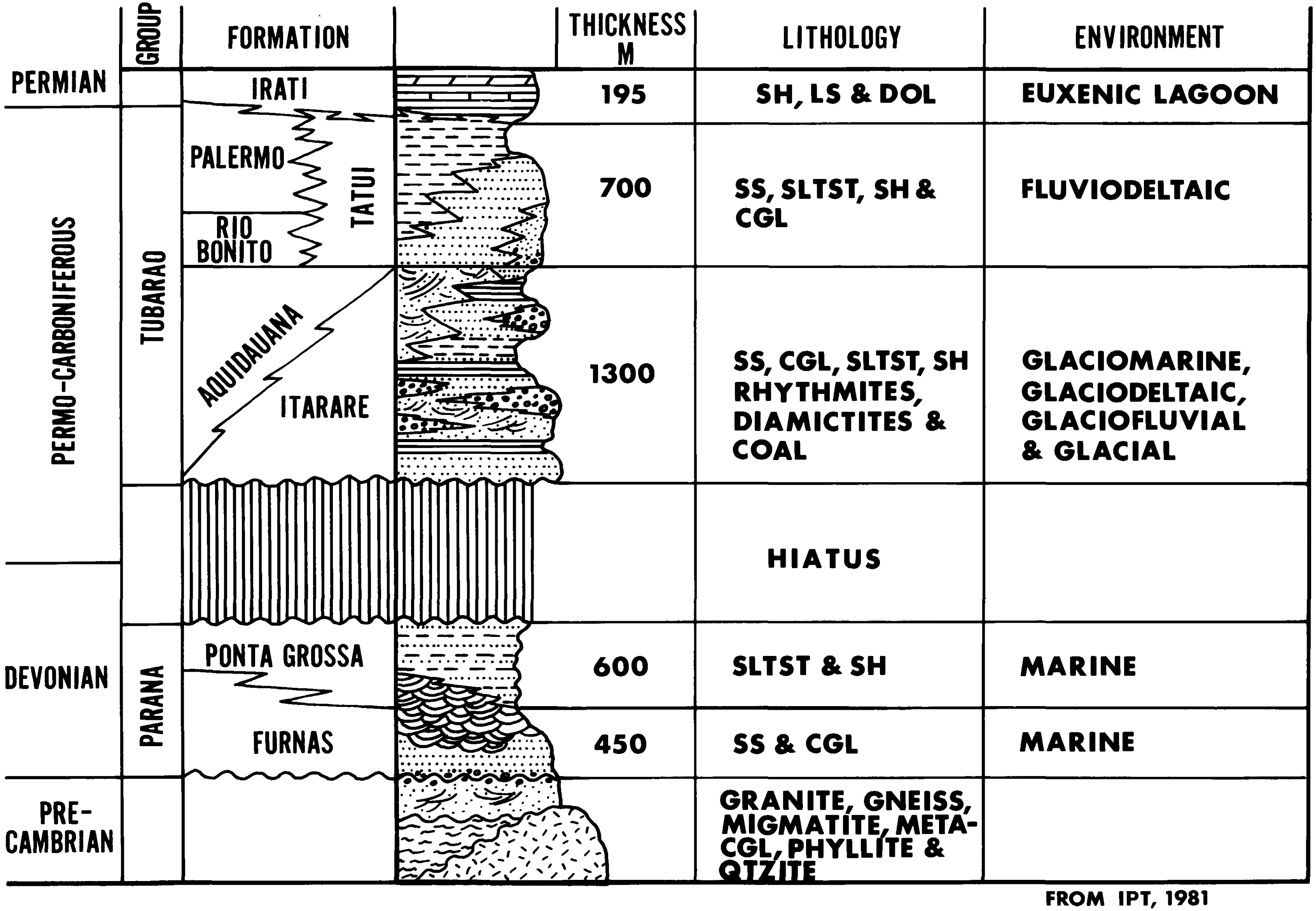


Figure 21.--A composite stratigraphic column of the Paleozoic rocks in northeast Parana Basin, State of Sao Paulo, Brazil. Adapted from Instituto de Pesquisa Tecnologica (1981).

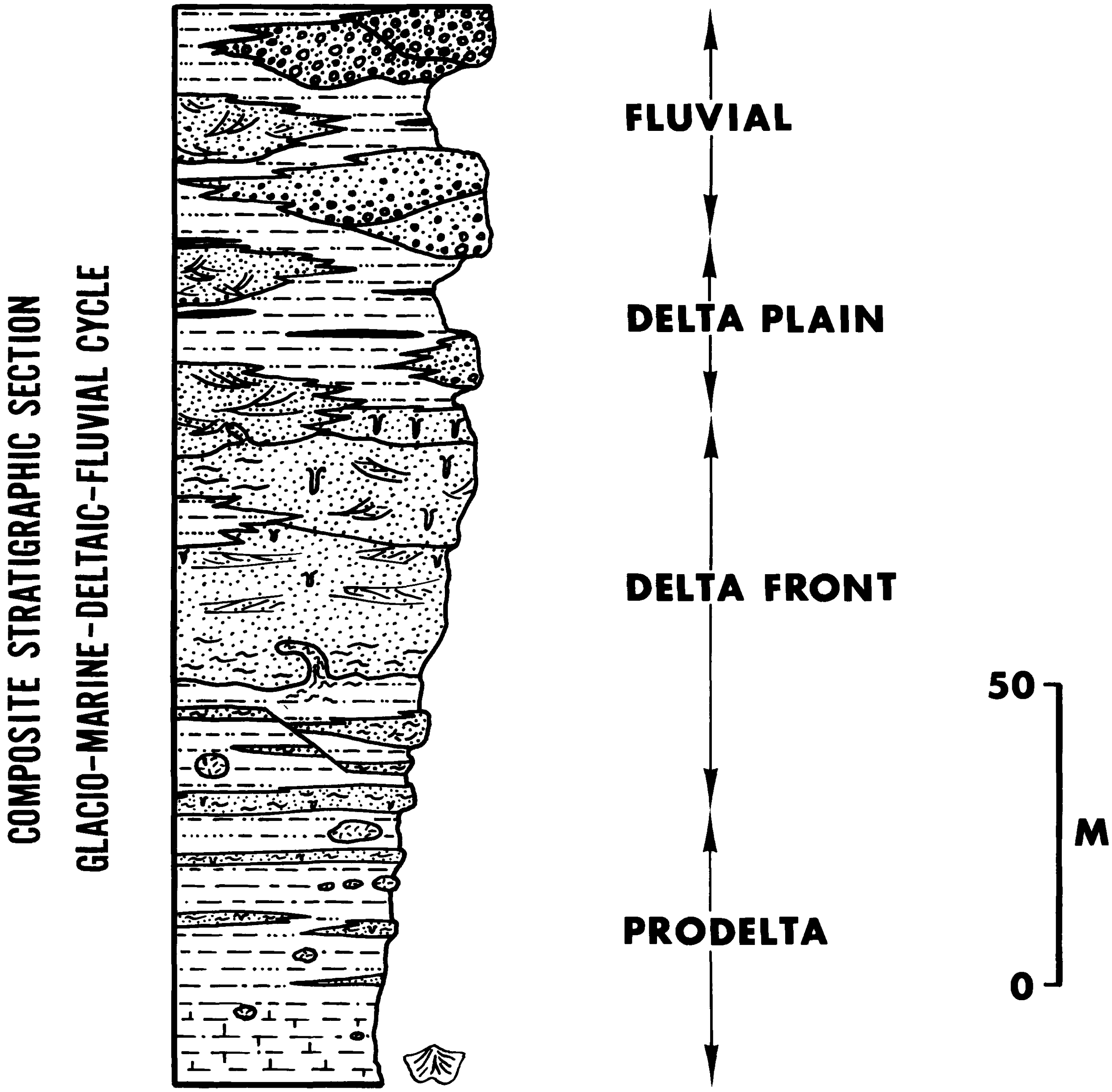


Figure 22.--A composite stratigraphic section of the proglacial deposits in the Itarare Formation.

rock fragments. Landward, the distributary channel sandstones merge laterally into "diamictites" or poorly sorted conglomeratic deposits. These "diamictites," which are deposited in multiple-stacked channels and are matrix supported, resemble debris flow deposits in braided channels and/or distal alluvial fans.

The coals in the glaciodeltaic and glaciofluvial deposits occur in the middle and upper part of the Itarare Formation as shown in figure 23. The coals in the Buri and Monte-Mor coal fields occur in the middle part of the Itarare Formation and the coals in the Cirquilho coal field are distributed in the upper part of the formation. On the basis of microflora zonations, the coals in the Buri and Monte-Mor coal fields were analyzed to coexist with brackish-marine intervals in adjacent areas in the lower part of Microflora II zone. A similar coexistence of coals and brackish-marine deposits occurs in the Cirquilho coal field in the upper part of the Microflora III zone. Thus, given these time-stratigraphic zonations, it is possible to plot local advances of brackish-marine waters against coeval coal swamps during Itarare time. These plots are given in a succession of paleogeographic maps in figure 24. The plots show that during the Lower Microflora II time coal swamps in the Buri and Monte-Mor coal fields, as well as the coal swamp in the Cirquilho coal field during Upper Microflora III time, were developed marginal to local advances of brackish-marine waters. However, as shown by the paleogeographic map of Middle Microflora II time, brackish-marine water advances were not necessarily accompanied by development of swamp. These regional advances of brackish-marine waters were probably controlled by eustatic rise of sea level due to deglaciation. The brackish-marine advances and their associated deposits were modified by northwest advances of continental glacier ice. Figure 25 illustrates a generalized model of deposition of the Itarare Formation showing supraglacial, subglacial, periglacial, and proglacial environments. The coals were chiefly accumulated in swamps in the glaciodeltaic environment and subordinately accumulated in the glaciofluvial swamps. The proglacial swamps formed in the Itarare Formation are a part of the overall global coal-forming swamps that existed during Permo-Carboniferous time prior to the breakup of the supercontinent Pangaea into the Gondwana and Laurasia subcontinents (fig. 26). In the State of Sao Paulo, Brazil, the Itarare coal swamps formed in proglacial environments marginal to a marine reentrant that occupied the Parana Basin in the southern part of South America.

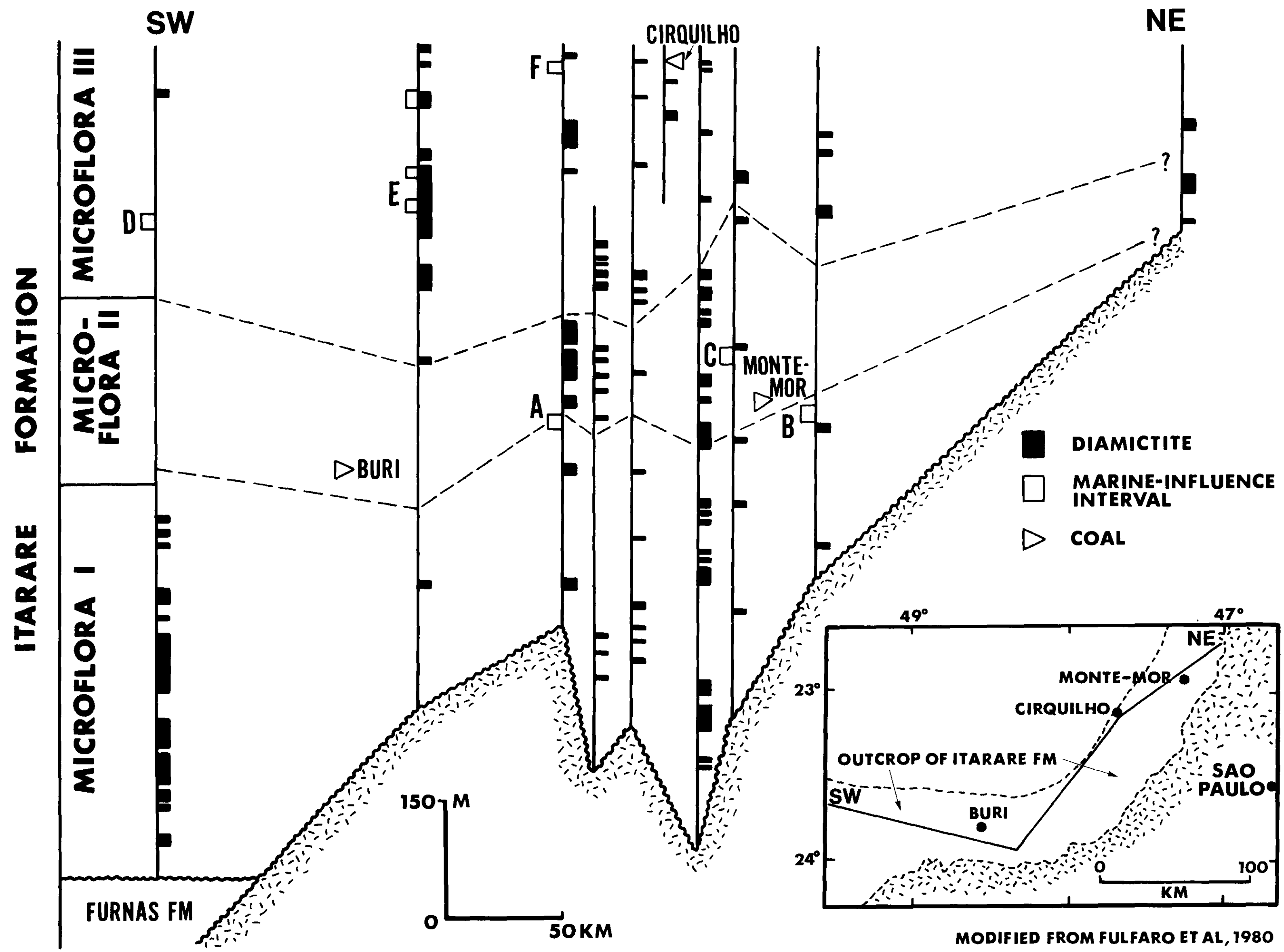


Figure 23.--A cross section and correlation of microflora zones of the Itarare Formation in the study area. Modified from Fulfaro and others (1980).

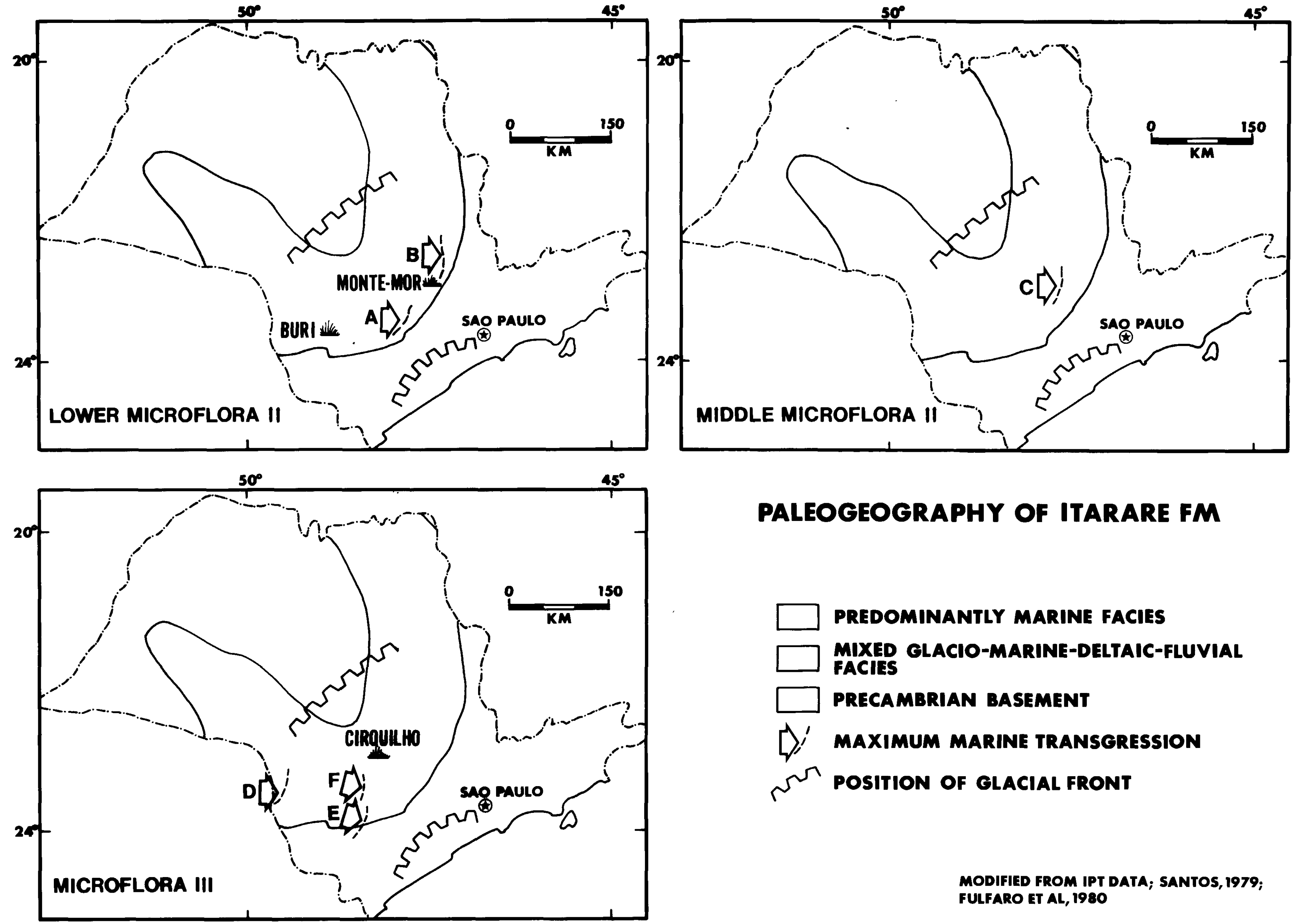
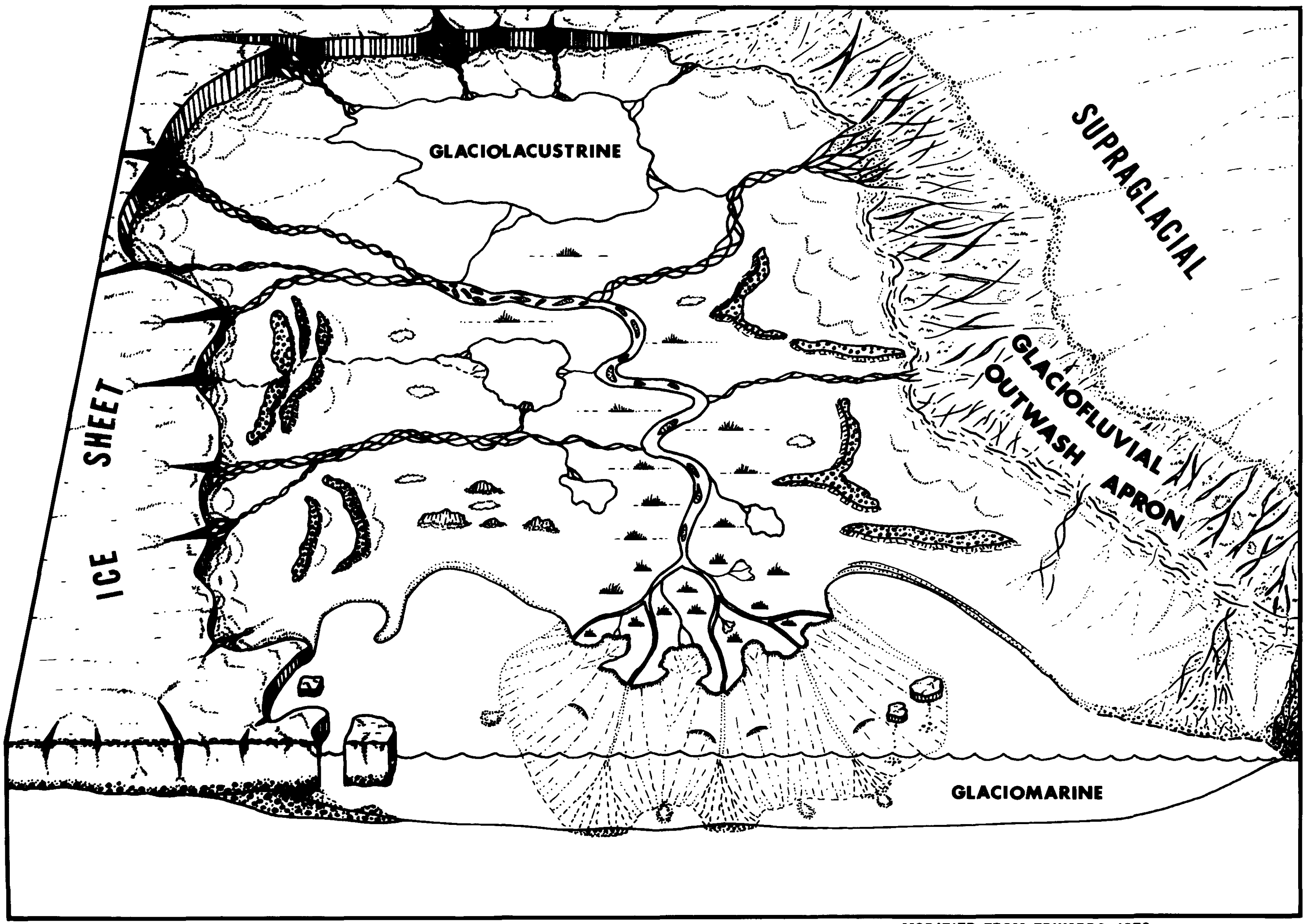


Figure 24.--Paleogeographic maps of the State of Sao Paulo, Brazil, during the Microflora II and III times. Modified from Santos (1979) and Fulfaro and others (1980).



MODIFIED FROM EDWARDS, 1978

Figure 25.--A generalized model of deposition of the Itarare Formation. Modified from Edwards (1978).

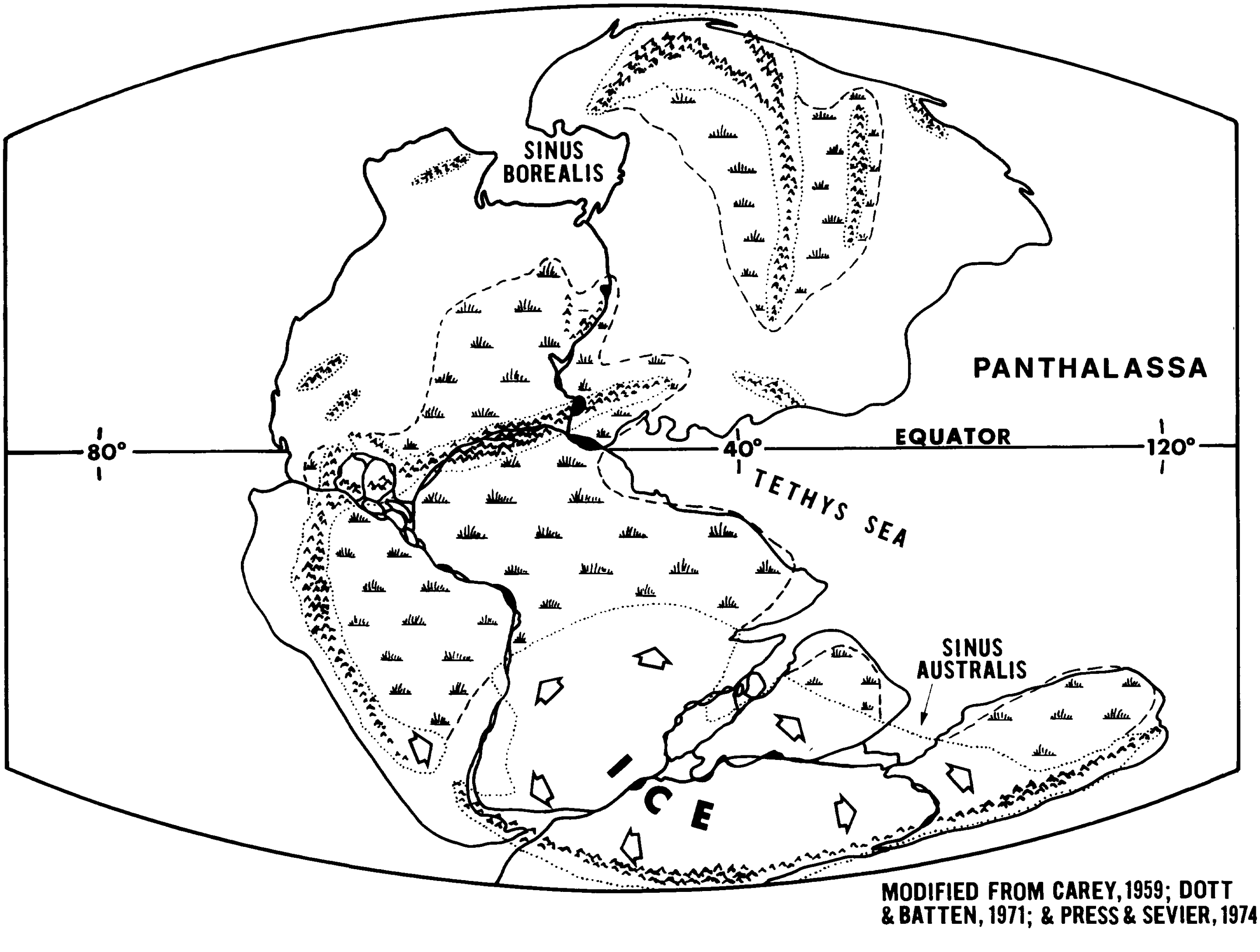


Figure 26.--A generalized paleogeographic map of the supercontinent Pangaea during Permo-Carboniferous time prior to continental drift. Modified from Carey (1959), Dott and Batten (1971), and Press and Sevier (1974).

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