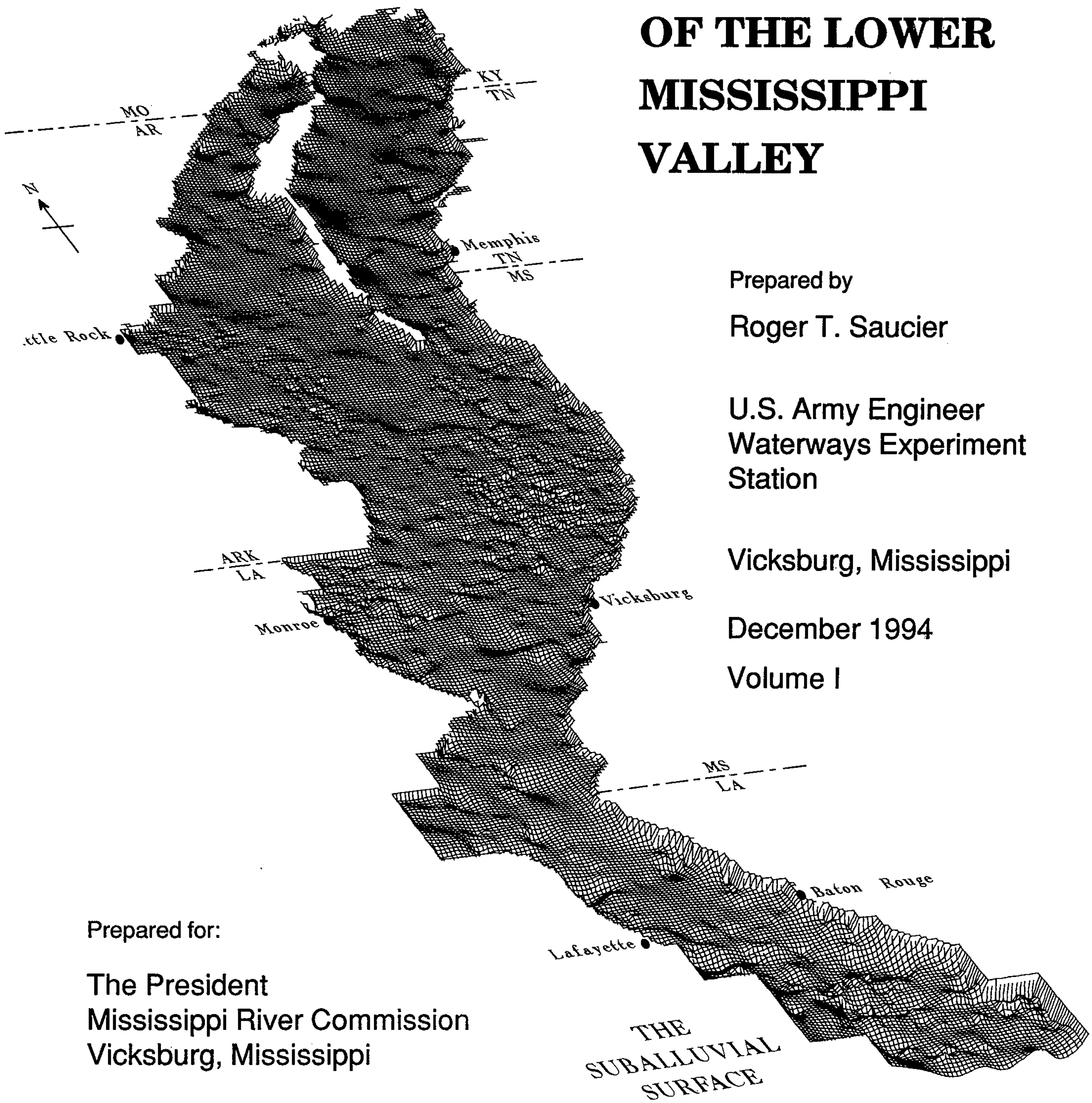
US Army Corps
of Engineers

GEOMORPHOLOGY AND QUATERNARY GEOLOGIC HISTORY OF THE LOWER MISSISSIPPI VALLEY



Prepared by

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Waterways Experiment
Station

Vicksburg, Mississippi

December 1994

Volume I

Prepared for:

The President
Mississippi River Commission
Vicksburg, Mississippi

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Preface

In May 1941, the U.S. Army Corps of Engineers, Mississippi River Commission (MRC), Vicksburg, Mississippi, hired Dr. Harold N. Fisk of Louisiana State University, Baton Rouge, Louisiana, as a geological consultant and tasked him to prepare a comprehensive summary of the geology of the Lower Mississippi Valley. The resultant product, *Geological Investigation of the Alluvial Valley of the Lower Mississippi River*, was published by the MRC in December 1944 and soon became *the* authoritative reference in the field and eventually a classic.

Although many of the concepts and interpretations set forth in 1944 have stood the test of time, others have not as new tools and techniques have emerged, new hypotheses have been offered and tested, and geological knowledge in general has advanced dramatically. To reflect the advances of the past 50 years, which include a voluminous but dispersed literature, huge amounts of new and sometimes unpublished data, and evolving concepts of valley history and chronology, the MRC agreed in October 1992 to sponsor the preparation of a new, comprehensive geological and geomorphological synthesis.

The synthesis was prepared between October 1992 and December 1994 by Dr. Roger T. Saucier while assigned part time as a Physical Scientist to the Earthquake Engineering and Geosciences Division (EEGD), Geotechnical Laboratory (GL), U.S. Army Engineer Waterways Experiment Station (WES), Vicksburg, Mississippi. General administration of the project was provided by Messrs. Frank Weaver, Lawrence Cave, and Tony Young of the Geotechnical and Materials Division, U.S. Army Engineer Division, Lower Mississippi Valley (LMVD), Vicksburg, Mississippi. Funds were provided by the U.S. Army Engineer Districts, Vicksburg, New Orleans, and Memphis, at the direction of the MRC.

Dr. Saucier was uniquely qualified to prepare the synthesis as a consequence of his more than 35 years of detailed mapping and investigation of multiple aspects of the engineering geology of the Lower Mississippi Valley while employed at WES and also serving as a private consultant. He has authored the only two limited attempts of the last 50 years to summarize aspects of the geology of the area, and his research focus, reflected in dozens of publications, has been on interdisciplinary considerations and applications ranging from archeology to paleoseismology. Dr. Saucier assembled and reviewed the literature for this synthesis, developed the outline and wrote the text, and prepared most of the figures and plates.

Technical guidance, consultation, critical review, and assistance in many aspects of the project were provided by Dr. Lawson Smith of the EEGD, WES, who also envisioned the need for the synthesis and served as a proponent of the project. Substantial assistance was provided by others in preparing two parts of the synthesis. Messrs. Tim Ethridge (LMVD), Mike Bishop (WES contractor), and William Murphy (EEGD) assembled and digitized the data and implemented the computer contouring of the suballuvial surface and prepared the graphical output. Mr. Ken Stroud, graduate student at the University of Southern Mississippi, conducted under contract to WES the assembly, evaluation, statistical analysis, and portrayal of the data on the geotechnical characteristics (Appendix A). Drafting and scientific illustration associated with the numerous figures and plates were accomplished by and under the direction of Messrs. William Park and Alan Middleton of the WES Visual Production Center, Information Technology Laboratory.

Peer review of the manuscript was provided by Dr. Whitney Autin of the Institute for Environmental Studies, Louisiana State University, and Dr. Lawson Smith and Mr. Joseph Dunbar, EEGD, WES.

The project was accomplished under the general supervision of Mr. Joe L. Gatz, Chief, Engineering Geology Branch, EEGD, Dr. A.G. Franklin, Chief, EEGD, and Dr. W.F. Marcuson III, Director, GL, WES.

At the time of publication of this report, Director of WES was Dr. Robert W. Whalin. Commander was COL Bruce K. Howard, EN.

Conversion Factors, Non-SI to SI Units of Measurement

Non-SI units of measurement used in this report can be converted to SI units as follows:

Multiply	By	To Obtain
acres	4046.873	square meters
acre-feet	1233.489	cubic meters
cubic yards	0.7645549	cubic meters
cubic miles	4.167571	cubic kilometers
degrees (angle)	0.01745329	radians
feet	0.3048	meters
gallons per minute	3.785412	cubic decimeters per minute
gallons (U.S. liquid)	3.785412	cubic decimeters
inches	25.4	millimeters
miles (U.S. statute)	1.609347	kilometers
pounds (mass)	0.4535924	kilograms
square feet	0.09290304	square meters
square miles	2.589998	square kilometers
tons (mass)	907.1847	kilograms

1 Introduction

Previous Investigations

With the possible sole exception of the insightful work of Lyell (1847), the earliest geological studies in the Lower Mississippi Valley area that reflect reasonably modern concepts generally date between the 1860s and the early 1900s. They include such notable works as those by Harris (1894) in Arkansas; Hilgard (1860) in Mississippi; Lerch, Clendenin, and Stubbs (1892), and Veatch (1906) in Louisiana; Safford (1869) in Tennessee; and Swallow (1877) in Missouri. Emphasis in these publications was on agriculture and mineral resources, but they also included descriptions of outcrops, springs, and special features and devoted considerable attention to groundwater. Without question, the focus was on the pre-Quaternary uplands. This reflects a trend that was to diminish but not disappear for many decades, i.e., geologists were interested in the uplands and "hard rock" and not alluvium. The only significant exceptions were studies by Marbut (1902) of the lowlands of southeastern Missouri and by Stephenson and Crider (1916) of a portion of northeastern Arkansas.

Fundamental aspects of the geology of the alluvial valley, such as the presence of glacial outwash (valley trains), multiple Mississippi River meander belts, and effects of sea level variations, were well known by the beginning of the 20th century. However, they were not comprehensively or systematically described by geologists during these early years. Soil scientists were actually the first to devote serious attention to the alluvial plain, and by the 1920s, there were published soil surveys for a number of counties in Arkansas, Louisiana, Mississippi, and Missouri. These constitute the first rather large-scale maps of Quaternary physiography and surficial deposits.

A major turning point in the documentation of Mississippi Valley Quaternary geology occurred in the early 1930s when the Louisiana Geological Survey initiated a series of truly comprehensive, large-scale, geological bulletins covering parishes in the southern part of that state (e.g., Howe and Moresi 1931, 1933). This series was later extended into central Louisiana (Chawner 1936) and also marks the first involvement of Fisk (1938) in Lower Mississippi Valley geology. The series focused on mineral resources (oil, gas, salt, and sulphur), and the first parish selected (Iberia) was because of its economically important salt domes. Nevertheless, the bulletins provided an unprecedented wealth of information on the Quaternary of the Mississippi Valley. During the decade of the 1930s, other states such as Mississippi and Tennessee initiated similar series but did not include counties in the alluvial valley area.

Prior to the 1930s, geological studies in the Lower Mississippi Valley area were largely descriptive, concentrating on outcrops, physiographic features, and stratigraphic correlations. In terms of alluvial and deltaic deposits, little attention had been devoted to relating them to geomorphic processes and explaining their origin and variability. Relatively few borings had been drilled; hence the nature of the subsurface beneath the alluvial plain was largely unknown. Consequently, geological knowledge was of limited utility to engineers who were concerned with siting dams, designing levees, employing river regulation, and similar activities.

Another major turning point occurred in the early 1940s when the Mississippi River Commission (MRC) retained Harold N. Fisk, a geologist at Louisiana State University (LSU), as a consultant. Between 1941 and 1948, Fisk and a team of graduate students prepared 30 special geological reports for the MRC on such topics as bank erosion, levee underseepage, faulting, and alluvial deposits and landforms, and geologic studies of proposed lock and dam sites and other structures in Louisiana, Arkansas, Mississippi, and Texas. Most of these reports were never published; however, during the eight-year period, Fisk did publish two monographs (Fisk 1944, 1947) of valley-wide scope that were to soon become widely known and used, and eventually to achieve the status of geological classics. These two works marked the beginning of engineering geology in the Lower Mississippi Valley.

Fisk's ability to relate the origin, distribution, and characteristics of deposits to fluvial, deltaic, and marine processes, and hence develop a predictive capability for engineers, was largely attributable to two significant developments. First, he was able to take advantage of the first complete coverage of the Lower Mississippi Valley by large-scale, vertical, aerial photography that was obtained by the U.S. Soil Conservation Service in the late 1930s. Second, he was able to obtain samples and logs from a large number of strategically placed, carefully drilled, and logged borings that penetrated the alluvial sequence. Those borings provided the first truly three-dimensional look at the deposits of the alluvial valley and deltaic plain.

During the decades of the 1950s and 1960s, there was a virtual explosion of the volume of geological knowledge, including that of direct relevance to engineering practice. Thousands of site-specific as well as topical studies were accomplished and published (see bibliography of Saucier 1966), covering virtually all fields of modern geology such as geomorphology, sedimentology, tectonics, seismology, paleontology, stratigraphy, and chronology. Major advances in knowledge resulted not only by the greatly increased interest and efforts by academia and government but also from the advent of new tools and techniques. Paramount among these are geophysics, radiocarbon dating, and remote sensing.

Upon the departure of Harold Fisk from LSU in 1948 (to Humble Oil & Refining Company), the focal point for specific engineering geologic investigations of the Lower Mississippi Valley, and the continuation of this tradition, shifted to the Geology Branch of the U.S. Army Engineer Waterways Experiment (WES) in Vicksburg, Mississippi. Between that date and the present, hundreds of geological studies of project sites and particular problems have been conducted. However, perhaps the most important contribution was the systematic large-scale (1:62,500) mapping of geomorphic surfaces and subsurface soils conditions and the inferred environments of deposition for essentially all of the Mississippi alluvial valley and its major tributaries within the jurisdiction of the U.S. Army Engineer Division, Lower Mississippi Valley. Eight folios of maps and cross sections, each covering a major basin area, have been published in limited editions (Fleetwood 1969; Kolb et al. 1968; May et al. 1984; Saucier 1964, 1967, 1969; Smith and Saucier 1971; Smith and Russ 1974) for use primarily by the U.S. Army Corps of Engineers.

Geological studies in the decade of the 1970s and continuing to date have been more precisely focused on particular problem areas and locations and much more sophisticated in approach. Two trends are evident. First, there has been a definite reach-out for additional tools and knowledge from related disciplines such as archeology, botany, chemistry, pedology, climatology, and oceanography. This has been necessary not only to resolve problems in chronology and paleoenvironmental reconstructions but also to better understand processes affecting the development of the area. Second, it has been possible to go beyond simple knowledge expansion to needed efforts in synthesis and reanalysis, reinterpretation, and revision.

Because of the intent of the present volume, attention is directed toward the status of efforts at summarization and synthesis rather than specific investigations. Since the first comprehensive look at the entire Lower Mississippi Valley by Fisk (1944), there have been only two attempts to give engineers a single source for an overview of the geology of the whole area with a discussion of the latest concepts. A brief summary of the Quaternary geology of the valley was prepared in the mid-1970s (Saucier 1974); however, it was intended primarily for cultural resources managers. It contained a new generalized geologic map (approximate scale 1:1,250,000), a discussion of problems with concepts, and the chronology of Fisk (1944), but it did not offer solutions. A more comprehensive overview of the Quaternary geology of the valley area, including a moderately detailed, full color, geologic map (scale 1:1,100,000), was published in 1991 (Autin et al. 1991). This work, the most definitive since 1944, provides the most up-to-date discussions of new concepts of valley origin and history, including a substantially revised chronology of events. It is balanced in scope and approach and intended for a general geology audience. A primary attribute is its widespread availability--a major benefit in view of the limited availability of the earlier work because of its rather obscure source.

Assessment of State of the Art

The latest synthesis of Lower Mississippi Valley geology (Autin et al. 1991) is modest in scope and has not been available very long. Therefore, the classic 1944 study by Fisk is still the most widely known, if not widely accepted, work of its type. Despite its age and the fact that many geologists recognize it has significant shortcomings, the 1944 work retains its appreciable status for several reasons, especially among engineers. For several decades it was the only reference of its type available, it is comprehensive in scope, and it was (for its vintage) magnificent in execution. It consists of two volumes: a volume primarily of text, and a pocket of 26 large, multicolored maps in six series. The maps truly are what has made the work so widely recognized and used. Although the text contains discussions that are pioneering contributions to the geology of the area, such as the effects of bed and bank materials on river behavior, they have been overshadowed by the materials with high graphics appeal.

In 1944, Fisk's contribution was a giant step forward in advancing the state of knowledge of the area, and it remained so for many years. However, as is almost always the case with such a comprehensive work, it soon became dated as new concepts emerged and new geological tools were developed. Normally when this happens, the newer concepts and data from later individual studies are periodically synthesized and new summaries prepared. However, in the case of Fisk's work, it influenced so many with its impression of detail, accuracy, and precision that it actually deterred later similar efforts (Saucier 1981).

Compounding the situation was the matter of normal problems in the interdisciplinary transfer of technical information. Practicing geologists recognized that the work was becoming outdated, but engineers were not (and could not have been expected to be) cognizant of the newer school of thought. Moreover, Fisk's work, especially his elaborate reconstructions of river channel patterns and chronology, became widely recognized by and of immense importance to archaeologists. They also were not privy to slowly evolving geological concepts and knowledge, most of which were not published and none of which were summarized in the literature in their field.

Possibly the most long-lasting contributions of Fisk in his tenure as consultant to the MRC were the conversion of geological information into a form directly useable in engineering practice, and the awareness he instilled in engineers of the value of a sound understanding of the affect of geologic processes in alluvial settings. Fisk's 1944 study epitomizes the former, and therein lies another problem. Fisk recognized that engineers could not use information that was presented in the context of multiple hypotheses, dissenting professional opinion, or that was heavily caveated with degrees of uncertainly. Therefore, he presented his results largely without qualification or alternate interpretations. Unfortunately, this sometimes involved omitting mention of features and topics for which he had no explanation. Eventually, significant problems in the use of his works developed among scientists in several disciplines.

Realization even by geologists of the extent and consequences of the limitations of Fisk's 1944 study has been slow in developing, especially as related to the chronology and history of river changes, terrace stratigraphy, and modes of valley formation. A few alternate interpretations were offered rather quickly (e.g., Doering 1956, Durham 1962, Leighton and Willman 1950), but hard evidence was much slower in materializing. Specific deficiencies in existing concepts became apparent, and a new chronological framework for terraces, valley trains (glacial outwash deposits), and meander belts eventually developed as a direct spinoff from the detailed mapping of environments of deposition at WES (Saucier 1968, 1978, 1987; Saucier and Fleetwood 1970). The first relatively complete (but not detailed) statement of new concepts and a suggested revised chronology of valley events appeared in Autin et al. (1991).

The most significant limitations of the 1944 work of Fisk are directly attributable simply to major advances in general geologic knowledge and theory that have developed within the past 50 years. Fisk assumed that glacioeustacy was the fundamental control on fluvial form and process, and he logically based his chronological concept on the best model available from the midwestern United States for glacial and interglacial cycles. However, that model has changed dramatically over the years. For example, Fisk based his chronology on the then-widely-accepted simple Pleistocene glacial sequence consisting of five glacial and four interglacial stages, whereas the more widely accepted present models are much more complex with possibly as many as eight cycles (Beard, Sangree, and Smith 1982; Richmond and Fullerton 1986) of varying duration and intensity. The new models are based heavily on oxygen isotope data from deep sea cores--a technology that did not exist in the 1940s.

If Fisk actually erred in judgment, it was with regard to his attempt to assign precise age estimates to all Mississippi Valley events, including estimates to within 100 years for all Mississippi River abandoned courses and cutoffs. These estimates have proven to be wrong in both relative and absolute terms, to have been based on wrong assumptions, and to be essentially beyond the state of the art even today. Even with radiocarbon dating and the extensive data from the archeological record, it may *never* be possible to do what Fisk attempted in 1944. As will be discussed later, there are still ambiguities in the fundamental relationships between whole abandoned meander belts that will require extensive field investigations to resolve before more detailed reconstructions are possible.

Even after extensive consideration, there are aspects of the mechanics of meander belt initiation, growth, and decay that are still unresolved. Meander belt life cycles apparently vary temporally and spatially, but the precise causal factors are not known. Hence, anything other than rough age approximations are meaningless and misleading.

Other aspects of Fisk's 1944 study that require revision include his overestimation of the effects of base level changes on the formation of the alluvial valley, his excessive emphasis on faulting and fracturing, his rejection of the eolian origin of loess, and his failure to recognize the effects of climate change on both landforms and geomorphic processes. Once again, Fisk only reflected the accepted hypotheses of his time, many of which have subsequently been rejected or significantly modified as geologic knowledge has increased.

Consideration of the large number of studies that have been accomplished since the works of Fisk will show that the state of the art of Lower Mississippi Valley geomorphology and geology is not well balanced, either topically or geographically. Despite the limitations previously mentioned, the area must rank as one of the best known large alluvial valleys in the world from the viewpoint of engineering geology. Subsurface conditions have been extensively and intensively explored primarily as a consequence of the history of major flood control, navigation, and water resources management projects by the U.S. Army Corps of Engineers. Consequently, much is known about geomorphic processes and the geotechnical properties of the deposits. Ironically, however, the alluvial valley is relatively poorly known from a purely geologic perspective, especially with regard to stratigraphy. Few studies of sedimentary mechanics and facies relationships have ever been undertaken.

Comparatively, the deltaic plain has been much more extensively explored than the alluvial valley *per se* and is much better known in a geological sense. That knowledge is a direct consequence of several decades of intense investigations by petroleum and geophysical companies aimed at both better understanding the Mississippi delta itself as well as the fundamental aspects of the occurrence and recovery of oil and gas in large deltaic systems. A very large percentage of the numerous stratigraphic, petrologic, structural, paleontologic, and chronologic studies that have been undertaken by universities and state agencies have been funded directly or indirectly by the petroleum industry. More recently, investigations of coastal and wetland processes relevant to deltaic ecology and land loss processes have contributed valuable new geologic data and concepts.

Intent and Scope of the Present Study

This volume is intended as the first comprehensive overview and synthesis of the geomorphology and geology of the Lower Mississippi Valley--both the alluvial valley and the deltaic plain--since 1944. However, it will vary from that earlier classic work in several respects.

A primary difference is an attempt in the present synthesis to present geological information and concepts to an audience of not only engineers and geologists but also scientists in a broad range of disciplines. It is primarily a document funded by and intended for the U.S. Army Corps of Engineers, but not just for the geotechnical, foundation design, potamology, and similar engineering functions of that organization. The synthesis also takes into consideration the informational needs of those concerned with such activities as basin planning, environmental assessments and mitigation, and cultural and natural resources management.

Engineering geology in its broadest context is the appropriate category for classifying this volume; however, it is not a "how-to" manual consisting of unequivocal answers. Neither is it primarily a theoretical treatise focusing on evaluating the strengths and weaknesses of alternative hypotheses and paradigms. The synthesis hopes to achieve a balance between the two extremes by presenting the generally accepted concepts and also by providing a discussion of the degree of confidence in the interpretations. It strives to separate pure speculation and hypothesizing (which unfortunately is all that exists in some cases) from that which is well founded and scientifically tested. It does not dwell on or attempt to resolve controversial issues.

Because of this approach, and in comparison to the emphatic and unequivocal nature of Fisk's 1944 study, some aspects of this volume may appear to be a major step backward rather than an advancement of knowledge. This is especially true with regard to the chronological reconstructions. However, as has been pointed out, the earlier study attempted to present age estimates that are now regarded as being unjustified in detail and precision. The present volume presents much less detail, but detail for which there is recognized evidence.

The organization and section content of this volume involves some difficult tradeoffs between logical order of discussion and data presentation and the interests and backgrounds of the user audience. For example, the decision to separate discussions of mode of origin, lithology, and geotechnical characteristics of depositional environments into separate parts--not necessarily logical to some--was done to accommodate the different interests and needs of geologists versus engineers. With a broadly multidisciplinary audience in mind, this writer intentionally adopted a writing style comparable to that contained in publications like *American Scientist* or *Scientific American.*

Geographically, this volume is restricted to the Lower Mississippi Valley below the approximate latitude of Cape Girardeau, Missouri, and to the lower reaches of its principal tributaries (Figure 1). Discussion and mapping is limited to the area downstream from Little Rock, Arkansas, on the Arkansas River, below Alexandria, Louisiana, on the Red River, and south of the Arkansas-Louisiana border on the Ouachita River. Considerable information on the Quaternary geology exists for the more upstream portions of those rivers, and interested readers are referred to the works by L.M. Smith (1986), Smith and Russ (1974), and Fleetwood (1969). Quaternary deposits and formations above the mouths of smaller tributaries are not mapped and discussed, but readers are referred to Saucier (1987) as an example of some of the information that is available. The entire Mississippi River deltaic plain and a portion of the chenier plain in southwestern Louisiana are included in the scope of this volume. Gould and McFarlan (1959) is the primary reference for

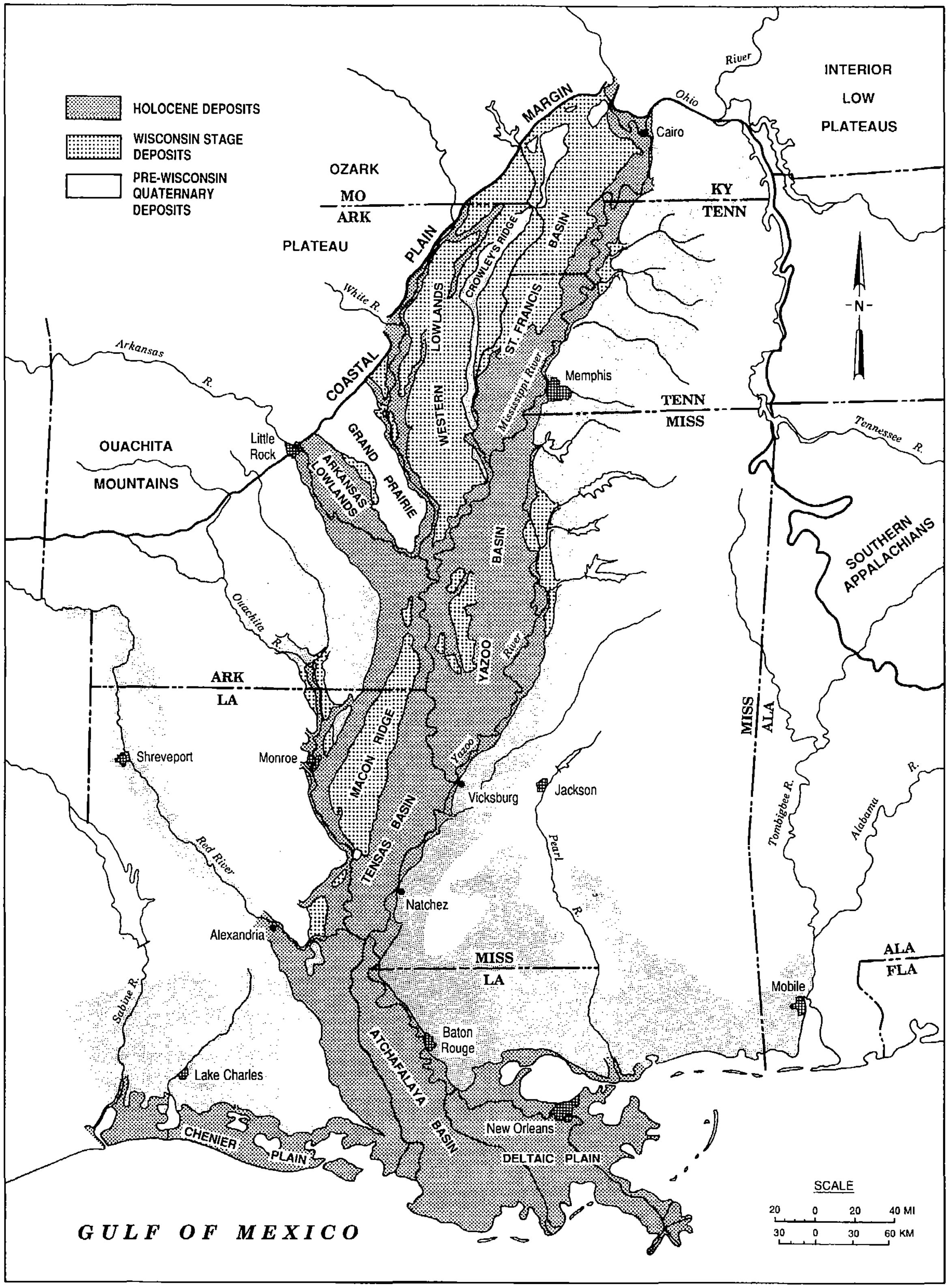

Figure 1. The Mississippi alluvial valley, deltaic plain, and chenier plain, and the total distribution of Quaternary deposits in the Lower Mississippi Valley portion of the Gulf Coastal Plain

information on the remainder (western portion) of the chenier plain. The discussion and mapping of the deltaic plain are limited to the subaerial portion of that feature: Coleman (1988) and Coleman and Roberts (1988) contain a summary of the geology of the offshore areas and subaqueous delta platform.

Quaternary deposits (fluvial terraces and related formations) of the uplands adjacent to the alluvial valley are discussed in this volume; however, detailed mapping is generally limited to a narrow band flanking the Holocene alluvial plain.

To avoid excessive length, readers are frequently referred to original sources for specific detail and documentation. For example, when radiocarbon dates are discussed, complete citations with error limits, laboratory numbers, precise locations, and similar data are intentionally not presented, even in tabular form. It is believed that readers must be cognizant of all aspects of the dates, especially their specific contexts, to allow use in future research or interpretations.

Methods and Limitations

There are known extensive data gaps in the knowledge of the geomorphology and geology of the Lower Mississippi Valley, but it was beyond the scope of this effort to undertake any field investigations or to develop any new information. Some limited new interpretations are offered, but these relate to resolving inconsistencies in existing data sets and the problems normally encountered when data from multiple sources and dates are synthesized into a single, and sometimes new format.

It is acknowledged that because this writer initiated and conducted many of the investigations on which this report is based and developed, it is heavily flavored by his experiences and professional opinion. Hopefully, however, it is not adversely biased or lacks objectivity. Where the writer's opinion does not reflect exposure to peer review or general acceptance, and where data from his unpublished files are used, it is so acknowledged.

Subsurface Investigations

Borings. For well over 100 years, it has been recognized that the Mississippi alluvial valley and deltaic plain are underlain by thick sequences of alluvial deposits. Lacking outcrops or bank exposures more than a few tens of feet high, even before the turn of the century, scientists logically took keen interest in samples from borings for clues to deltaic plain formation, and engineers searched for answers as to the thickness and nature of the deposits that were providing such poor foundation conditions (Hilgard and Hopkins 1874, 1884; Wilson 1881).

Although the MRC took a pioneering role in the engineering applications of subsurface investigations, until the 1930s most data were generated as a result of water wells drilled for irrigation in the alluvial valley area, and borings and wells drilled as part of petroleum exploration and development in the deltaic plain area. More than five decades ago, Russell (1940) observed that the Mississippi alluvial valley must be the best explored thick alluvial sequence in the world. Since that time, the amount of subsurface data available has increased by orders of magnitude.

By far the largest collection of subsurface data on the area exists in the files of WES. As part of the more than 30-year effort to systematically map the depositional environments of the Mississippi Valley, it has been standard practice to collect as much subsurface data as are practicably available. Routinely surveyed sources include U.S. Army Engineer District offices, state highway departments, the U.S. Geological Survey groundwater division offices, the U.S. Soil Conservation Service, private well drilling firms, petroleum companies, consulting foundation engineering firms, and published reports. As a consequence of the WES mapping efforts for the Corps, the WES files contain data on an estimated 250,000 borings.

In amassing these data, seldom has it been possible to examine actual cores or samples. This generally has been possible only in terms of special projects, such as Corps locks and dams, where detailed geologic investigations have been involved. Rather, the typical data used in the regional mapping effort have consisted of descriptive driller's logs or laboratory logs and some geotechnical test results.

Interpretations of subsurface environments of deposition and their stratigraphic relations generally have relied on traditional means of correlation using lithologic data and extrapolation from surface mapping. Geotechnical properties of the deposits are usually evaluated in a vital second step in the interpretive process. Contour maps of key horizons, such as the suballuvial surface, have been prepared where appropriate; however, the principal means of portraying subsurface interpretations has been by way of geologic cross sections. In association with the quadrangle mapping of depositional environments conducted by WES, approximately 10,000 miles of cross sections have been prepared. From those cross sections, more generalized transvalley sections have been prepared (modified from Autin et al. 1991) and are shown in Plate 16.[1]

The quality of the subsurface data is directly related to the purpose of the drilling. For example, water wells seldom are logged by geologists or soils engineers or with the needs or interests of those persons in mind; hence, the accuracy and precision of the data are usually low. On the other hand, borings made by engineering firms or federal and state agencies are of much higher quality and usefulness. Even the latter sometimes do not provide all of the

[1] All plates are in Volume 2.

detail that a geologist would desire, since soils characteristics such as color, minor sedimentary structures, and inclusions are not emphasized.

Regionally, the amount of subsurface data on which the quadrangle mapping of depositional environments has been based (and hence this synthesis reflects) varies considerable in both quality and quantity. Relatively, more data are available for the deltaic plain than the alluvial valley, much more data are available along the present Mississippi River channel (because of levee and revetment projects), and the data are concentrated in areas of higher population density. By way of contrast, there are 1:62,500-scale quadrangles in the Western Lowlands area that have been mapped using no more than a few dozen water well logs, whereas there are quadrangles in the New Orleans, Louisiana, area that have been mapped using over 10,000 borings made for engineering projects.

In the alluvial valley area, less than 20 percent of all borings, excluding water wells, are more than 50 ft deep and hence do not penetrate through the entire alluvial sequence. In the deltaic plain area, the average boring depth is considerably greater, and perhaps as many as 50 percent penetrate the Holocene alluvial sequence and reach the underlying Pleistocene-age deposits. The effects of these significant depth and regional variations in subsurface data are discussed in later sections that consider the nature of the suballuvial surface and the nature of the geotechnical characteristics of various depositional environments. It is important to note in overview that perhaps 80 percent of the total knowledge of subsurface conditions is restricted to about 30 percent of the total area.

Geophysical surveys. High-resolution remote sensing of subsurface conditions and characteristics has been a recent and very limited advent in the Lower Mississippi Valley. Refraction seismic surveys have been extensively conducted for decades in the deltaic plain area; however, these provide little or no useful data on deposits within the upper several hundred feet. Water-borne reflection seismic profiling (acoustic subbottom profiling) is a highly useful technology in many areas, such as the shallow continental shelf, but is severely limited by the nature of the deposits in inland portions of the Mississippi Valley. Systems that operate at frequencies necessary for high resolution are adversely affected by organic-rich deposits and cannot penetrate thick masses of coarse-grained deposits such as characterize the lower portion of the alluvial valley sedimentary sequence. Conversely, systems that operate at frequencies (and power levels) capable of penetrating the thick coarse-grained deposits cannot adequately resolve stratigraphic details.

The few acoustic subbottom profiles that have been available for geologic interpretations have been of limited value in discerning faults and mapping the configuration of the suballuvial surface in the deltaic plain (e.g., see Kolb, Smith, and Silva 1975). There has been only one noteworthy application of this technology in the alluvial valley area for engineering applications (Murphy 1983). More recently, both land-based and water-borne reflection seismic surveys have been extensively employed in a portion of southeast Missouri in an attempt to better understand the structure of the New Madrid seismic zone (Crone et al. 1986, VanArsdale et al. 1990). These have been moderately successful thanks in part to recent advances in data recovery resulting from systems using digital data processing and enhancement.

Aeromagnetic and gravity surveys have been made over the entire Mississippi Valley and adjacent areas, but these have not provided data of sufficient resolution to be of use in engineering applications. Electrical resistivity, ground penetrating radar, and similar techniques have been employed for investigations of small project sites at several locations in the area; however, these also have not contributed information of significance to this regional synthesis.

Map and aerial photo Interpretation

Basic quadrangle-scale mapping by WES of the depositional environments in the Lower Mississippi Valley has been accomplished primarily by aerial photo interpretation. Because of the general lack of outcrops, except along ridges and upland margins, and the large size of most alluvial landforms, aerial photos constitute an indispensable tool for initial landform identification and delineation. Subsequently, the photo interpretation is typically refined using subsurface data, topographic information, soils maps, and limited field reconnaissance. When the earlier quadrangles were prepared, topographic data were generally available at 5- or 10-ft contour intervals but at a scale of 1:62,500. At present, data at similar contour intervals are available for the entire Mississippi Valley area at a scale of 1:24,000. Many quadrangles at both scales are in their second or third revised editions.

Since the mapping of the first quadrangles in the Yazoo Basin about 35 years ago, it has been found to be essential to use multiple aerial photo coverages flown at different dates. Multiple photo coverage provides opportunities to observe features under differing vegetation and soil moisture conditions and often to observe subtle geomorphic differences on one coverage that may be invisible on another.

The most typically used coverage has been individual frames and index mosaics of vertical, black and white photography at a scale of 1:20,000 obtained from the Agricultural Stabilization and Conservation Service of the U.S. Soil Conservation Service. Coverage from this source is available at approximately 10-year intervals; however, by far the most valuable coverage was the initial photography flown in the early 1930s over much of the deltaic plain and in the late 1930s over most of the alluvial valley. These photos are of exceptional quality and allow interpretation of conditions as they existed before the adverse effects of extensive agriculture, land levelling, and other cultural modifications. Photo coverage by the U.S. Army Engineer District offices has been the next most valuable source and is available primarily along the major river systems. High altitude coverage is available from the

U.S. Geological Survey at various dates but is of limited use because of its small scale (1:50,000 or smaller).

Remote sensing coverage such as manned spacecraft photography and especially color infrared imagery has been available during the last two decades and has provided a supplemental tool of moderate value. However, its greatest utility has been in mapping land use patterns and changes, and processes such as shoreline erosion and channel changes rather than basic geomorphic interpretations. Similarly, satellite coverage such as Landsat imagery and side-looking airborne radar (SLAR) imagery has been of limited value and used primarily for considerations of lineaments and faulting (e.g., see Schweig and Marple 1991).

Geomorphic mapping and studies in the dynamic landscapes of the alluvial valley and deltaic plain have also benefitted from the use of the moderately accurate historic map series of the last 150 years or so. The first maps of significant value are the township plats of the U.S. Land Office Survey which, in the Mississippi Valley area, generally date to the 1830-1850 period. Subsequently, the next rather large-scale map series were the surveys of the Mississippi River by the MRC in the period 1871-1880, and these were followed at close intervals (e.g., 1881-1892, 1902-1904, 1913-1916) by surveys at scales ranging from 1:20,000 to 1:63,360. Such historic maps have been quite helpful in determining natural drainage patterns and the extent of wetlands and hydrographic features before modification by man.

Georcheological investigations

Applications. Archeological evidence is abundant that the initial human occupation of the Lower Mississippi Valley area occurred at least 12,000 years ago and occupation has been continuous ever since. Throughout this time, humans have had to modify their settlement patterns, adjust their life styles and exploitation of natural resources, and even sometimes migrate in response to cultural as well as environmental conditions. It is no wonder, therefore, that in the highly dynamic landscapes of the alluvial valley and deltaic plain, archaeologists have depended heavily on geomorphologists for an understanding of how the physical environment has changed in time and in space and influenced prehistoric human behavior. On the other hand, because of a need for more and better tools to understand landform and landscape evolution and changing natural processes, geomorphologists and geologists have relied heavily on archaeologists. Hence, the development of an understanding, interdependence, and degree of interdisciplinary cooperation has yielded significant results. Over time, scientists have come to view archeological sites as a component of the natural landscape in the same context as natural levees, beaches, or abandoned stream channels--each has a role to play and a story to tell in reconstructing geologic history. There are numerous examples in the literature of how the development of a new concept in archeology has quickly provided answers to puzzling questions in geomorphology, and vice versa.

Archeological sites have immense importance as chronostratigraphic markers: in many cases, they provide the most reliable evidence as to the age of a landform or an event. Artifacts diagnostic of a particular cultural period, although they may be undated directly in a particular context, can often provide age estimates to within several hundred years based on extrapolations from dated sites. Fortunately, artifacts exhibit distinctive styles of decoration, manufacturing techniques, and form that have definable temporal and spatial ranges. Even more precise dates are possible where wood, charcoal, bone, or other organic matter can be assayed by the radiocarbon method. Typically, archeological sites are most abundant on ridges or topographically high areas adjacent to streams or lowlands, but they provide only *minimum* age estimates for the landform on which they are located. In a highly characteristic scenario, sites on natural levee ridges or terrace margins were first occupied after these landforms were already well developed and no longer subject to flooding and significant sediment accretion. In most cases, however, it is not possible to determine just how soon after landform development the first human occupation took place.

The types and distribution of archeological sites in a region can be as diagnostic as their actual composition in understanding processes and landscape evolution. The presence or absence of sites (or site components) of a particular age may be indicative of drainage system changes or stream channel shifts. Sites would not be expected to be present in an area during times of heavy sedimentation such as when a new meander belt was being formed. Conversely, a heavy concentration of sites in an area may suggest that a meander belt was recently abandoned with some streamflow but no major flooding present. In the deltaic plain area, the presence of sites (usually shell middens) in the midst of broad expanses of interdistributary marshes usually indicates the existence of buried natural levees or beach ridges associated with an older, subsided delta complex. Determining the elevation of the base of a partially buried site of a given age can even be a means of estimating minimum subsidence rates.

The floral and faunal content of archeological sites can be an important clue to regional ecological conditions and environmental change. With the knowledge that prehistoric populations did not travel long distances for their daily subsistence and there were no changes in dietary preference, the abundance of remains of aquatic species in a site may indicate the extent of nearby wetlands, such as an oxbow lake or swamp. In the deltaic plain area, the species composition of mollusks in shell middens is an indicator of the salinity regime of nearby waters. In turn, salinity can be an indicator of the stage of a subdelta in its characteristic growth/decay life cycle. It is not uncommon for the vertical sequence of shells in a midden to reflect a change from saline to freshwater conditions (or vice versa) within a matter of a few hundred years.

In broader context, archeological sites can sometimes indirectly contain evidence of climatic change. Sites often provide conditions favorable for the preservation of pollen, phytoliths, and other ecofacts that allow the reconstruction of former vegetative assemblages. Changes in vegetation patterns may be the result of the influence of variations in precipitation and temperature on local hydrologic conditions.

Within the last two decades, geoarcheological studies have been shown to have direct application to geological problems in addition to those concerned with geomorphic processes and paleogeographic reconstructions. Saucier (1977a) initiated a line of investigation that has proven valuable in identifying and dating paleoliquefaction events indicative of prehistoric earthquakes of moderate intensity. Subsequent studies of this type have helped to better define the seismic risk in the New Madrid zone.

Previous investigations. The history of archeological investigations in the Lower Mississippi Valley is similar in many respects to that of the geology of the area. The first intensive period of archeological activity occurred in the 1870s, but the first "scientific" archeology did not begin until the early 1930s (Jeter et al. 1989). Logically, these early investigations focused on locating, collecting, and making limited test excavations in the large earth mounds and mound complexes of the alluvial valley of which there were thousands.

Investigations of a truly geoarcheological nature also date to the 1930s. One of the first to reflect the use of site locations in paleogeographic reconstructions was by Kniffen (1936) who interpreted stages in the formation of the St. Bernard subdelta (delta complex) of the Mississippi River based on the relative ages of shell middens. He also found evidence in the middens of changing environmental conditions due to subdelta growth and decay.

Kniffen's pioneering work and influence (and decades of subsequent teaching at LSU) had lasting impacts in several ways. They were instrumental in prompting an assessment of the relationship between changing settlement patterns and the development of the entire deltaic plain (McIntire 1958) and a number of intensive efforts in paleoenvironmental reconstructions of particular areas such as the Pontchartrain Basin (Saucier 1963). Kniffen's influence was also a factor in perhaps the most significant archeological development to take place in the region--the formation in the late 1940s of the Lower Mississippi Survey (LMS), a joint effort between the LSU School of Geology, the University of Michigan, and the Peabody Museum of Harvard University. In addition to hundreds of studies of individual sites, the LMS produced two comprehensive overviews of the archeology of the upper and central parts of the alluvial valley (Phillips, Ford, and Griffin 1951; Phillips 1970). Both of these works relied heavily on correlating sites with stream channels, as mapped by Fisk, and produced site inventories that have been valuable references ever since.

In contrast to the situation regarding periodic syntheses of geologic knowledge of the region, there have been several major archeological overviews produced in the last several decades. Some of these have focused on particular river basins such as the Red River area by Davis (1970), particular cultural periods such as the Povery Point culture by Webb (1977), individual states or regions such as the central Mississippi Valley by Morse and Morse (1983), the state of Louisiana by Neuman (1984), and the entire area by B. D. Smith effective for a different Quaternary time-span, has ranges of precision error based on inherent limitations, and can be adversely affected by contamination from various sources. Perhaps the most severe limitation of all techniques is the limited occurrence of dateable materials that are in context, i.e., that are truly indicative of the deposits or events for which an age estimate is desired.

The oldest and by far the most widely used method in the Lower Mississippi Valley area is radiocarbon (or C_{14}) dating. This isotopic technique can provide statistical age estimates on organic materials to a practical maximum of about 35,000 years before present (BP) and, thanks to a recent breakthrough, to beyond 50,000 years BP through the use of accelerator mass spectrometry.

The availability of radiocarbon dates for geomorphological studies varies greatly in type and by area. In the alluvial valley area, there are approximately 400 known dates for which WES has relevant information in a database. Of that number, only about 30 percent are from geological contexts, mostly peat or organic clays from abandoned channels or backswamp areas. Due to apparent contamination by either younger or older organic matter (e.g., Thorne and Curry 1983) or for unknown reasons, about one fourth of the 30 percent has been considered to be anomalous. Rejection of radiocarbon dates as being anomalous is usually a judgmental matter where the dates are in conflict with stratigraphic or other evidence deemed to be more definitive, or where there is an illogical chronostratigraphic sequence (i.e., an older date overlying a younger date in a normal depositional sequence). The remaining 70 percent of the dates are from charcoal or bone from archeological sites. Those dates are generally more indicative and reliable than those from geological contexts but often do not directly date the landforms on which the sites are situated (minimum ages only).

Obtaining stratigraphically meaningful geological dates on alluvial valley sedimentary sequences has also been difficult because of a general scarcity of preserved organic matter and has been successful in only a few intensively studied pollen cores (e.g., Royall, Delcourt, and Delcourt 1991). The dating of freshwater shells and wood fragments from cores, often the only possibility, is often unreliable. In contrast, the deltaic plain is characterized by extensive peat and highly organic clay deposits (Fisk 1960). As a consequence, there are over 1,000 known published radiocarbon dates for that area, the vast majority of which are accepted as valid dates from geological contexts. This extensive radiocarbon dataset has allowed a detailed reconstruction of delta complex formation and quantification of rates of subsidence and sea level rise. However, in contrast to the alluvial valley, there are relatively few dates from archeological contexts since charcoal is not well preserved in many coastal shell middens. Except for some instances of inferred petroleum hydrocarbons in Holocene sediments, contamination is not a significant problem in dating deltaic plain deposits.

Thermoluminescence (TL) is a radiogenic dating method applicable to both geologic and archeological samples less than 500,000 years old. It has been used on samples of prehistoric pottery from archeological sites in the alluvial valley area and more extensively to date loess deposits in the uplands bordering the valley. TL dating can be applied to minerals that were exposed to sunlight (light bleaching) prior to deposition, and hence it is more effective on eolian than on fluvial deposits.

Amino acid racemization is a recently developed, biochemical, relative dating method with potential wide application to both freshwater and marine archeological and geological samples. The method is based on determining ratios of protein amino acids in fossils such as shell and bone. To date, amino acid chronologies (aminostratigraphy) has been attempted primarily on gastropod shells from loess deposits.

Dendrochronology is the only other numerical dating method that has been applied with success in the Lower Mississippi Valley area. The detection and dating (to within a few years) of forest disturbance due to prehistoric earthquake activity, the dating of archeological sites and features, and paleoclimatic reconstruction have been attempted using data from bald cypress trees in Arkansas (Stahle and Wolfman 1985).

When all dating methods are considered, the total of about 450 numerical dates from the alluvial valley area has been wholly inadequate to develop a chronology comparable to that of the deltaic plain. This lack of basic chronologic information becomes much more critical with increasing age of landscapes. The available dates, especially the ones from archeological sites, are predominantly limited to the Holocene: for example, very little chronologic control of any type exists for most Pleistocene landforms and deposits.

Information Sources

This synthesis is based heavily on information contained in published sources, however obscure they may be. It is not, however, meant to be an exhaustive summary of the available literature since many thousands of items would be involved. Cited references are primarily for the benefit of readers that wish to identify other related works and for documentation of a particular topic. Where statements or discussions reflect unpublished sources, they are appropriately identified.

Illustrations

No new map or photo interpretation was performed as part of the compilation of this synthesis other than to resolve minor matching conflicts between maps prepared at different dates by different authors. Rather, the total of 215 published quadrangle-scale maps by WES (in eight folios) showing the distribution of depositional environments were compiled into eleven maps at the reduced scale of 1:250,000 (Plate 3). These were augmented with interpretations taken from 16 quadrangles being mapped at the time of preparation of this synthesis (1993) and from preliminary data for 21 quadrangles not yet mapped. Most of the latter occur in the deltaic plain in areas where previous geologic mapping has been adequate for the purposes of this synthesis. The eleven 1:250,000-scale compilations constitute Plates 4 through 14 of this report and form the basis for many of the discussions and interpretations. Base maps used for the compilations are standard U.S. Geological Survey 1- x 2-deg sheets obtained without the brown (contours), green (woodlands), red (road classification), and yellow (urban areas) overlays.

The landform/depositional environment classification used in Plates 4-14 is fundamentally a slightly simplified version of that used in the quadrangle-scale mapping by WES (Plate 3), and the colors are comparable except for the Pleistocene terraces. However, the alphanumeric designations are a modified version of what was used in Autin et al. (1991). This scheme allows better differentiation of features according to age and the river responsible for their formation (e.g., Arkansas River versus Mississippi River). Because of the smaller scale of the eleven plates, it was necessary to omit certain features such as swales in point bar accretion areas and the distribution of natural levee deposits in the alluvial valley area. Otherwise, the level of detail and accuracy are comparable.

Plate 15 portrays the hypsometry of the suballuvial surface, and Plate 16 contains typical cross sections showing the relationship between the Late Pleistocene and Holocene alluvial and deltaic deposits and the underlying formations. Plates 17-27 portray contours on the suballuvial surface. The base maps for this series are exactly the same as those used for the mapping of the environments of deposition (Plates 4-14). A discussion of the data and techniques used in the contouring of the suballuvial surface is presented in Chapter 5.

A detailed chronologic reconstruction of stages and events in the history of the Lower Mississippi Valley, including Mississippi River channel changes, is not possible as Fisk attempted to do 50 years ago. Therefore, a different approach was selected for this synthesis. Plate 28 (A-M) portrays the probable landscapes as they existed at thirteen key times in valley history, based on available data from all sources. No attempt has been made to estimate the precise location of the Mississippi River or its main tributaries; rather, the paleogeographic maps reflect only the interpreted locations of whole meander belts. Reliable determination of the ages of particular abandoned river channels (cutoffs) is beyond the state of the art for even late prehistoric times except in a few locations where archeological evidence is abundant and has been intensively studied. Readers concerned with *historic* period river channel changes should consult the compilation of comparative channel positions as determined from surveys dating to 1765 (Mississippi River Commission 1938) or the numerous other maps that exist.

Applications

This synthesis is envisioned as a basic source of information on the latest concepts and interpretations for those aspects of the Quaternary geomorphology and geology of interest primarily to engineering geologists. The level of detail of data presentation is oriented toward readers with an interest in the entire valley or major regions therein rather than specific locations. However, it should be also be useful to the latter because of reference to relevant literature and data sources.

The synthesis is intended for a technical, broadly interdisciplinary audience of scientists and engineers. Accordingly, it seeks to explain principles and concepts with minimal use of nomenclature and scientific jargon.

Geotechnical engineering applications of the synthesis should include site selection and evaluation of structures of a wide variety of types, analyses of general foundation conditions, location of construction materials, and evaluation of groundwater conditions. Hydraulic engineering applications should include studies of river behavior and potamologic considerations. Planning applications should include environmental assessments, wetlands distribution, cultural resources surveys, resource management, and habitat development activities.

Ideally, this synthesis should be one of the first references used, especially for those unfamiliar with the area. In a dynamic environment such as the Lower Mississippi Valley, a knowledge of natural processes and resulting landforms is an *essential* starting point in understanding man/land relationships. One cannot successfully accomplish environmental management, resource stewardship, or infrastructure development without understanding and appreciating landscape evolution.

2 Geographic and Physiographic Settings

The Lower Mississippi Valley

Definitions

The terms Lower Mississippi Valley, Mississippi alluvial valley, and Mississippi alluvial plain are in widespread use but unfortunately sometimes are erroneously used synonymously. While such terms are not rigidly definable as unequivocal physical features, they have subtle but significant differences and have changed in definition over time.

The Lower Mississippi Valley is that part of the overall Mississippi River system of the United States that lies between the latitude of Cape Girardeau and the Gulf of Mexico (Figure 1). Alternately, the nearby town of Cairo, Illinois, at the confluence of the Mississippi and Ohio rivers, is often regarded as the northern limit of the Lower Mississippi Valley. To the east, the boundary is distinct and is marked by bluffs separating the valley from dissected Coastal Plain uplands of Tertiary age. However, to the west, the boundary is difficult to define because of the merging of valleys of principal tributaries with the main Mississippi Valley.

This writer suggests that, for the purposes of this report, the Lower Mississippi Valley be defined as the greater Quaternary valley and deposits of the Mississippi River and its principal tributaries within the limits of the Coastal Plain (excluding the Red River in Texas). As shown in Figure 1, this would include the valley *per se*, the deltaic and chenier plains, fluvial and coastal terraces, and loess deposits of the uplands. Detailed mapping of these deposits is best presented in Saucier and Snead (1989).

The Mississippi alluvial valley was defined by Fisk (1944) as that area characterized by landforms and deposits created since the last glacial maximum. He regarded these as belonging to the geologic period called the Recent, which approximates but is not precisely the same as the Holocene. Russell (1940), from the perspective of responses to sea level variations, introduced the term Recent since it would be stratigraphically easier to recognize in the Gulf Coast area than the Holocene, which is defined on the basis of midwestern United States glacial stratigraphy. The former covers a time-span of about 18,000 years, while the latter is restricted to the last 10,000 years according to the views of most geologists.

Within the alluvial valley, Fisk recognized the alluvial plain and several older upland areas. The alluvial plain was defined as those areas (Holocene), such as the Mississippi River meander belts, that are (or were) subject to flooding by the rivers in their present regime. This definition is still valid. Upland areas included those of Tertiary age (e.g., Crowley's Ridge) as well as braided-stream terraces (valley trains) like Macon Ridge and those of the Western Lowlands that date to the last glacial cycle. The alluvial plain and the older uplands are easy to recognize and delineate. However, in Fisk's delineation of the alluvial valley, he also included several areas, such as the Grand Prairie, that are now believed to date to a still-earlier glacial cycle (the Sangamon Stage). Consequently, it is appropriate to refine and restrict the definition. Herein, the alluvial valley is considered (based on chronologic rather than geomorphic criteria) as that segment of the Lower Mississippi Valley that is characterized by landforms and deposits that are *primarily* of Wisconsin and Holocene age (Figure 1). There are still some problems with this definition in terms of certain terraces in Louisiana, as will be discussed later, but it is a generally satisfactory solution to this nomenclature problem.

In Louisiana, the alluvial valley merges with the deltaic plain, and in turn, the deltaic plain is contiguous with the chenier plain to the west (Plate 1). From a geographical perspective, a line drawn across the southern extent of bounding valley walls of pre-Wisconsin age separates the alluvial valley from the deltaic plain. This line is generally referred to as the Donaldsonville-Franklin line, named after the towns that approximate its limits. However, from a geologic perspective, the deltaic plain extends some 80 mi farther inland to the head of the Atchafalaya River (Figure 1). That stream represents the first (upstream-most) true Mississippi River distributary that discharges to the Gulf of Mexico.

General characteristics

One of the most popular and widely referenced concepts of Fisk (1944) is that the alluvial valley is a compound feature, i.e., a "valley-within-a-valley." He postulated that throughout the Quaternary, the Mississippi River formed a series of valleys, each of which was at a slightly lower elevation due to a progressive decline in sea level (the ultimate base level). Each successive valley was not only lower but also narrower, such that remnants of the older valley fills remained as terraces. During the last several decades, it has been determined that many of the older terraces of Fisk are pre-Pleistocene in age and not of Mississippi River origin (Autin et al. 1991). For this reason and because of other evidence, it is now apparent that the alluvial valley is indeed the lowest level, but in many areas it is as wide or wider than it has ever been in Quaternary times. Enough terrace remnants exist to make Fisk's concept valid in theory, but unfortunately misleading.

Between Cape Girardeau and the Gulf of Mexico, a distance of about 600 mi, the alluvial valley varies in width between about 30 and 90 mi (Figure 1). The deltaic plain extends about 150 mi in both east-west and north-south directions. As such, the alluvial valley has an area of about 33,450 sq mi and the deltaic plain an area of about 15,430 sq mi. The combined areas thus extend across parts of seven states, with the largest amounts being in Louisiana and Arkansas.

Within the alluvial valley, the Holocene alluvial plain occupies an area of 15,260 sq mi, only about 46 percent of the total. The remaining 54 percent of the valley is characterized by braided-stream terraces (valley trains) of Early and Late Wisconsin age. Valley trains dominate the alluvial valley landscape north of the latitude of Memphis, Tennessee, and occur to a lesser extent as far south as the Red River.

Average floodplain elevations of the alluvial valley decline from about 325 ft mean sea level (msl) in extreme southern Illinois to about 40 ft at the head of the deltaic plain, an average downvalley slope of only 0.6 ft/mi. Considered as a whole, the deltaic plain (and the adjacent chenier plain) has an average elevation of only about 5 ft due to the huge expanses of intratidal marshes. Average relief in the upper part of the alluvial valley approximates 25 ft and declines progressively southward. Uplands bordering the alluvial valley, and Crowley's Ridge within the valley, typically attain elevations of 200 ft or more above those of the adjacent floodplain. Upland elevations also steadily decline southward and eventually reach sea level at the Donaldsonville-Franklin line.

Alluvial Valley Segment

Major divisions

Upland remnants of Tertiary age and terraces and ridges of Wisconsin and pre-Wisconsin age serve to subdivide the Mississippi alluvial valley into six major lowlands or basins (Plate 1). In turn, each basin area is further subdivided into smaller units by ridges of Wisconsin and Holocene age. Each of the six major basins is a true topographic depression and definable hydrologic unit with a bounding interfluve. In all cases, drainage is from north to south into a major collecting stream after which the basin is named.

Readers should be aware that in the case of those major basins that are bounded by a Mississippi River meander belt, a significant part of the basin area actually consists of a part of the meander belt ridge itself. This is because the meander belt ridges, consisting of point bar accretion, abandoned channels, and natural levees, extend 5 to 10 mi from the river on both sides.

Western Lowlands. This basin is located in southeastern Missouri and northeastern Arkansas and is the second largest basin with an approximate area of 6,800 sq mi. It extends over a north-south distance of about 225 mi from the vicinity of Cape Girardeau to the Helena, Arkansas, area. The Ozark escarpment to the north and the Grand Prairie to the south form its western boundary while Crowley's Ridge and the Commerce Hills form its eastern boundary. It is predominantly an area of Early Wisconsin-age glacial outwash through which tributary streams and local drainage have formed narrow valleys and floodplains. Essentially all drainage enters the western side of the lowlands from the Ozark Plateau via (from north to south) the St. Francis, Black, Current, Spring, White, and Little Red rivers. Their flow combines to form the White River which discharges from the lowlands into the Arkansas River. The Cache River and Bayou De View are the only two streams of consequence that originate within the lowlands area.

From Cape Girardeau southward for about 50 mi, the Western Lowlands averages less than 10 mi wide and is interrupted by several small upland "islands" such as Hickory Ridge. From there to the vicinity of Jonesboro, Arkansas, the lowlands widen to about 20 mi and then abruptly widen again to more than 30 mi. Highest elevations occur adjacent to Crowley's Ridge and decline in a step-like fashion to the west. Lowest elevations occur along the floodplains of the Black and White rivers.

St. Francis Basin. This basin, also in southeastern Missouri and northeastern Arkansas, extends between the vicinities of Cairo and Helena, a distance of about 190 mi. It is bounded on the west by Crowley's Ridge and the Commerce Hills, and on the east by the present meander belt of the Mississippi River between the two above-mentioned towns. Throughout most of its length, the basin has a rather constant width of about 40 mi but sharply narrows south of Memphis, Tennessee. Roughly, the northwestern two thirds of the basin is characterized by Wisconsin-age glacial outwash and the southeastern one third by Holocene Mississippi River meander belts.

The principal stream of the basin is the St. Francis River which flows into it through a gap in Crowley's Ridge and discharges from it into the Mississippi River at Helena. Principal tributaries within the basin are the Little River, Pemiscot Bayou, and the Tyronza River. Lowest average floodplain elevations occur near the center of the basin along the Little River Lowland. Within the area of glacial outwash, elevations decline eastward away from Crowley's Ridge rather than westward as they do in the Western Lowlands.

The St. Francis Basin is sometimes grouped together with several small, narrow lowlands in extreme western Kentucky and Tennessee that lie between the present Mississippi River meander belt and the uplands to the east to form the Eastern Lowlands. However, this term is not widely used.

Yazoo Basin. This basin located in northwestern Mississippi is the largest in the Mississippi alluvial valley with an area of about 7,600 sq mi. It extends about 200 mi from Memphis to Vicksburg and is a little over 60 mi wide at the latitude of Greenwood, Mississippi. Uplands form the basin's eastern boundary throughout, and the present meander belt of the Mississippi River forms its western boundary. Glacial outwash deposits comprise less than 5 percent of the total basin area: the remainder consists of Holocene meander belt and backswamp environments.

Lowest points in the basin area occur just north of Vicksburg. Most streamflow comes from the Coldwater, Tallahatchie, Yokona, and Yalobusha rivers and several smaller streams that discharge from the uplands into the eastern side of the basin and combine to form the Yazoo River at Greenwood. The Yazoo River becomes tributary to the Mississippi River at Vicksburg. Interior drainage of the basin is by way of a complex system of sluggish streams that eventually join the Big Sunflower or Bogue Phalia rivers or Deer Creek and thence into the Yazoo River. Virtually all basin drainage is controlled by abandoned Mississippi River meander belt features or distributary channels.

Arkansas Lowland. This basin located in east-central Arkansas is the smallest of the major basins with an area of only about 1,325 sq mi. It extends approximately 75 mi from Little Rock to the small settlement of Arkansas Post. The Grand Prairie forms its northern and eastern boundary, and the present meander belt of the Arkansas River forms its western and southern limit. No major streams enter or originate within the basin, and all drainage is controlled by abandoned meander belts of the Arkansas River. All deposits and landforms are of Holocene age and are a product of the Arkansas River.

Boeuf Basin. It is described as a long and narrow depression that lies south of, and geologically is a continuation of the Arkansas Lowland in southeastern Arkansas and northeastern Louisiana. The basin extends in a sinuous fashion from the vicinity of Pine Bluff, Arkansas, southward for a distance of about 190 mi to Sicily Island, Louisiana. It is bounded on the west by uplands, on the north by the present meander belt of the Arkansas River, and on the east by Macon Ridge. Throughout much of its length, it has an average width of only about 10 mi and nowhere is more than 25 mi wide. Holocene deposits (predominantly backswamp deposits) and landforms originating with the Arkansas River characterize the entire basin area.

The dominant drainage feature of the northern two thirds of the basin is Bayou Bartholomew which occupies an abandoned course of the Arkansas River. Although the basin is named for the Boeuf River, this stream only enters the basin from Macon Ridge about 40 mi above its southern limit. In reality, the largest stream in the basin is the Ouachita River which enters its western side in the vicinity of Monroe, Louisiana, and flows south in an abandoned Arkansas River meander belt. The Boeuf River becomes tributary to the Ouachita River just north of Sicily Island.

Tensas Basin. The southernmost basin of the alluvial valley extends from near the mouth of the Arkansas River in eastern Arkansas to near the mouth of the Red River in east-central Louisiana, a distance of about 180 mi. The basin is bounded by Macon Ridge and uplands south of Sicily Island on the west, and the present Mississippi River meander belt on the east. The Tensas Basin is quite narrow but widens steadily to the latitude of Vicksburg, and thence varies in width between 25 and 45 mi. It is characterized entirely by Holocene meander belt and backswamp deposits.

Lowest elevations in the basin occur just east of Macon Ridge and in the southern part between Sicily Island and the Red River. The Tensas River and Bayou Macon drain the upper two thirds of the basin while Black River, which carries the flow of the Ouachita River south of Sicily Island, is the main stream in the lower part of the basin.

Crowley's Ridge. Though not a major division of the Mississippi alluvial valley *per se*, this ridge nevertheless is a feature of comparable significance. It is an upland remnant that separates what was once the main valley of the Mississippi River (the Western Lowlands) from the main valley of the Ohio River (the Eastern Lowlands). It extends as an unbroken ridge for about 125 mi between the Marianna and St. Francis River Gaps, but it actually extends for 215 mi between Helena and Thebes Gap if several geologically related isolated upland areas are included (including the Commerce Hills). The northern half of Crowley's Ridge has an average width of about 10 mi, but it narrows abruptly near Jonesboro and has an average width of less than 5 mi from there southward.

Topographically, Crowley's Ridge closely resembles the maturely dissected uplands of similar origin that lie to the east of the Mississippi alluvial valley. It attains elevations of 100 to 250 ft above the adjacent Wisconsin-age valley trains, and the greatest local relief occurs south of Jonesboro where the ridge is narrowest.

Grand Prairie and Macon Ridge. While not in the category of major alluvial valley divisions, these two ridges smaller than Crowley's Ridge are worthy of discussion because they serve as important interfluves between major basins. The Grand Prairie in east-central Arkansas projects southeastward from the Tertiary-age uplands east of Little Rock for a distance of about 70 mi. This Pleistocene terrace has a rather constant width of about 25 mi and has an elevation that is 20 to 40 ft higher than the adjacent Arkansas Lowlands to the west or the White River Lowland to the east. The Grand Prairie was created primarily during the Sangamon Stage by the Arkansas River. Macon Ridge in extreme northeastern Louisiana and southeastern Arkansas is a 135-mi-long prominent ridge that lies between the Boeuf and Tensas Basins. It reaches a maximum width of about 25 mi in northeastern Louisiana about 30 mi north of Sicily Island. The ridge is consistently higher on its eastern side where elevations are 20 to 30 ft higher than in the adjacent Tensas Basin. On the western side, elevations of the ridge are approximately the same as those in the Boeuf Basin, and it is sometimes difficult to distinguish the two at the surface. Macon Ridge consists almost entirely of Early Wisconsin-age glacial outwash and is a continuation of the valley train in the Western Lowlands.

Wisconsin and Holocene uplands and ridges

The five topographically prominent landforms of Wisconsin or Holocene age within the Mississippi alluvial valley considered important in its history and development are: Flatwoods Terrace, Marksville Hills, Sikeston Ridge, Charleston Fan, and Tiptonville Dome (Plate 1). They actually represent four different modes of origin, however.

The Flatwoods Terrace in the lower Boeuf Basin and the Marksville Hills (also known as the Avoyelles Prairie) just south of the present course of the Red River in Louisiana are isolated units of once more continuous fluvial terraces. They have been left as "islands" of high ground by subsequent lateral and vertical stream erosion. In the St. Francis Basin in southeastern Missouri, Sikeston Ridge is a well-defined, 30-mi-long erosional remnant of Early Wisconsin Ohio River glacial outwash that is surrounded by Late Wisconsin Mississippi River glacial outwash. Immediately east is the Charleston Fan which is a 100-sq-mi discrete unit of Late Wisconsin glacial outwash thought to be attributable to a brief pulse of meltwater flow through Thebes Gap. To the south lies the Tiptonville Dome which is situated between the present Mississippi River and the uplands to the east. Together with the adjacent, smaller Ridgley Ridge to the south, they constitute what some workers refer to as the Lake County uplift. The compound feature, lying about 10 ft above surrounding areas, is the result of neotectonic surface deformation during the New Madrid earthquake series of 1811-1812.

Holocene meander belts and distributaries

As will be discussed in detail later, it is highly characteristic for streams that carry large suspended-sediment loads and periodically overtop their banks and build natural levees to occasionally divert flow to new courses. A short-lived and relatively small diversion creates only a crevasse splay. A more persistent diversion usually leads to the development of a distributary channel smaller than the parent stream, and a permanent diversion leads to the creation of a new meander belt and eventual capture of the parent stream. The Holocene alluvial plain of the Mississippi alluvial valley is dominated by abandoned distributaries (not to be confused with deltaic distributaries) and meander belts of the Mississippi, Arkansas, and Red rivers. Each is a low, broad ridge that is a mile to several miles wide, at least several tens of miles long, and 5 to 10 ft higher than the adjacent floodplain areas.

As shown in Plate 1, there are 34 discrete segments of abandoned meander belts and five distributaries of the three major rivers that determine the physiography of the major basin areas. Correlation of these segments from basin to basin to determine their relative and numerical ages has been and continues to be perhaps the most difficult problem in alluvial valley chronostratigraphy. The correlations shown in Plate 1 form the best available chronologic model and also the basis for discussions presented later in this report. Proper names assigned to the distributaries and meander belt segments are for convenience in discussions and are based on the largest or most continuous streams presently occupying the segments. It should be noted that in some cases even short segments may be occupied by several named streams or a single stream with different names for different segments or reaches.

Wisconsin and Holocene lowlands and gaps

Twenty-four second-order lowland areas are present in the several major basin areas of the Mississippi alluvial valley and are generally recognized by proper names (Plate 1). Most of them fall into four categories based on mode of origin. Some are the Holocene-age valleys of streams of moderate size such as the Black River and White River Lowlands within the Western Lowlands. Some are the products of shifting and downcutting by braided streams during episodes of valley train deposition, such as the Cairo Lowland and the Morehouse Lowland in the St. Francis Basin. Some like the Bogue Phalia Basin and the Quiver River Lowland in the Yazoo Basin are semienclosed depressions lying between abandoned meander belt segments or distributary ridges. Those lowland areas exist because they represent areas with relatively lower sedimentation rates. Finally, others such as the Catahoula Lake Basin and the Dismal Swamp Lowland are depressions of multiple origin that are sufficiently low or poorly drained to contain permanent lakes or swamps. Numerous gaps in upland ridges occur in all parts of the alluvial valley, but only five are generally regarded as significant geographical features (Plate 1). Each apparently developed as a result of the headward erosion of one or more small ephemeral streams, but each is significant because all were later occupied by larger streams that were able to achieve a gradient advantage by diverting through them.

Deltaic and Chenier Plain Segments

Uplands, distributary ridges, and cheniers

In areas as low and flat as the deltaic and chenier plains, landforms that are as little as 5 to 10 ft above the surrounding landscape are often visible for miles, and a difference in elevation of inches may mean the difference between marsh and forest vegetation. Therefore, prominent ridges are exceptional and conspicuous.

Excluding outliers of the coast-parallel terrace that forms the northern limit of the deltaic and chenier plains, there are only four features of pre-Holocene age that are present. These features are Avery Island, Belle Isle, Cote Blanche Island, and Weeks Island, which are four of the five so-called Five Islands of south Louisiana (Plate 1). Each is a piercement-type salt dome about a mile in diameter in which diapirism has uplifted Pleistocene deposits of Wisconsin age to elevations as high as about 150 ft above sea level. The fifth of the salt domes, Jefferson Island, while not technically in the Holocene deltaic plain, is sufficiently close to be included herein.

The vast expanse of the deltaic plain was formed by deltaic progradation and the coalescing of shifting delta complexes in which abandoned distributaries are numerous. Following the precedent of Frazier (1967), the term *delta complex* is used in preference to the terms *subdelta* or *delta lobe* which, although in widespread use, are less definitive. In general, the distributaries radiate in a fan shape from the apexes of the delta complexes, declining in width and elevation with increasing distance. Since subsidence has been a prevailing regional process during the Quaternary, there is a direct relationship between the size of distributaries and their age. The youngest are the topographically most prominent while the oldest, if not actually buried, are hardly discernible features.

The distribution of deltaic environments of deposition as shown in Plates 12, 13, and 14 represents a generalization of the pattern of deltaic distributaries with many of the smaller ones omitted because of scale limitations. A further generalization is portrayed by Plate 1 in which only four are identified because of their size and strategic location. Next to the present course of the Mississippi River below Donaldsonville, the Bayou Lafourche distributary is the youngest and has the best developed natural levees. Farther east, where the delta complex is older and distributaries are fewer, the Bayou des Familles, Bayou la Loutre, and Metairie Bayou-Bayou Sauvage distributaries and their natural levee ridges are prominent landforms despite the fact that elevations do not exceed 10 ft.

Abandoned distributaries are the characteristic topographically prominent landforms of the deltaic plain, whereas cheniers fill this role in the chenier plain. Cheniers are stranded Gulf of Mexico beaches within large expanses of Holocene marsh that developed on prograding mudflats. A typical chenier ridge is less than 10 ft high, a few hundred feet wide, but miles or tens of miles long. In prehistoric times, the ridges supported a forest vegetation dominated by live oaks: the term chenier is derived from the French word, "chene," meaning oak. Several series of coast-parallel cheniers are present and are discussed more fully later.

Interdistributary lowlands and basins

Atchafalaya Basin. Geologically and geomorphologically, this basin of south-central Louisiana represents a transition between the Mississippi alluvial valley and that portion of the deltaic plain characterized by distributary ridges and intratidal marshes. This largely uninhabited and forested area of backswamp of about 3,800 sq mi in extent (25 percent of the total area) is the single largest major division of the deltaic plain. It extends for over 120 mi between the mouth of the Red River to the vicinity of Houma, Louisiana, and has maximum width near Baton Rouge, Louisiana, of about 50 mi. The Teche meander belt ridge forms the western boundary of the basin, and it is bordered on the east by the present Mississippi River meander belt and the Bayou Lafourche distributary.

The Atchafalaya River, currently a major distributary of the Mississippi River, is the principal stream within the basin and flows southward through the approximate center of the basin. The river breaches the Teche meander belt ridge at Morgan City, Louisiana, and discharges into Atchafalaya Bay. In prehistoric times, the discharge of the river was augmented with substantial flood runoff from the Mississippi and Red rivers into the upper part of the basin. Extensive shallow lakes existed up to several decades ago in the southern part of the basin, but they have been rapidly filling because of increased Mississippi River discharge.

In contrast to the Atchafalaya Basin which basically is a freshwater environment, the remainder of the deltaic plain and the chenier plain are primarily brackish to saline. The present deltaic plain landscape is composed of several large interdistributary basins dominated by intratidal marshes and numerous shallow lakes and ponds. Each basin has a small area of freshwater swamp in the inland-most portions and becomes progressively more saline seaward. There are numerous smaller interdistributary depressions in the deltaic plain, but because most are fan-shaped seaward and are not constricted in that direction, they are not formally designated basins. An example is the depression that lies south of the Bayou la Loutre distributary and east of the present Mississippi River that opens out into Chandeleur Sound.

Pontchartrain Basin. Bounded by pre-Holocene terraces to the north and the present Mississippi River meander belt and the Metairie Bayou-Sauvage distributary to the south, this basin is about 80 mi long and 30 mi wide. Lake Pontchartrain, the largest lake in the deltaic plain, is the dominant feature of the basin and is connected with the Gulf of Mexico by two passes at its eastern end. Several small streams discharge into the basin from the uplands to the north and west, notably the Amite River, but no natural drainage enters from the south.

Barataria Basin. With a length of about 90 mi, this basin is bounded on the west by the Bayou Lafourche distributary, the Mississippi River meander belt on the north, and the Bayou des Familles distributary and the Mississippi River on the east. No streams enter the basin, and local runoff moves through Bayou des Allemands into Lake Salvador and thence through a series of bayous and lakes into Barataria Bay and the Gulf of Mexico.

Terrebonne Marsh. The large expanse of marsh south and west of Houma has been designated as such to facilitate subsequent discussions although it is not formally designated. This deltaic plain unit lies south of the Teche distributary ridge and east of Atchafalaya Bay and encompasses a number of small distributaries on its eastern side (see Plates 14 and 15).

Adjacent Uplands

The Mississippi alluvial valley is directly bordered by older formations and deposits of the Gulf segment of the Coastal Plain Province except to the northwest where it lies adjacent to the geologically much older Ozark Plateau (Figure 1). Northwest of the linear and well-defined Ozark Escarpment, the plateau is a broad uplift or dome consisting of several subplateaus formed predominantly on limestones and dolomites ranging in age from Cambrian to Pennsylvanian (Figure 3). These rocks are maturely dissected and are capped with deep residual soils containing large amounts of chert. The larger streams are incised several hundred feet into the plateau, are bordered by steep bluffs, and have narrow floodplains typically less than a mile wide. Numerous large springs significantly augment the discharge of most of the streams that enter the alluvial valley. In general, the Ozark Plateau contains the most rugged topography and the largest amount of rock outcrops in close proximity to the alluvial valley.

The Ozark Plateau is bordered on the south by the 25- to 35-mile-wide, east-west trending Arkansas River valley, and in turn, the valley is bordered on its south by the Ouachita Mountains. The latter physiographic unit consists of folded and faulted sandstones and shales of Pennsylvanian and Mississippian age occurring as east-west trending ridges and valleys. These formations lie adjacent to the Mississippi alluvial valley in only a small area near Little Rock.

The Western Hills portion of the Gulf Coastal Plain lies south of the Ouachita Mountains and immediately west of the Mississippi alluvial valley from Little Rock south to near Alexandria. This physiographic unit consists of moderately dissected, largely unindurated deposits of Tertiary age that outcrop as broad belts paralleling the Coastal Plain margin. Being an eroded offlapping sequence, the belts are progressively younger in age toward the south. The few formations that contain lithified materials form prominent cuestas or wolds. One such cuesta, the northeast trending Kisatchie Wold, consists of Miocene and Oligocene siltstones and sandstones and is responsible for the marked narrowing of the Mississippi alluvial valley at the latitude of Sicily Island. Maximum relief is generally in the range of 50 to 100 ft, and the major streams have valleys that are several miles wide. Because of abundant supplies of sediment from easily eroded deposits and progressive stream downcutting since Tertiary times, the larger stream valleys typically are flanked by broad fluvial terrace sequences of Quaternary age.

The Eastern Hills portion of the Gulf Coastal Plain lies immediately east of the alluvial valley from the mouth of the Ohio River south to between Natchez, Mississippi, and Baton Rouge. This moderately dissected upland area closely resembles the Western Hills in terms of relief, topography, lithology, and origin; however, it differs in that the Tertiary formations are overlain over large areas by late Tertiary and Quaternary fluvial and eolian deposits. Deposits of fluvial sands and gravels of Pliocene or younger age cap many hills in a 10- to 50-mile-wide belt adjacent to the valley from Kentucky to Louisiana.

RELATIVE GEOLOGIC TIME				EST. TIME (MY)
ERA	PERIOD		EPOCH	
CENOZOIC	Quaternary		Holocene	
			Pleistocene	2-3
	Tertiary		Pliocene	12
			Miocene	26
			Oligocene	37-38
			Eocene	53-54
			Paleocene	65
MESOZOIC	Cretaceous		Late Early	136
	Jurassic		Late Middle Early	190-195
	Triassic		Late Middle Early	225
PALEOZOIC	Permian		Late Early	280
	Carboniferous Systems	Pennsylvanian	Late Middle Early	
		Mississippian	Late Early	345
	Devonian		Late Middle Early	395
	Silurian		Late Middle Early	430-440
	Ordovician		Late Middle Early	500
	Cambrian		Late Middle Early	570
PRECAMBRIAN				3600

Figure 3. Standard geologic column applicable to the Lower Mississippi Valley area

In turn, a 30- to 60-mi-wide belt of Late Quaternary loess overlies the Tertiary and older Quaternary deposits immediately east of the alluvial valley margin through this same area. Several discrete loess sheets have been recognized that together form an eastward-thinning belt which reaches maximum thicknesses of more than 75 ft in the Vicksburg-Natchez area. Especially near the immediate upland margin or bluff, the loess deposits are deeply dissected and form extremely rugged topography with local relief of over 100 ft.

Between the Western Hills and the chenier plain on the western side of the alluvial valley, and between the Eastern Hills and the deltaic plain on the eastern side, the uplands consist of slight to moderately dissected coast-parallel terraces of fluvial and marine origin. These also exist as offlapping sequences in which the older terraces are farther north and higher in elevation. Local relief is as much as several tens of feet on the older terraces but is less than 5 ft on the youngest ones adjacent to the chenier and deltaic plains. In Louisiana, the Great Southwest Prairies is the physiographic term applied to the younger coastwise terraces west of Lafayette, Louisiana (Plate 1). East of the Mississippi River, the coastwise terraces occur in what is generally called the Florida Parishes portion of the state.

3 Geologic Processes and Controls

Virtually all Lower Mississippi Valley landforms and deposits are the result of fluvial, eolian, or marine processes. Of these, fluvial processes are predominant and include the products of both inorganic and organic sedimentation. All three geomorphic processes function in both constructional and erosional modes: at any given time, landforms are both being created and destroyed. The particular landforms created are the result of variations in the quantity and physical characteristics of the sediment, the energy of the transporting medium, and the environmental setting. For example, silt being transported as suspended load in floodwaters overtopping a streambank will build a natural levee in a broad, flat basin, whereas similar material concentrated in a small stream valley will develop an alluvial fan at its mouth. Wave action working on a large supply of sand will create a beach, whereas similar wave action without a sediment supply will result in shoreline erosion. Specific landforms and geomorphic processes are discussed in detail in Chapter 6 of this report.

Geomorphic processes determine how, when, where, what, and if sediments are deposited. Landscapes, however, involve the products of geomorphic processes in the context of regional settings and geologic controls. Scale and time become important factors. Meander belts, for example, consist of numerous individual natural levees, abandoned channels, and other landforms, but the types, number, and distribution of meander belts depend on valley size, shape, slope, interactions among meander belts, and time. With increasing size of features or landscapes and increasing age, various broad, regional controls or geologic processes become increasingly important. The following sections of this chapter discuss the nature and effects of five regional controls that have strongly influenced how individual landforms have responded to produce different landscapes.

Glaciation, climate, relative sea level, tectonism, and subsidence are regarded as the dominant geologic controls influencing the Lower Mississippi Valley area during the Quaternary. The rate of introduction and the composition, texture, and quantity of the sediment supply to the area is recognized as another independent variable, but with the exception of those major changes brought about by waxing and waning glaciation (see following section), this control is difficult to assess and quantify and is not explicitly discussed.

Emphasis throughout this chapter is on the variable nature of these regional geologic controls over time. If any of these controls, such as climate or sea level position, had remained constant, the resulting landscapes would be strikingly different. Despite a natural tendency to do so, essentially no landscapes have achieved a state of equilibrium or maturity in the sense that a life cycle has been completed. Rather they are in a state sometimes referred to as dynamic equilibrium wherein only a temporary balance exists between process and form.

Readers are cautioned that this discussion of processes and controls refers only to those of a *geologic* nature. Possibly the most consequential influence of all--that of humans--is not discussed directly. Anthropogenic changes, sometimes dramatic, have directly and indirectly affected the physical landscape in all parts of the Lower Mississippi Valley. These include tillage, drainage, channelization, land clearing, leveeing, dredging, reclamation, mining, groundwater withdrawal, etc. Humans have not eliminated any geomorphic processes or introduced any new ones; however, they have greatly changed the rates at which most have been operating. Moreover, humans have functioned as a trigger mechanism in changing the timing of when geomorphic thresholds are crossed.

Continental Glaciations

Without question, the single most significant geologic process to affect the geomorphology of the Lower Mississippi Valley has been continental glaciation. Indeed, had it not been for the disarrangement of preglacial drainage in the midcontinent area by the first glaciation, the Mississippi River system with the Ohio and Missouri rivers as major tributaries would not exist (Fisk 1944). Ice sheets that formed during the Pleistocene in the midwestern United States did not extend southward beyond southern Illinois and hence did not directly affect the valley. However, the valley repeatedly served as the sluiceway through which huge quantities of meltwater and outwash were funnelled to the Gulf of Mexico. Farther south, coastal areas responded to severe cyclical variations in sea level. Correlation of deposits and landforms in all parts of the Lower Mississippi Valley with those of the glaciated midwestern United States therefore is vital but is complicated by a lack of a direct stratigraphic link (Autin et al. 1991). It is also apparent that the problem is made even more complex by the fact that responses within the valley to this geologic control varied both spatially and temporally.

Three basic issues are involved in understanding the influence of continental glaciations and are discussed below. These issues are (a) identifying the specific ways in which various portions of the area responded, (b) determining the number and magnitude of the glaciations, and (c) determining the chronology of the glacial sequence.

Glacial response model

For decades, the accepted view of the geomorphic response of the valley area to continental glaciations has been the glacioeustatic model introduced by Fisk in 1944. In that model, a major fall in sea level during glacial stages caused deep and extensive valley entrenchment, whereas a major rise in sea level during interglacial stages initiated valley alluviation and eventually deltaic progradation. Since there was a secular decline in interglacial-stage sea level positions during the Pleistocene and uplift in interior areas, a net result of glacioeustatic responses was the formation of a suite of fluvial terraces with each successively lower terrace representing a younger interglacial stage.

Portions of this process-response model are no longer considered valid, and in general, it is regarded as overly simplistic (Autin et al. 1991). The model overestimates the impact of sea level variations on the more inland portions of the valley, incorrectly interprets the role of outwash deposition in valley formation, and does not consider the strong influence of climate. For example, it is now widely recognized that certain terraces formed primarily during times of waxing (expanding) glaciation rather than during waning (retreating) glacial conditions or interglacial periods.

The currently accepted process-response model is shown in Table 1. It differentiates between the responses that took place in the major physiographic units during a typical glacial-interglacial cycle and in reality is an integrated geologic/eustatic/climatic model. The model is time-dimensionless because the various glacial cycles varied appreciably in length, but it does illustrate how different areas responded in different--and sometimes opposite--ways at a given point in a cycle. It is important to consider that process-response relations change between global and local scales. For example, with respect to climate, most of the Lower Mississippi Valley area clearly recorded the disintegration of the last continental (Laurentide) ice sheet, but it may or may not have recorded minor sea level fluctuations and probably rarely recorded local climatic events.

Glacial sequence

The traditional model of glacial chronostratigraphy for the midwestern United States, based on the European model developed before the turn of the century, recognizes five major glacial stages (from oldest to youngest, the Nebraskan, Kansan, Illinoian, Early Wisconsin, and Late Wisconsin) and five interglacial stages (from oldest to youngest, the Aftonian, Yarmouthian, Sangamonian, Peorian--also called Mid-Wisconsinan or Farmdalian, and the Holocene). This is the model used by Fisk (1944) as the basis for identifying the four major fluvial terraces of the Lower Mississippi Valley (from oldest to youngest, the Williana, Bentley, Montgomery, and Prairie).

The model was developed primarily using terrestrial glacial stratigraphy from the United States Midwest and, at least in early years, assumed each

TABLE 1. PROCESS MODEL SHOWING REGIONAL RESPONSES TO BASIC GLACIAL/INTERGLACIAL CYCLE IN THE LOWER MISSISSIPPI VALLEY (FROM AUTIN ET AL.1991)

Glacial Cycle		Sea Level Response	Coastal/Deltaic Response	Alluvial Valley Response	Tributary Stream Response	Upland Response
Interglacial		Highstand	Deltaic and Chenier Plains	Aggradation	Stability and Soil Formation	Slow Degradation Soil Formation
		Minor Oscillations		Meander Belts and Soil Formation		Soil Formation
			Delta Lobes on Shelf	Minor Degradation	Meander Belt Formation	
			Rapid Shoreline Transgression			
	Waning Glaciation	Rising		Valley Train Development	Aggradation	Loess Deposition (Local Aggradation)
			Trench Filling		Possible Alluvial Drowning in Lower Reaches	
				Maximum Aggradation	Instability	
Glaciation	Glacial Maximum	Lowstand	Broad Exposed Shelf	Outwash Deposition and Initial Aggradation		
			Shelf-Margin Deltas		Degradation	Major Erosion and Dissection
				Degradation	Terrace Formation	
			Stream Entrenchment and Extensions		High Discharges	
	Waxing Glaciation	Falling				
			Rapid Shoreline Regression	Stream Regime Change (Meandering to Braided)	Regime Adapts to Increasing Discharges	Slow Degradation

stage involved a single major ice advance and a subsequent retreat. However, during the last several decades, considerably more definitive evidence for developing a glacial sequence has come from studies of loess stratigraphy, pollen cores, deep-sea sediment cores, and ice-cap cores. While allowance was made for far greater resolution of events, this new body of evidence has complicated the picture in one respect. The Mississippi alluvial valley has responded to glacial conditions just in the midwestern United States, whereas the coastal area has responded to sea level conditions. The latter is a reflection of glacial conditions worldwide, and it is known that individual ice sheets were not synchronous in all parts of the globe.

Based on oxygen isotope data from deep-sea cores, currently the most definitive global-scale chronostratigraphic technique, there were at least 17 complete glacial-interglacial cycles during a period that might represent little more than the last half of the Quaternary period (Morrison 1991). Loess sequences in China indicate at least 44 cycles occurred during the Quaternary. It must be realized, however, that all of these cycles were not of comparable duration or amplitude and represent a substage (stade) rather than a full glacial stage.

It has recently been pointed out (Anklin et al. 1993) that for most of the Quaternary, climatic conditions were much less stable with more frequent major oscillations (and sea level variations) than during the last 10,000 years. Despite this complexity, there is still considerable merit in using a generalized model or sequence for the Lower Mississippi Valley. The model presented herein (Figure 4) is adapted from Beard, Sangree, and Smith (1982), who developed it for the Gulf Coast region using paleontologic, sedimentologic, and seismic evidence. Eight major glacial cycles are recognized and correlated with traditional stage nomenclature. Although many may regard this model as overly simplistic, it is compatible with the state of knowledge of Lower Mississippi Valley chronostratigraphy. Considering an inability, for example, of even confidently correlating whole terrace complexes with the correct major glacial stages, it is meaningless at this time to contemplate valley responses to more subtle glacial variations. A more complex model may be extremely valuable for correlations within the thick sedimentary sequence underlying the deltaic plain and continental shelf, but it has little current relevance to geomorphic considerations related to engineering applications.

The Beard, Sangree, and Smith (1982) eustatic cycle model appears to generally correlate with the traditional midwestern glacial model (Figure 4), but use of the traditional nomenclature for conceptual stage names presents problems. Boellstorff (1978) and other workers regard the model as overly simplistic and advocate redefining or abandoning the nomenclature entirely. Both Richmond and Fullerton (1986) and Morrison (1991) have proposed slightly differing versions of a revised model based on the latest interpretations of midwestern glacial stratigraphy. An attempt to merge the two is presented in Figure 4. It will be noted that the names for stages older than the Sangamonian have been abandoned in favor of a new nomenclature and those younger have been redefined. In this synthesis, most discussions are in terms

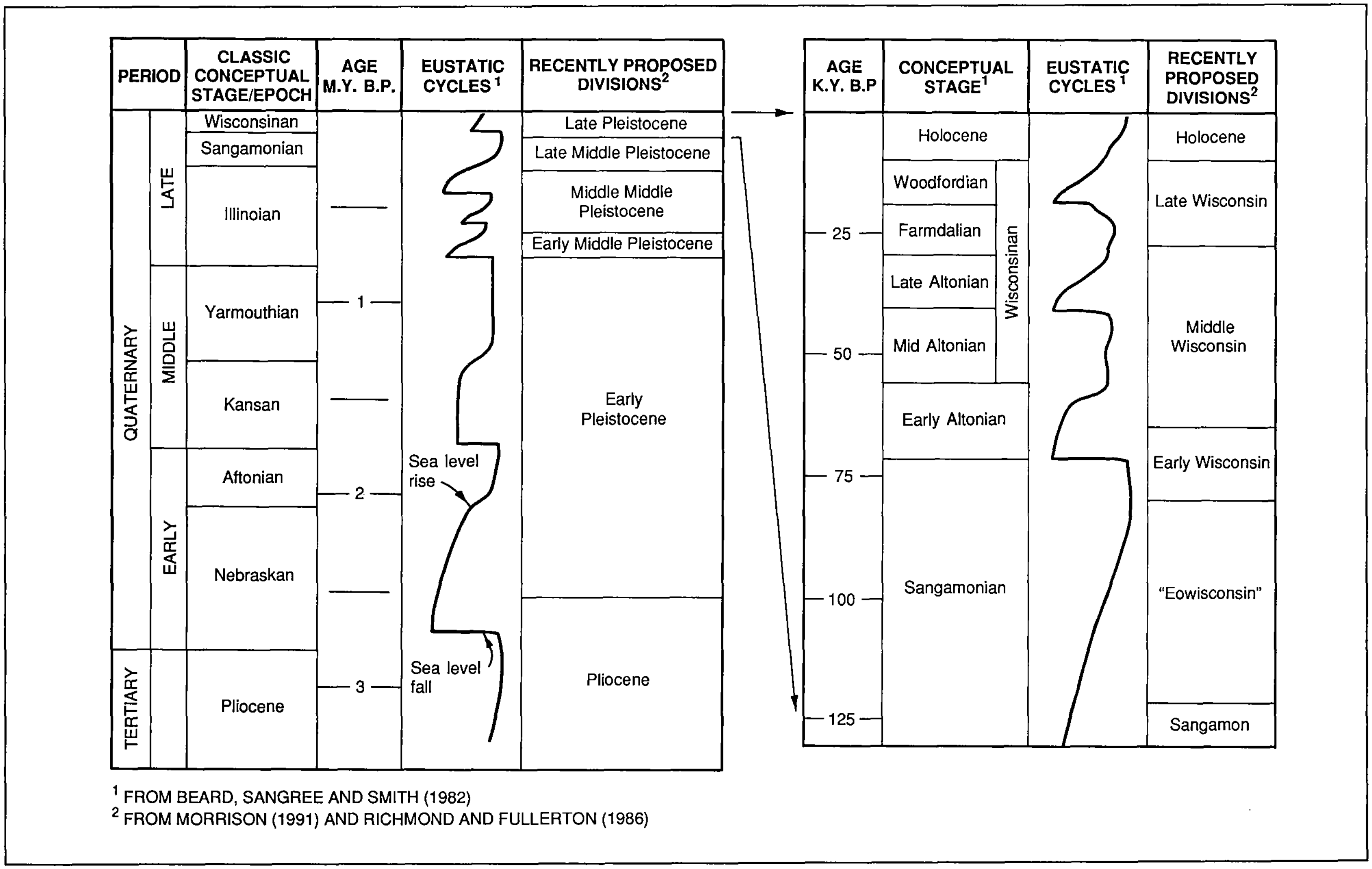

Figure 4. Geologic column for the Quaternary period showing conceptual stages and a generalized sea level curve indicating eustatic cycles (modified from Beard, Sangree, and Smith 1982)

of the new nomenclature, but references to older names are sometimes included where confusion may result from prior use in cited literature.

Glacial chronology

As recently as the 1950s, estimates of the total length of the Quaternary period ranged from as little as 500,000 to 600,000 years to a widely recognized maximum of about 1.2 million years. With the coupling of oxygen isotope data with radiometric dates (fission-track method) and paleomagnetic stratigraphic interpretations, the Tertiary (Pliocene)-Quaternary boundary has been progressively pushed back to as far as 1.65 million years (Richmond and Fullerton 1986) to 2.8 million years (Beard, Sangree, and Smith 1982). A value of 2.48 million years is used in this report (Figure 4) based on the interpretation of Morrison (1991). Readers should be aware, however, that the indicated dates for the various stages are controversial within the profession. Despite the availability of proportionately many more radiometric dates, there is even disagreement over the ages of the more recent stages but largely because of differences in stage definition.

The Pleistocene-Holocene boundary is less controversial although it is arbitrarily defined rather than based on a distinctive stratigraphic break. Both geologists and archeologists generally place the boundary at 10,000 years B.P. (Hopkins 1975), but the date of 12,000 years occasionally appears in the literature. The date of 10,000 years has been proposed for formal adoption even though it does not coincide with an abrupt climatic change, and sea level was appreciably below its present level and rising at a rapid rate. Essentially, no geologists currently accept the position advocated by Russell (1940) that the Holocene (which he called the "Recent") should be defined as postdating the last glacial low stand of sea level (now estimated at about 18,000 years B.P.) despite the logic for its use in the Gulf Coast area.

Emphasis in this report is strongly oriented toward the Sangamon and Wisconsin stages (Figure 4). Although these stages represent perhaps less than 6 percent of Quaternary time, the preponderance of observable evidence within the Lower Mississippi Valley pertains to this 130,000+-year time span. In turn, since nearly half of the area of the alluvial valley and all of the deltaic and chenier plains were formed during the Holocene, discussions are dominated by events of the last 10,000 years.

Climate

Considerations in this chapter pertain only to the primary effects of climate and climate change during the Quaternary period. They do not include discussion of secondary effects such as sea level fluctuations or changes in river regime (from braided to meandering) because of the growth and decay of continental ice sheets in response to global-scale climatic change.

Climate reconstruction

Dating back to the mid-nineteenth century, early interest in Pleistocene climate focused on temperature changes, especially during glacial stages in near-glacial areas. Toward the end of the last century, recognition of pluvial lake conditions in the western United States caused attention to be focused on precipitation changes and the interrelationships between temperature and precipitation. Evidence of the magnitude and regional patterns of climate change developed slowly thereafter and was based largely on stratigraphy. A major upsurge in interest and advance in knowledge did not take place until the 1970s when modern palynological techniques were developed and implemented. Since then, climate change has been deduced primarily from paleovegetation reconstructions from cores in lake sediments and peat deposits and based on comparisons to modern analogs. A second upsurge in interest recently has been triggered by concern about the "greenhouse effect" and the climatic effects of anthropogenic aerosols, as well as aided by the numerical modeling of atmospheric circulation patterns and atmospheric-oceanic interrelations.

However effective these techniques have been, they have been applicable mostly to just the latter part of the Wisconsin stage. Therefore, climatic conditions in earlier glacial stages can only be inferred from application of a glacial-cycle model. Vegetation reconstructions using palynological records in the Lower Mississippi Valley area first began less than two decades ago (Delcourt and Delcourt 1977), and the database for this area is still sparse compared to adjacent regions. There can be no doubt that paleoclimatic reconstruction is a developing science and very much in its infancy with regard to the Lower Mississippi Valley.

Traditionally, investigations of the geomorphic role of climate have been most concerned with gradual, long-term changes in mean annual precipitation, i.e., periods of centuries or longer when conditions were generally wetter or drier than at present. There can be no doubt that these conditions occurred periodically throughout the Quaternary and had consequential geomorphic effects, but more recently attention has turned to other types of responses. Knox (1983) has been a leading proponent of major fluvial responses to rather abrupt changes in the frequency and intensity of storminess and consequent flooding due to changes in atmospheric circulation from zonal to meridional regimes. A cursory consideration (Saucier 1985) has shown that there may be a correlation between the onset of episodes of increased or decreased flooding frequency in the Midwest and the probable dates of the initiation of new meander belts and delta lobes in the Lower Mississippi Valley. This is an intriguing area for future research.

Effects of climate change

Direct effects of climate on area landscapes involve changes in erosion and sedimentation rates and patterns and changes in stream regimes. The latter are triggered by changes in the relationship between discharge and sediment concentration. Development of a conceptual model of the response of a fluvial system to climate change has been difficult because of delayed response times, asynchronism of response within a system, and variations from one system to another (Blum 1990), but Knox (1983) has had considerable success for streams in the Midwest. Recognition of climate as a significant causal mechanism in the Lower Mississippi Valley has been slow because the various manifestations in area fluvial systems can be attributed with equal or greater confidence to other processes.

Indirect effects of climate change include changes in vegetation communities. Significant increases or decreases in ground cover can cause landscape stabilization or destabilization with resultant changes in the rates and patterns of erosion and deposition. It has long been recognized that changes such as from forest to grassland vegetation cause appreciable changes in sediment yields from uplands.

In the Lower Mississippi Valley, the most obvious indication of climate change is not within the fluvial systems but rather is the extensive blanket of loess. Advocating the views of Russell (1944), Fisk (1951) did not accept that loess was eolian, but evidence to the contrary is overwhelming and an eolian origin is now universally accepted as the only viable explanation. Supporting an eolian origin of loess has been the more recent recognition of large areas of sand dunes in the upper portions of the alluvial valley (Saucier 1978). As will be discussed later, loess is the direct result of deflation of silt from glacial outwash deposits (valley trains) by seasonally strong, primarily northerly and northwesterly, late glacial-stage winds.

Recognition of possible effects of climate change on the Holocene Mississippi River meander belts and floodplain has been a difficult and frustrating endeavor pursued by this writer for several decades. Perhaps the biggest problem is the apparent insensitivity (higher response thresholds) of a system as large as the Mississippi River to climatic influences and its tendency to respond in a diffuse rather than an abrupt or sharp manner. Distinctive differences in abandoned channel and course size and geometry between meander belts have long been considered (Saucier 1974, 1985) as possible evidence of response of the Mississippi River to climatically-induced changes in discharge; however, pieces of the puzzle simply do not fit and a satisfactory model has not emerged.

Evidence of fluvial responses to climate change is apparently abundant on Gulf Coast streams such as the Sabine and Pearl rivers (Figure 1) that directly enter the Gulf of Mexico. Evidence also occurs on some Mississippi River tributaries, notably the Ouachita River. On all of these streams, the Deweyville complex (Bernard 1950, Gagliano and Thom 1967, Saucier and Fleetwood 1970) is considered by many workers to be the direct result of a major stream regime change caused by glacial-stage increased precipitation and runoff. Therefore, this distinctive and conspicuous terrace complex would not be the product of interglacial-stage, high-sea level alluviation of the type envisioned by Fisk (1944) in his glacial-stage model.

Discussion of the climatically-induced responses of smaller upland streams is beyond the scope of this report. It has traditionally been accepted that alluviation accompanies wetter climatic episodes while floodplain stability and soil formation accompany drier climatic episodes. That generalization is at best an oversimplification. As pointed out by Schumm (1973), many drainage systems experience complex responses in which increased erosion and headward extension occur in upland tributaries at the same time that alluviation occurs in the main valleys. In the uplands adjacent to the Lower Mississippi Valley, such possible indicators of climatic responses as cut and fill sequences and buried paleosols are definitely present and widespread and have been investigated (including dating) by Grissinger, Murphey, and Little (1982) and others. Discernible evidence of climate effects is not present in the uplands *per se*, and there is no reason to believe that there have been dramatic vegetation changes in Quaternary times.

Climate change record

Eight radiocarbon-dated paleoenvironmental records from Arkansas (3), Louisiana (1), Missouri (3), and Tennessee (1) were determined from pollen and sedimentary sequences in cores. Those records constitute essentially all the evidence for climate change in the Lower Mississippi Valley area (Royall, Delcourt, and Delcourt 1991). These data only pertain to the period from 18,000 years B.P. to the present and hence postdate the last glacial maximum. During that time, a sequence of four different climate conditions has been recognized that apparently had discernible impact on geomorphic processes.

From the time of maximum glaciation about 18,000 years B.P. to about 15,000 years B.P., the climate of the central and upper portions of the Lower Mississippi Valley was considerably colder and wetter than at present (Royall, Delcourt, and Delcourt 1991). The alluvial valley landscape was dominated by outwash-carrying braided streams, and the valley train surface supported a boreal swamp forest dominated by spruce. Although this was a time of high water tables and increased effective precipitation, it was also a time of seasonal windiness with dune formation and loess deposition in the uplands. The latter environment was dominated by a cool-temperate deciduous forest. Increased discharges in tributary streams caused cyclical valley downcutting and are manifest by paleochannels considerably larger than those of the present streams (the Deweyville complex). Little is known about the southern part of the alluvial valley at this time, but the coastal zone (well south of its present position because of lowered sea level) possibly was drier and only a little cooler than present (Barry 1983).

The interval from about 15,000 to 9,500 years B.P. witnessed a climate warming, but conditions were still cooler and wetter than present. Due to retreat of the ice sheet to north of the Great Lakes area, glacial outwash deposition ceased in the alluvial valley area before the end of this interval. Boreal floodplain vegetation was replaced by a cool-temperate mixed deciduous hardwood forest. Groundwater levels declined, discharges in tributary streams decreased to present levels, and loess deposition ceased. It was during this time of climate amelioration that humans first appeared on the scene and became permanent inhabitants. The Mississippi River abruptly shifted from a braided to a meandering regime in the upper part of the alluvial valley, but this change was in response to the cessation of glacial meltwater discharge through the system rather than a climate change.

Perhaps the most famous and widely debated climate episode of the Holocene occurred after 9,500 years B.P. and reached its peak between about 7,500 and 4,000 years B.P. This is the Altithermal (also sometimes referred to as the Hypsithermal, thermal maximum, or climatic optimum) which was the time of maximum postglacial warmth and dryness. It is principally noted for an eastward shift in prairie vegetation from the Great Plains area into the upper Midwest. Because the Altithermal was time-transgressive, estimates of its onset and duration in the Lower Mississippi Valley area vary considerably depending on the location of the evidence. There is considerable disagreement over just how warm and dry this area became, but vegetation in at least the upper parts of the alluvial valley responded with an increase in grasses and herbaceous species (King and Allen 1977). It is generally believed that significant reductions in surface-water area and vegetation changes took place only in the already drier upland areas and that plant communities along streams responded little if at all. To what extent human populations responded to the Altithermal is much debated. Morse and Morse (1983) are perhaps the strongest advocates of a major influence on the locations of permanent settlements. In considering human responses, it should be kept in mind that it is likely that different river systems and even different parts of a given river system responded geomorphologically in different ways to the drier climate.

Within the alluvial valley, there is only one known instance where there may be direct geomorphic evidence for reduced stream discharges during the Altithermal. On the alluvial fan of the Current River (Western Lowlands), there are three abandoned courses out of a total of six that are distinctively smaller in several geomorphic parameters than either older or younger ones (Price et al. 1981). Archeological evidence suggests the smaller channels, indicative of reduced discharge, may be contemporaneous with the Altithermal. The Current River fan is discussed in Chapter 5 and illustrated in Figure 9.

A correlation has been sought between climate (as the causal mechanism) and Mississippi River meander belts in the Yazoo Basin that obviously were formed by a river carrying less discharge than at present (Saucier 1985). Development of these meander belts during the Altithermal when Mississippi River discharge may have been reduced is an attractive explanation and would be strong evidence that the Lower Mississippi Valley responded to this climate change. However, despite uncertainties in the ages of the meander belts, there appears to be no temporal correlation between the meander belts in question and the Altithermal episode. Further, it must be considered that because of the large longitudinal extent of the greater Mississippi River system, a climatically-induced reduction in flow in one part of the system (e.g., the Missouri River Basin) may have been compensated for by conditions elsewhere.

With the gradual dissipation of the relative warmth and dryness of the Altithermal, slightly cooler and wetter conditions returned to the Lower Mississippi Valley area no later than about 4,000 years B.P. and have persisted to the present. The temperatures and precipitation amounts of this period are probably comparable to what they were during the early Holocene (before the Altithermal). However, there is reason to believe that since about 4,500 years B.P., a meridional circulation regime has been more dominant than it was previously. Hence, the climate has been more variable within a wider range of extremes.

Sea Level Variations

Effects of sea level change

Through geologic time, many processes have affected the level of the sea, including changes in the volume of the ocean basins because of plate tectonics, seafloor spreading and underwater volcanism, variations in earth rotation, and others (Pirazzoli 1991). However, because of their high frequency and intensity, this study considers only Quaternary-period glacioeustatic changes, i.e., those attributable primarily to the cyclical accumulation of continental ice and, to a lesser extent, changes in ocean water temperature.

Since sea level is the "ultimate" base level, variations in that level have had the greatest geomorphic effect by causing changes in stream gradients. During glacial stages when sea levels fell, streams downcut and formed entrenched valleys. During interglacial stages when sea level rose, streams aggraded and filled the entrenchments. This response was strongest nearest the Gulf shoreline and on those streams directly tributary to the Gulf. Fisk (1944) overemphasized the effect of glacial-stage sea level lowering on the Mississippi alluvial valley by postulating that the entire valley was swept clean of sediments and deeply entrenched. Nevertheless, the lower portion of the valley did experience valley degradation and some scouring on several occasions during the Quaternary. Streams that directly discharged into the Gulf also experienced some entrenchment but again not to the extent that Fisk envisioned. He failed to consider the appreciable lengthening of stream courses that took place because of the retreat of the shoreline well out onto the continental shelf during at least the latest major sea level lowering. This channel lengthening process actually caused little change in stream gradients.

The most profound effects of glacioeustatic sea level variations were felt beneath the deltaic and chenier plains. Besides the shallow entrenchment of older coastal plain surfaces, each major sea level fall and shoreline regression allowed the previous subaqueous deposits to become emergent and exposed to long periods of subaerial weathering. This has produced a sequence of buried oxidized and desiccated soil horizons that have profound geotechnical significance. Each major sea level rise has resulted in the inundation and erosion of the preexisting land surface and the deposition of a typical onlapping sedimentary sequence. Toward the end of each major sea level rise episode, a new deltaic plain formed and its rate and pattern of development was directly influenced by the rate and magnitude of sea level fluctuations.

Reconstructing sea level history

No aspect of Gulf Coast Quaternary geology has probably been more widely investigated and debated than the history of sea level variations, especially the rise from the last (Late Wisconsin) major low stand to the present level. There are literally thousands of papers and reports exclusively or partly on this subject--at least nine curves representing different interpretations have been offered since the 1960s (Pirazzoli 1991). Various workers have used widely varying lines of evidence, including submerged shoreline features, oxygen isotope data, and archeological data from the continental shelf, but most of the curves are based on radiocarbon dates on buried or submerged marsh deposits and other organic materials. Uncertainties associated with these data include determining the exact stratigraphic relationship between the deposit and true level of the sea at the time of deposition, and correcting for subsidence, uplift, or tectonic deformation. A further complication is that most of the preserved stratigraphic evidence relates to the temporary high still stands and not to the lower positions that occurred in each fluctuation.

Sea level curves for the entire Quaternary, in other than a diagrammatic sense, have been attempted by few workers because of the general lack of information on earlier glacial cycles. Figure 4 portrays a generalized curve showing the timing of postulated cycles but does not provide a clear indication of the magnitude of sea level changes. This figure does indicate an important observation, however. The Illinoian stage, once thought to represent a single major glacial cycle, probably includes three separate low stands and two intervening relatively high high stands that were more than minor fluctuations.

It has traditionally been believed that at some point in each of the major interglacial stages, sea level attained an elevation higher than at present, and the older the stage, the higher the level. Reasonable estimates are not available from the central Gulf Coast area for the probable actual sea levels during interglacial stages of the Early Pleistocene (Figure 4). Evidence has been cited from the Atlantic Coastal Plain suggesting that sea level in late Pliocene times was 230 ft and possibly as much as 350 ft above present (Fairbridge 1960), but Colquhoun et al. (1991) believe other factors such as an increase in oceanic-basin volume or regional upwarping may be responsible for the observed effect. Nevertheless, it can be safely assumed that during Early Pleistocene glacial stages, sea level was on the order of several hundred feet lower than at present.

Late Quaternary sea level curve

Figure 5 is offered as a sea level curve for the last 150,000 years that is most compatible with evidence from both the Mississippi alluvial valley and the deltaic plain areas. It has been adapted from Revelle (1983) and Goodwin et al. (1991) but contains significant interpretation and judgment by this writer. In view of the lack of agreement among various geologists and oceanographers, undoubtedly this curve will provoke arguments among many. Readers should be aware that there are slight differences in time estimates and nomenclatures between Figures 4 and 5 because of the different sources and the different interpretations involved. For example, in Figure 4, the stage preceding the Sangamon is called the Late Middle Pleistocene, whereas in Figure 5, the name Illinoian is used. Differences of a few thousand years in age estimates for the stages is irrelevant to this report because there is no evidence from the Lower Mississippi Valley so precise as to favor one interpretation over the other.

In the original sources for Figure 5, the post-Sangamon stage is referred to as Wisconsinan rather than Wisconsin as is the case in Figure 4. To avoid confusion, the latter spelling is consistently used throughout this report.

It is even more apparent from Figure 5 than from Figure 4 that when data are relatively abundant as they are in the later stages, the curve becomes complex with frequent sea level rises and falls. Evidence suggests that both low stands and high stands were brief (at most, several thousand years long) and that major sea level rises consisted of frequent, brief still stands separated by periods of rapid sea level rise. For example, Penland et al. (1991a) have recently suggested that during the last major sea level rise (the Holocene transgression), there were several intervals in which sea level may have risen faster than 2 cm/year (5.7 ft/century). A possibility of this type focuses attention on the sharply contrasting situation of the last several thousand years when sea level has been unusually stationary from a geologic point-of-view.

There are several aspects of the post-Illinoian stage sea level curve that apparently bear a critical relationship to deposits and events in the Lower Mississippi Valley. First, at two different times in the Sangamon stage, sea level reached an elevation up to about 20 ft above the present level. As will be discussed later, there are extensive beaches and shoreline features in coastal Louisiana that are related to one or both of these high stands.

Second, there were two episodes between about 110,000 and 70,000 years B.P. (a 40,000-year-long period) when sea level fluctuations were of sufficient magnitude to affect the gradients (and regimes?) of streams near the coast, causing possible entrenchment and aggradation. Eustatic events of this period (heretofore largely unrecognized or disregarded) may hold a clue to the more appropriate chronostratigraphic positioning of landforms and deposits which, to date, has been in serious conflict with observable evidence.

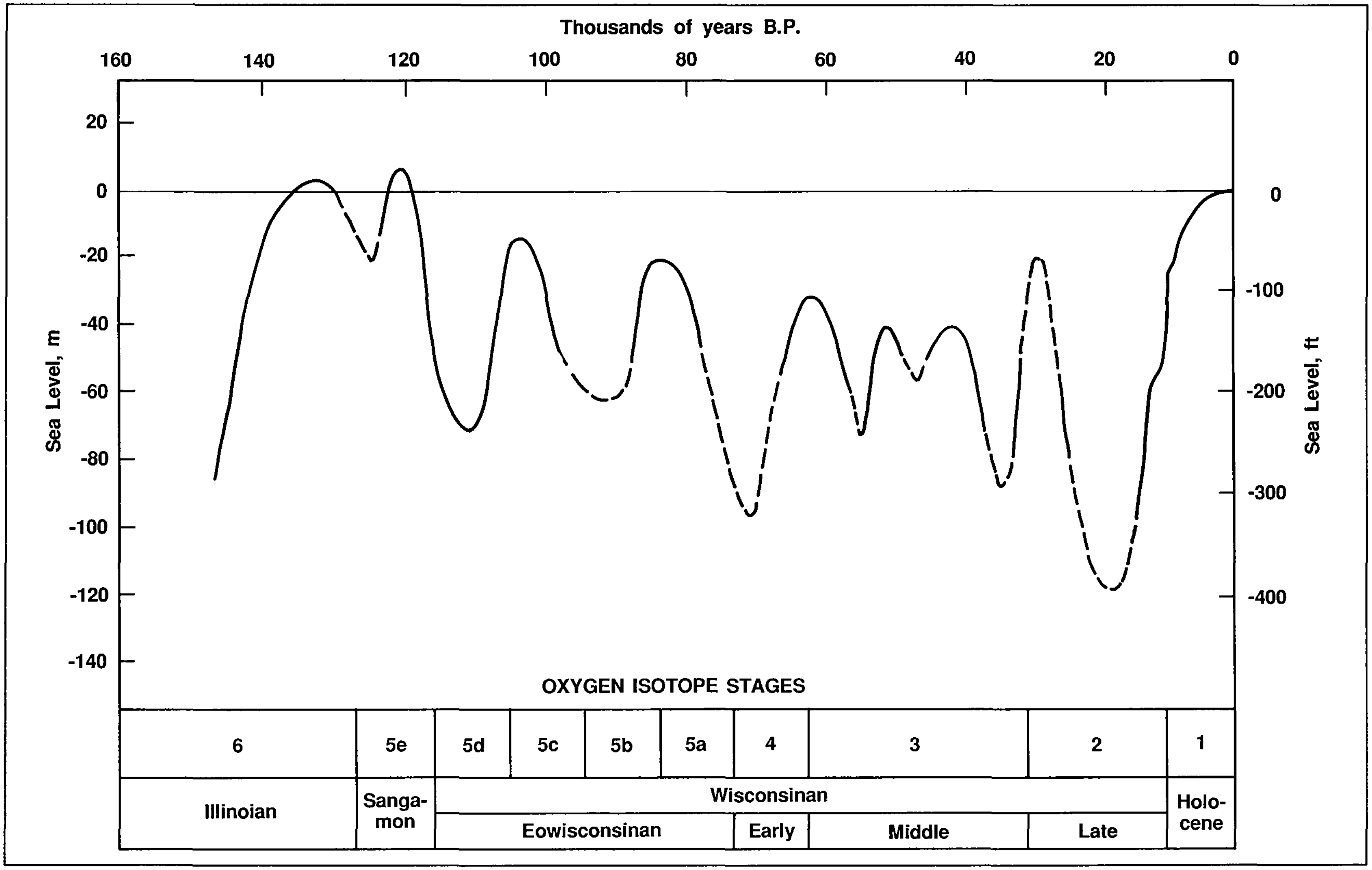

Figure 5. History of post-Illinoian sea level variations in the Gulf Coast area (dashed lines indicate interpretations by the author) (adapted from Revelle 1983, Goodwin et al. 1991, and other sources)

Third, there is a much debated possibility of a relatively high sea level stand sometime during the Middle Wisconsin stage (Figure 5). Considerable evidence from various locations in the southeastern United States, including numerous radiocarbon dates, has been cited (see Saucier 1977b) for a sea level close to or perhaps even higher than present (Figure 4). However, there are reasons to believe the dates may be anomalous, and they are not supported by oxygen isotope or other evidence from elsewhere in the world (Bloom 1983, Morner 1971). Notwithstanding, the Middle Wisconsin was a time of considerable glacial retreat in North America, and the Mississippi alluvial valley contains abundant evidence of a major episode of glacial outwash deposition and valley train formation (Saucier 1968). The writer now agrees that evidence against a major high stand at that time is compelling, but he still believes that a high stand that was between 50 and 100 ft below present took place sometime after 70,000 years B.P. and before 22,000 years B.P.

A fourth critical aspect of the curve is the magnitude of the Late Wisconsin stage low stand between about 22,000 and 18,000 years B.P. Influenced by the prevailing opinion of their time, Fisk and McFarlan, Jr. (1955) believed that sea level fell as low as 450 ft below present. More recent estimates tend to range from a maximum of 150 m (492 ft) to a more probable value of 120 to 130 m (394 to 426 ft). Both the magnitude of this low stand and its duration were important controls on the degree of weathering and entrenchment of the preexisting deposits beneath the deltaic plain.

The fifth critical aspect involves the timing of the last few tens of feet of sea level rise during the last several thousand years. More specifically, when did sea level reach its present level? There is a small but vocal minority that believe sea level actually exceeded its present level about 6,000 to 5,000 years B.P. (Fairbridge 1960). Most workers may argue over details, but they believe that sea level rose rapidly to within 10 ft of its present level about 4,000 years B.P. Since then, it has slowly but steadily risen to its present level. An interesting recent interpretation by Penland et al. (1991b) involves a rapid rise in sea level to the -20-ft level about 6,500 years B.P., a still stand until about 4,000 years B.P., and then a rapid rise to its present level by about 3,000 years B.P. Although this scenario seemingly is in conflict with certain archeological evidence, it has important implications as far as formation of the deltaic and chenier plains are concerned and is discussed in more detail later. Tanner (1991) and Stapor, Mathews, and Lindfors-Kearns (1991) have recently gone so far as to try to relate chenier formation in southwestern Louisiana to sea level fluctuations as small as 3 to 10 ft during the last 3,500 years.

Tectonics and Diapirism

The Gulf section of the Coastal Plain Province has been described as being one of the least complicated in the United States (Walker and Coleman 1987), but this does not mean that it has not been geologically active. To the contrary, throughout its history, it has been and continues to be affected by uplifting and downwarping over areas ranging from a few miles to hundreds of miles in extent. Essentially no part of the Lower Mississippi Valley has been unaffected by vertical movements of one type or another.

The first chapter in the tectonic history of the area begins with continental rifting that created the Gulf of Mexico Basin in Late Triassic or Early Jurassic times (Figure 3). Rifting continued until the basin achieved its present configuration by the Early Cretaceous (Buffler 1991). Initial deposits in the basin (of Jurassic age) consisted of huge masses of evaporites (primarily salt) that were laid down directly on Precambrian basement rocks in what is known as the Gulf Salt Basin. The basement rocks plunge southward into the basin from a depth of less than 3,000 ft south of the Ouachita Mountains in southern Arkansas (Figure 1) and lie at a depth estimated at about 50,000 ft beneath the present continental shelf off Louisiana.

Along the north side of the Gulf Basin, crustal downwarping during the Cretaceous period formed the Mississippi Embayment, a northward synclinal projection of the Coastal Plain that lies between the Southern Appalachians and the Ouachita Mountains (Figure 1). Beneath the Embayment, the folded and faulted belt (Ouachita orogeny) of Paleozoic rocks of the latter areas has been downwarped by several thousand feet.

At the end of the Upper Cretaceous, a marine transgression from the Gulf created water depths in the Embayment estimated at about 1,000 ft (Goldthwaite 1991). Shortly thereafter, the Embayment began filling with thick wedges of noncarbonate clastic sediments from the north and northwest and was largely filled during Paleocene and Eocene times. Sediments eventually spilled over a ledge or sill at the mouth of the Embayment and started filling the Gulf Basin. In post-Eocene times, the Embayment has experienced several cycles of transgression and regression of the sea and the deposition of alternating sequences of fluvial and marine clastics and carbonates, but the overall effect was a single major regressive sedimentary cycle continuing throughout the Tertiary into the Quaternary. As a consequence, the edge of the continental shelf prograded southward about 200 mi to its present location.

During the Upper Cretaceous, the lower parts of the sedimentary sequence of the Embayment were disturbed by volcanic intrusions that caused local uplifts, including the Monroe Uplift and the Jackson Dome. Those tectonic events are described in Chapter 4.

Development of the Gulf Coast geosyncline, a basin within the Gulf Basin, facilitated the accumulation and preservation of the large volumes of sediment that spilled out of the Embayment. The geosyncline is an actively subsiding area extending along the Gulf Coast from Texas into Louisiana and underlying the deltaic and chenier plains and the continental shelf. Deposition of the tens of thousands of feet of sediments during the Cenozoic era has caused the mobilization of the thick mass of Jurassic salt and the development of a complex band of coast-parallel, penecontemporaneous (growth) faults (Walker and

Coleman 1987) in south Louisiana. These large, deep-seated, down-to-the-coast, normal faults formed in the late Paleozoic and Mesozoic eras and are still active. Although fault movements occur, they are dominantly nonseismic, but not necessarily exclusively (Lopez 1991).

With great depth of burial, salt behaves as a viscous fluid through time. It becomes less dense than overlying sediments and rises and pierces through them to form salt domes in a process known as diapirism. These dome-, cylindrical-, or mushroom-shaped salt masses, thousands of feet tall and up to 6 mi in diameter, occur by the hundreds in two trends across the Lower Mississippi Valley (see Chapter 4). Only a few salt domes have reached the surface (e.g., the Five Islands in south Louisiana) and the others lie at depths of hundreds to thousands of feet beneath the surface (Murray 1961). Besides being the source of exploitable salt, sulphur, and gypsum, the domes have been responsible for trapping enormous quantities of hydrocarbons. Some shallow domes have affected near-surface Quaternary deposits and are of geomorphic significance as will be discussed later.

In addition to regional manifestations of tectonic controls, the entire lower part of the Lower Mississippi Valley has been affected by progressive, broad seaward tilting. The southern portions of each of the fluvial sequences, which are attributable to a major glacial cycle that form terrace complexes in the alluvial valley area, consequently plunge into the subsurface beneath the deltaic plain area (King 1965). The oldest sequence, which forms the highest terrace complex upvalley, lies at the greatest depth beneath the deltaic plain and has the steeped slope. The "hinge line" that separates the uplifted-terrace situation to the north and the buried deltaic sedimentary-sequence situation to the south approximates the northern edge of the deltaic plain and has migrated southward throughout the Quaternary. The locus of greatest deltaic sedimentation (depocenters) during the Quaternary has shifted from south-central Louisiana, where the oldest sedimentary sequences are thickest, to southeastern Louisiana, where maximum sedimentation is occurring in the present sequence.

For many years, geologists sought to identify the types, locations, and rates of tectonism with attention on the long spans of geologic time and little consideration of the extremely brief (in a relative sense) Quaternary period. It has long been known that the coast-parallel growth faults are currently active and that salt domes are continuing to rise vertically. Little or no attention was given to the possibility of continuing movement in other areas of doming or upwarping, especially those that originated in the Cretaceous. However, within the last two decades, Schumm, Watson, and Burnett (1982) and Schumm (1986) have focused attention on neotectonic (post-Miocene) controls and have argued that continued movement in areas like the Monroe Uplift have been sufficient to affect Quaternary terraces and even the Holocene floodplain. Further, Jurkowski, Ni, and Brown (1984) have presented data from precise levelling indicating detectable uplift in southwestern Mississippi during just the last several decades, and Cox (1994) believes that basin asymmetry resulting from preferred river migration along the Arkansas, Saline, and Ouachita rivers since the early Pleistocene may be a result of fault-block tilting. Although uplift or tilting may be occurring, it has been difficult to separate geomorphic responses to tectonism from those caused by normal fluvial processes: hence, the results of these studies are subject to varying interpretations.

Without question, the most tectonically active part of the Lower Mississippi Valley is the New Madrid Seismic Zone, an extensive area centered in the St. Francis Basin that has been the site of large-magnitude earthquakes, faulting, doming, and subsidence. Because of the special geological and engineering importance of this zone, it is the focus of discussions presented in Chapter 8.

Subsidence

In simple terms, subsidence involves the lowering of the elevation of a land area in relation to sea level. This change in relative elevation may be the result of true sea level rise, crustal sinking (e.g., geosynclinal action), consolidation of sediments, local consolidation of landforms or structures caused by their weight, tectonic activity, or any combination of these processes.

In the Mississippi alluvial valley (excluding the New Madrid Seismic Zone), consolidation of sediments is the only one of the five processes that is active. However, because the fine-grained alluvial deposits are not especially susceptible to compaction and are relatively thin, subsidence amounts and rates are both low. In the deltaic plain area, on the other hand, all five factors involved in subsidence are present, and subsidence is a major control in all aspects of this area's geomorphology (Kolb and Van Lopik 1958).

Estimates of present subsidence rates in coastal Louisiana primarily have been made by using tide gages (Ramsey and Penland 1989), by comparing present elevations of objects with their inferred original elevations, and by the radiocarbon dating of buried peat horizons (assumed to have originated at or slightly above sea level). Qualitative estimates and observations regarding regional variations have been made using archeological sites (Kniffen 1936, Saucier 1963).

Throughout the Quaternary, the entire deltaic plain area has been affected by crustal sinking and the consolidation of the thousands of feet of sediments in the Gulf Basin. During times of waning glaciation, the area has also been affected by sea level rise. Taking into consideration an estimated present rate of sea level rise of 0.32 ft/century, Kolb and Van Lopik (1958) estimated a basic subsidence rate of 0.78 ft/century for all of southeastern Louisiana for the present. Factoring in tectonics (growth faults), consolidation of the highly compressible Holocene deltaic deposits, and consolidation caused by weight of structures, local subsidence can be locally highly variable and as much as 2.0 ft/century. For comparison, Saucier (1963) estimated a subsidence of 0.39 ft/century for the Pontchartrain Basin area. In a regional context, maximum rates occur in the area of the modern active delta and decline to the north and west. In a local context, maximum rates occur in interdistributary basins and are noticeably lower on natural levees and over masses of point bar sediments.

Against this background of "normal" subsidence, human activities have caused acute cases where because of the short times involved, *rates* are not relevant, but actual amounts are. For example, in the New Orleans area, the lowering of water tables due to artificial drainage and the consequent desiccation of organic-rich interdistributary deposits has caused subsidence of as much as 6 to 7 ft within several decades over a large part of the city (Kolb and Saucier 1982). Similar amounts of subsidence have occurred in several large (several square mile) impoundments in various parts of the deltaic plain where drainage and reclamation for agriculture has been attempted. Also in the New Orleans area, groundwater offtake for industrial use from aquifers as deep at 700 ft has caused subsidence in parts of that city of about 1.7 ft (Kazmann and Heath 1968).

It is a fair assessment that not a single deltaic plain landform or depositional environment has been unaffected directly or indirectly by subsidence. Effects of subsidence are apparent in landform shape and distribution and in deltaic plain hydrology and vegetation patterns. However, from a geomorphic perspective, the greatest result has been a vertically stacked cyclical sedimentary sequence (Coleman 1988) that has operated at high frequencies. Delta complexes partially to completely overlap one another rather than being spatially separated but stratigraphically adjacent.

4 Regional Geologic Framework

Major Formations

Data sources and limitations

Table 2 is a generalized, composite stratigraphic column for the Lower Mississippi Valley area. It was assembled from many sources, including the geologic maps of the seven states involved, the synthesis by Cushing, Boswell, and Hosman (1964), and special publications such as Dockery (1981) and Howe (1961). Where more than one name has been used in different areas for a given formation or unit, preference was given to that name used in the area with the largest outcrop closest to the alluvial valley or the names used for outcropping portions of units versus those for deeply buried units.

Table 2 is intended to present brief lithologic descriptions and relative ages of the major formations or units that occur immediately adjacent to or at shallow depths (several hundred feet) beneath the alluvial valley and deltaic and chenier plains. Deposits such as those of Jurassic age that lie at great depths within the Mississippi Embayment or the Gulf Basin are of little or no engineering consequence and are not discussed herein. Table 2 does not attempt to indicate the stratigraphic relations between units, such as whether they are part of a continuous depositional sequence or separated by discontinuities. This type of information, where relevant, is presented in the following discussions.

Table 2 is directly linked with Plate 2 and should be used in conjunction with that illustration. The latter depicts the locations of the major units, usually at the group or series level.

The latest edition of the Geologic Map of the United States (King and Beikman 1974) was the principal source of data for Plate 2, but modifications were made using more recent state geologic maps (e.g., Snead and McCulloh 1984, Haley 1976), groundwater investigations by the U.S. Geological Survey, and a map of the geology of the suballuvial surface by Krinitzsky and Wire (1964). These various sources presented numerous conflicts and

TABLE 2. GENERALIZED, COMPOSITE STRATIGRAPHIC COLUMN FOR THE LOWER MISSISSIPPI VALLEY

ERA	SYSTEM	SERIES	GROUP	FORMATION OR UNIT	THICKNESS RANGE (FT)	LITHOLOGY
Cenozoic	Quaternary	Holocene		Alluvium	0-400	Unconsolidated clays, silts, and sands deposited in fluvial, deltaic, lacustrine, and marine environments.
		Pleistocene		Valley trains	50-300	Two sequences (Early & Late Wisconsin) of braided-stream deposits consisting of massive sands & gravels.
				Loess	0-75	Five sheets of tan to light brown, lightly calcareous, massive, eolian silts of Late to Middle Pleistocene age.
				Deweyville Complex	40-80	Fluvial terrace with thin fine-grained topstratum and thick coarse-grained substratum.
				Prairie Complex	60-200	Diverse time-transgressive depositional sequence representing fluvial to marine environments.
				Intermediate Complex	50-150	Fluvial terrace deposits of well-oxidized clays, silts, sands, and gravels. Includes Montgomery terrace.
	Tertiary	Miocene		Upland Complex	20-100	Well-dissected deposits of highly-oxidized, fluvial (braided-stream) sands and gravels. Includes Bentley and Williana terraces and Citronelle and Lafayette fms.
				Pascagoula	0-200	Gray fluvial to estuarine clays and sandy clays with layers of sand and sandstone. Occasionally fossiliferous.
				Hattiesburg	0-450	Hard, gray clays with claystone and thin, greenish sandstone and cemented sand layers. Includes Fleming formation of Louisiana.
		Oligocene		Catahoula	0-350	Gray to white, tuffaceous siltstones and sandstones with layers of loose, fine sands and thin clay layers.
			Vicksburg	Bucatunna	30-40	Dark brown, lignitic clays of marine or estuarine origin. Few thin siltstone layers.
				Byram	40-50	Highly fossiliferous marine clays and sandy marls with zones of nodular or lenticular limestone.
				Glendon Limestone	30-40	Alternating thick layers of hard, sandy limestones and clayey, sandy marls.
				Mint Springs	20	Fossiliferous, sandy and clayey marls with occasional phosphatic and lignitic pebbles.
				Forest Hill	0-150	Clayey, lignitic silts irregularly interbedded with fine, cross-bedded sands and thin layers of clayey lignite.
		Eocene	Jackson	Yazoo Clay	0-500	Dark gray, massive clays with widely scattered, irregular zones of silty clays. Occasionally fossiliferous.
				Moodys Branch	0-40	Fossiliferous, sandy and clayey marls with occasional layers and nodular zones.
			Claiborne	Cockfield	200-400	Lenticular, alternating, thin strata of gray to gray-brown clays and light gray silts or silty sands. Scattered lignite fragments and layers.
				Cook Mountain	130-160	Thick, brown, hard clays and reddish, clayey limonite alternating with thin beds of glauconitic sands.
				Sparta Sand	400-500	Massive, light gray, fine to medium sands interbedded with thin layers of brown, lignitic sandy clays. Incl. Memphis sand.
				Cane River	0-200	Green and brown, calcareous, glauconitic, and fossiliferous clays, marls, and sands. Includes Kosciusko fm.
				Carrizo Sand	0-190	Light gray to brownish-gray, fine to coarse, micaceous sands.
			Wilcox	Undiff.	100-920	Fine to medium, lignitic, sands and sandy clays and lignite. Massive sands, some coarse and graveliferous, in upper and basal portions.
		Paleocene	Midway	Porters Creek Clay	200-670	Massive, gray, fissile shales, clay shales, and clays with sandy clay beds.
				Clayton	0-60	Gray, calcareous, glauconitic, fossiliferous shales with scattered lenses of white limestone near base.

(Continued)

TABLE 2. (Concluded)						
ERA	SYSTEM	SERIES	GROUP	FORMATION OR UNIT	THICKNESS RANGE (FT)	LITHOLOGY
Mesozoic	Cretaceous	Gulfian		Arkodelphia Marl	0-200	Dark blue, fossiliferous marls interbedded with sandy limestones. Includes Owl Creek fm.
				Nacatoch Sand	100-500	Massive crossbedded, yellowish to gray, fine sands interbedded with sandy limestones and marls. Incl. Ripley fm.
Paleozoic	Pennsylvanian	Atokan		Atoka	1400+	Dark brown to black, thinly bedded, hard shales with layers of dark gray to brown, fine-grained sandstones.
		Morrowan		Bloyd Shale	0-150	Dark brown to black, carbonaceous clay shales interstratified with thin layers of limestones.
				Hale	50-275	Yellowish-brown to brown, massive and cross-bedded, medium grained sandstones.
	Mississippian	Chesterian		Pitkin Limestone	0-100	Includes Fayetteville Shale and Batesville Sandstone. Massive layers of gray, fossiliferous limestones.
				Ruddell Shale	10-400	Black, fissile, carbonaceous shales with concretions and thin layers of limestones.
	Ordovician	Osagean		Chattanooga Shale	0-85	Includes Clifty Limestone and Penters Chert. Fissile, black, carbonaceous shales with nodules and concretions.
		Cincinnatian		Maquoketa	10-60	Thinly laminated, silty shales with shaly lenses of limestones. Includes Thebes fm. (sandstone).
		Mohawkian		Kimmswick	50-150	Coarsely crystalline, white to light gray, medium bedded to massive limestones. Contains chert.
				Joachim	0-160	Yellowish-brown, argillaceous dolomites with interbedded limestones and shales in lower part.
				Dutchtown	170	Dark blue to black, fossiliferous, medium to thinly bedded limestones and dolomites with sandstones and shales.
				St. Peter	10-100	Massive, well-sorted, white, quartzose sandstones but locally orthoquartzites.
				Everton	400	Gray, sandy dolomites interbedded with fine-grained sandstones, limestones, and cherts.
		Canadian		Black Rock	100+	Fossiliferous, cherty dolomites and limestones with gray and brown silts.
				Smithville	150	Fossiliferous dolomites with small amounts of chert.
				Powell	150-175	Brown, medium to finely crystalline dolomites with thin beds of green shales and fine-grained sandstones.
				Cotter	200-450	Light gray to light brown, medium to finely crystalline, cherty dolomites with thin beds of green shale.
				Jefferson City	125-350	Brown, medium to finely crystalline dolomites and argillaceous dolomites.
				Roubidoux	100-250	Light gray to brown, bedded, finely crystalline dolomites with chert and some sandstones.
				Gasconade	300-700	Light brownish-gray, cherty dolomites.

inconsistencies in both interpretations and nomenclature that had to be resolved, sometimes arbitrarily by this writer.

It must be emphasized that Plate 2 is generalized and simplified. Numerous small outcrops, especially in the Paleozoic area of Arkansas and Missouri, were omitted because of scale limitations. Moreover, bands of alluvium and sometimes Quaternary terraces along small streams were not delineated for the same reason.

Identification and delineation of formations beneath the alluvium of the alluvial valley is generalized, not because of too much data, but rather too little data. Where formations outcrop, geologists have been able in most cases to make positive identifications based on lithology and paleontological studies. However, even though many borings and wells have been drilled into the formations beneath the Holocene and Late Pleistocene alluvium, there have been very few cases of attempts at positive identification of the formations involved. In most cases, drillers are only concerned with determining the depth where alluvial sands and gravels overlie highly consolidated clays, silts, or sands rather than the geologic age of the latter materials. Since most of the pre-Quaternary formations have similar lithologies, more than visual inspection of a few feet of core is needed for age determinations.

Descriptions

The cycles of sea level regressions and transgressions that led to the deposition of the thick masses of sediment in the Gulf Basin occurred in various depositional environments that ranged from terrestrial to estuarine to deltaic to shallow marine to deep marine. The more continental deposits include particle sizes that range from clays to gravels while the more marine deposits are dominated by limestones and dolomites. In general, the Quaternary and Tertiary formations consist of deposits that are unlithified--true rock caused by the induration of sediment by pressure or cementation is relatively rare. In contrast, the Paleozoic formations are dominated by sedimentary rocks, many of which are quite hard and erosion resistant.

Because of their importance, the Quaternary units listed in Table 2 are the subject of detailed discussions in Chapter 6 of this report, and brief descriptions are not presented herein. Rather, this chapter focuses on formations that range in age from late Tertiary (Miocene) to early Ordovician. In general, the youngest formations occur farthest south, and the oldest ones in the Ozark Plateau area to the north and northwest of the alluvial valley. They are described below, from youngest to oldest.

The Pascagoula and Hattiesburg formations of Miocene age consist of hard, gray, green or blue, freshwater to brackish-water clays and sandy clays. These materials are irregularly layered with fine sand, weak sandstone, and cemented sand which are resistant enough to erosion to form ledges and sills when exposed in creeks. Some thin lenses of small, black, chert gravel have been observed in the formations.

Pascagoula and Hattiesburg deposits are exposed at the surface in the uplands of southwestern Mississippi generally south of the latitude of Vicksburg. Within 10 mi or so of the alluvial valley, they are heavily veneered by graveliferous deposits of the Upland complex (early Quaternary) and are exposed only in deep valleys and along creeks. They have been observed in a similar context in the western portion of the Florida Parishes of Louisiana to a distance of about 20 mi south of the Louisiana-Mississippi state line (Plate 2). Beneath the alluvial valley, they comprise the suballuvial surface roughly south of a Vicksburg-Sicily Island line and extend to within about 10 mi north of Baton Rouge where they are thought to have been truncated by an east-west trending fault (Saucier 1969). In the lower Red River valley, deposits of the Fleming formation, a stratigraphic equivalent, comprise the suballuvial surface from the Alexandria area southward (Kolb 1949).

The Catahoula formation, originally believed to be of Miocene age and now assigned by some to the Oligocene, directly underlies the Hattiesburg formation and consists of gray to white, tuffaceous (volcanic ash-rich) siltstones and sandstones with some loose quartz sand. In outcrops, it is distinctive because of its light color, and it is hard enough to form prominent cuestas or ridges in central Louisiana and southwestern Mississippi.

West of the Mississippi alluvial valley, this formation forms the base of the hills at Sicily Island, outcrops in the banks of the Ouachita River, and constitutes the southwest-trending, prominent ridge from there to the vicinity of Alexandria (Chawner 1936). East of the alluvial valley, it occasionally outcrops between Natchez and Vicksburg and is perhaps best exposed in the banks of the Mississippi River at Grand Gulf, Mississippi, about 25 mi southwest of Vicksburg. Deposits of this formation have been identified beneath the suballuvial surface between the two areas but do not appear to have been sufficiently resistant to erosion to form a noticeable ridge or topographic high.

A relatively thin sequence of fossiliferous, estuarine to marine deposits that constitutes the Vicksburg Group of Oligocene age lies beneath the basically freshwater Catahoula formation. The uppermost unit is the Bucatunna formation which consists of hard, dark brown, carbonaceous, and lignitic clays with a few siltstone layers. Beneath this formation is the Byram formation which consists of highly fossiliferous, calcareous, glauconitic clays and sandy marls with zones of nodular or lenticular limestone. By far the most conspicuous unit of the Vicksburg Group is the Glendon Limestone. This formation is characterized by alternating thick (1- to 3-ft) layers of white to light gray, hard, sandy, abundantly fossiliferous limestones and clayey to sandy marls. The Mint Springs formation, which is of shallower-water origin, underlies the Glendon Limestone and is basically a massive bed of gray, highly fossiliferous, sandy marl with lignitic and phosphatic pebbles.

Some workers place the underlying Forest Hill formation in the Jackson Group of Eocene age because of its decidedly freshwater origin and its separation from the Mint Springs formation by an unconformity. However, following the interpretation of Dockery (1981), herein it is considered as the oldest unit of Oligocene age. This formation is composed of irregularly interbedded, discontinuous, thin layers of clayey, lignitic silts and fine, cross-bedded sands.

Five formations of Oligocene age form a narrow outcrop band that trends northeastward from the uplands northwest of Sicily Island across the alluvial valley area to the uplands in southwestern Mississippi (Plate 2). All formations are prominently exposed in the banks of the Mississippi River and the bluffs to the east from about 20 mi northeast of Vicksburg to several miles south of that city. Over that distance, it drops in elevation by about 100 ft but plunges steeply southward into the subsurface thereafter.

Two formations comprise the Jackson Group of upper Eocene age in Louisiana and Mississippi, but elsewhere the group is undifferentiated. The Yazoo Clay, by far the thickest of the two, consists of massive, dark gray to brown, occasionally lignitic, and fossiliferous clays with scattered thin lenses and zones of silty clays and fine sands. The unit becomes less fossiliferous, more lignitic, and significantly sandier toward the northeast, reflecting less marine conditions toward the upper part of the embayment at the time of deposition. Below the Yazoo Clay is the thinner but distinctively different Moodys Branch formation. This unit is also fossiliferous but is a glauconitic, sandy, and clayey marl with occasional, large, limestone nodules.

Jackson Group deposits occur in two distinctly different outcrop areas in the Lower Mississippi Valley. The first occurs north of and in an outcrop band similar to but wider than the Oligocene-age deposits. They are exposed over a distance of about 10 mi in the bluffs west of the Ouachita River in central Louisiana and over about twice that distance in the bluffs east of the Yazoo Basin in western Mississippi. A considerably wider subcrop band is shown beneath the alluvial valley, based on occasional borings that have encountered the relatively east-to-identify dark-colored, hard clays (Kolb et al. 1968).

The second Jackson Group outcrop area is a broad (50- to 75-mi-wide), mostly continuous zone that extends from the uplands of southeastern Arkansas northeastward for more than 250 mi into the uplands of western Tennessee and extreme southwestern Kentucky. In the latter areas, the deposits occur only in the bluffs along the Mississippi River (Hardeman 1966). Because of their decidedly sandy nature, there is disagreement as to whether they are assignable to the Jackson Group or the lithologically similar Claiborne Group. Beneath the alluvial valley, the distribution shown is that of King and Beikman (1974); however, Krinitzsky and Wire (1964) do not show Jackson Group deposits constituting the suballuvial surface any farther than about 20 mi north of Memphis.

The Cockfield formation, the upper unit of the Claiborne Group, consists of highly lenticular, alternating, thin beds or strata of gray to gray-brown clays and light gray silts or silty sands. Lignite fragments are scattered throughout the deposits, and layers of impure lignite up to several feet in thickness occur. Some massive sand zones occur near the base of the formation. The underlying Cook Mountain formation includes thick beds of chocolate brown, fossiliferous, hard clays and reddish, clayey limonite alternating with thin beds of fine, glauconitic sands. Next in turn is the Sparta Sand (and its probable equivalent, the Memphis sand), the thickest unit of the Claiborne Group. This formation consists of light gray, massive, fine- to medium-grained sands interbedded with finely laminated, lignitic, brown sandy clays. Dark brown lignitic clays and thin lignite beds occur in the upper part of the formation.

Below the Sparta Sand lies the Cane River formation which contains green and brown, calcareous, glauconitic, and fossiliferous clays, marls, and sands. The Kosciusko formation, Winona Sand, and Zilpha Clay are probable stratigraphic equivalents. At the base of the Claiborne Group lies the Carrizo Sand which consists of light gray to brownish-gray, fine- to coarse-grained, micaceous sands.

The five formations of the Claiborne Group collectively have the largest outcrop (and subcrop) area of any mapped unit in the Lower Mississippi Valley area, occurring in northeastern Louisiana, western Mississippi, much of eastern Arkansas, the "Bootheel" area of Missouri, and portions of western Tennessee and Kentucky (Plate 2). In general, deposits of this group surround the outcrop area of the Jackson Group and extend over a north-south distance of nearly 400 miles. Because the Claiborne Group contains some of the more important aquifers in the region (e.g., the Sparta Sand), and because significant hydraulic interconnections are known to occur between the Tertiary aquifers and the alluvial aquifer, there have been several attempts to differentiate the Claiborne Group deposits at the suballuvial surface (e.g., Krinitzsky and Wire 1964; Hosman, Long, and Lambert 1968; Dalsin 1978). Results of these investigations cannot be summarized and presented in this report because of the significant differences and conflicts in interpretation, the resolution of which is beyond the scope of this study. It appears safe to say only that the younger formations, such as the Cockfield, occur over at least the western half of the outcrop area, whereas the other formations occur toward the east, such as in the eastern and northeastern portions of the Yazoo Basin area.

Deposits of the Wilcox Group (designated a formation in some areas), the thickest Tertiary unit in the Mississippi Embayment, consist of fine- to medium-grained sands and sandy clays, in part lignitic, and carbonaceous clays and lignite. Massive sands, occasionally coarse and graveliferous, occur in the upper and basal portions of the unit. Although there is considerable disagreement over the extent of Wilcox Group deposits in southeastern Missouri and western Kentucky and Tennessee, a narrow band flanking the Ozark Escarpment from just south of Little Rock to the vicinity of Cairo, as shown in Plate 2, is agreed to by most workers. Accordingly, most Wilcox Group deposits lie beneath the alluvial valley and have limited upland outcrops.

The Porters Creek Clay of the Midway Group consists of massive, dark gray to black, fissile shales, clay shales, and hard clays with some interbedded or laminated fine sand in the upper part of the unit. The underlying thin Clayton formation, the oldest Tertiary unit in the embayment area, consists of gray, calcareous, glauconitic, fossiliferous shales or hard clays with scattered lenses of white limestone near the base of the unit. Outcrops of Midway Group deposits are quite limited, occurring in narrow bands south and northeast of Little Rock and across the northern portion of Crowley's Ridge and into the Commerce Hills. Being the oldest deposits within the Coastal Plain, they unconformably overlie the Paleozoic rocks in a narrow band in the suballuvial surface of the Western Lowlands just east of the Ozark Escarpment. Deposits of Cretaceous age are poorly represented in proximity to and beneath the alluvial valley area, and only two formations have been identified. The younger, the Arkadelphia Marl (and the equivalent Owl Creek formation), consists of dark blue, green, yellow, and brown, fossiliferous marls interbedded with sandy clays and limestones. The older, the Nacatoch Sand (and the equivalent Ripley formation), is composed of massive, yellowish to gray, cross-bedded, fine sands interbedded with sandy limestones, marls, and carbonaceous clays. The principal outcrop area for Cretaceous deposits occurs north of the Midway Group in the northern part of Crowley's Ridge and the Commerce Hills in southeastern Missouri and continues across the Mississippi River into extreme southern Illinois. Otherwise, a small area of undifferentiated Cretaceous deposits overlies Paleozoic rocks just west of the Ozark Escarpment in the vicinity of Newport, Arkansas.

The Paleozoic rocks that constitute the uplands immediately west of the Ozark Escarpment between the vicinity of Little Rock and the Mississippi River at Cape Girardeau range in age from Pennsylvanian to Ordovician and are divisible into 19 formations. They also outcrop along the northern edge of Crowley's Ridge and the Commerce Hills and form the suballuvial surface beneath the Advance and Drum Lowlands. East of the outcrop band, these "bedrock units" plunge steeply into the subsurface of the Mississippi Embayment at rates of 40 to 80 ft/mi. In the vicinity of Greenville, Mississippi, they lie at a depth of about 5,000 ft.

Because of their similar lithologies and limited relevance to this synthesis, each of the Paleozoic formations is not discussed separately herein. The reader is referred to Table 2 for specific descriptions. In overview, it can be stated that the Pennsylvanian and Mississippian system units consist primarily of shales and fine-grained sandstones interlayered with lesser amounts of limestones. In contrast, the thick sequence of Ordovician units consists dominantly of dolomites with subordinate amounts of limestones, shales, and sandstones in the form of thin beds. Most of the formations contain abundant nodular chert and are capped with deep, cherty, residual soils.

Readers should be aware that there are outcrops of limestones of Devonian and Silurian age in Thebes Gap and in the uplands of extreme southwestern Illinois that are not included in Table 2 or Plate 2 because of their extremely limited extent. Similarly, there are igneous intrusions in the vicinity of Little Rock that are not shown for similar reasons.

Major Structural Features

The Mississippi Embayment, a relatively symmetrical syncline, and the northern flank of the Gulf Basin into which the embayment merges are the primary structural features of the Lower Mississippi Valley area (Murray 1961). In the former, Paleozoic rocks are downwarped by as much as 10,000 ft roughly midway between the Ouachita Mountains to the west and the Southern Appalachians to the east.

Because of the influence of several secondary structural features, the axis of the embayment follows a sinuous, north-south trend roughly along the eastern side of the Mississippi alluvial valley (Figure 6). The embayment widens noticeably in eastern Arkansas into the Desha Basin of southeastern Arkansas, which is actually the eastern portion of the larger Arkoma Basin. In western Mississippi, the embayment narrows and its axis swings to the southeast as a consequence of the Monroe Uplift to the west and the Jackson Dome to the east. South of these uplifts, the embayment broadens into the east-west trending Gulf Basin of south Louisiana, and the Paleozoic rocks plunge to depths as great as 20,000 ft in the vicinity of Vicksburg and 30,000 ft in the vicinity of Baton Rouge.

It is certainly not coincidence that the broad "S" shape of the Mississippi alluvial valley from southern Illinois to the Gulf of Mexico mimics the axis of the Mississippi Embayment. However, the alluvial valley is offset to the west of the embayment axis rather than directly overlying it; hence, the valley does not overlie the deeper part of the syncline where presumably subsidence and downwarping have been greatest. It is logical to assume that if the structure of the basement influenced the configuration of the alluvial valley, the effects of secondary structural features such as the Monroe Uplift would be most strongly manifest. While there appear to be general relationships between the features and the valley, they are subtle and equivocal. Much stronger relationships between structure and physiography are evident in terms of the smaller features such as salt domes, faults, and igneous intrusions.

The most prominent secondary structural feature is the Monroe Uplift of southeastern Arkansas, northeastern Louisiana, and west-central Mississippi (Figure 6). This feature is a broad, relatively flat-topped dome that has been active during the Mesozoic and Cenozoic eras. It may be an isostatic adjustment of the basement in response to the formation of adjacent basins. Although the uplift may be experiencing neotectonic movement that has warped some Quaternary terraces (see Chapter 3), the effects of the feature are most evident in the distribution of the Jackson and Claiborne Group deposits (Plate 2). In the subsurface, older Tertiary formations markedly thin onto the uplift and exhibit unconformities that reflect intermittent movement over a long period of geologic time.

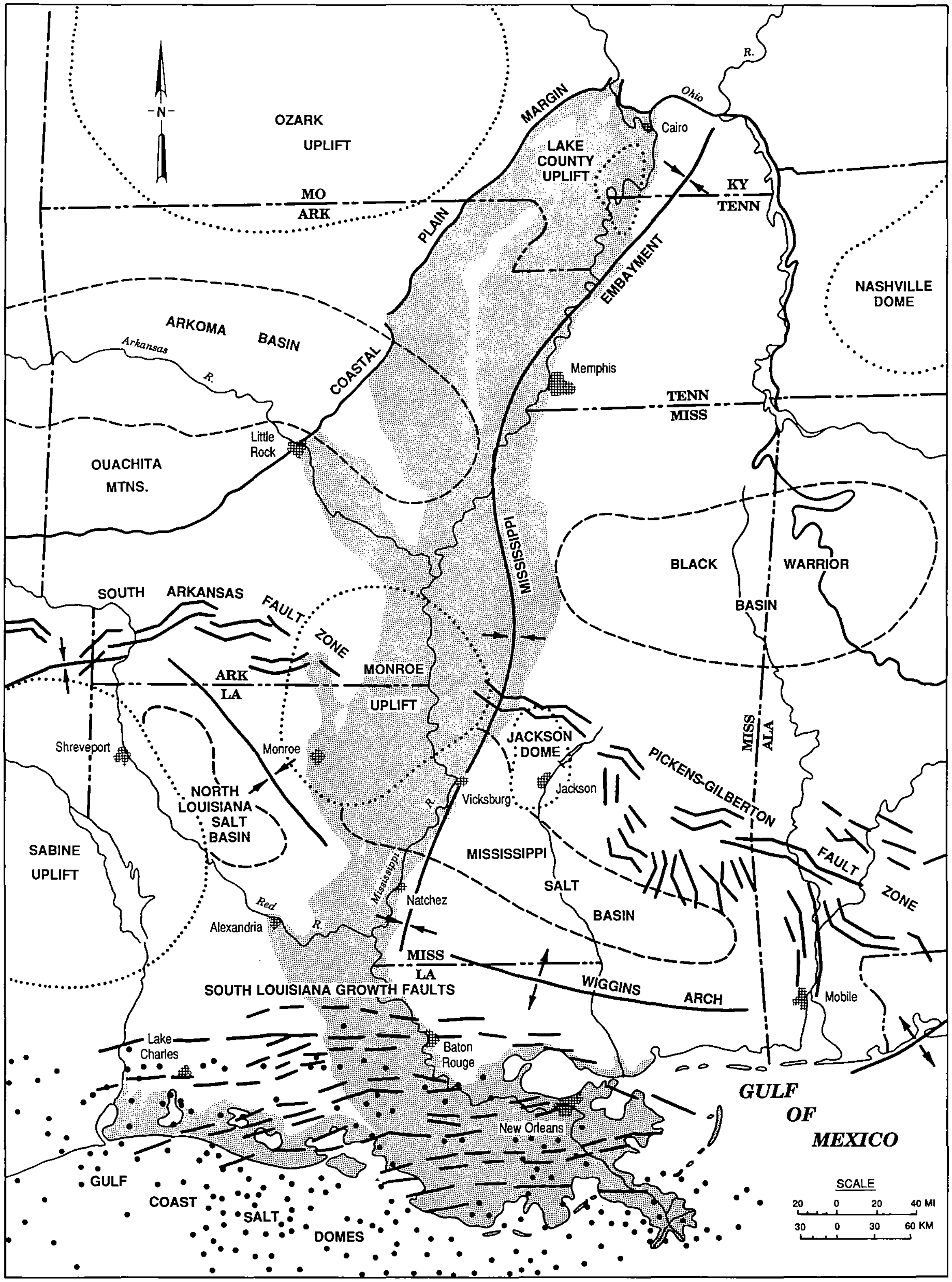

Figure 6. Major structural features of the Lower Mississippi Valley (modified from Autin et al. 1991)

Just east of the Monroe Uplift in central Mississippi is the Jackson Dome. This structural feature is much smaller (25 mi in diameter) than the Monroe Uplift but has considerably greater structural magnitude. The geologic histories of the two uplifts are generally similar (e.g., movements in the Mesozoic and Cenozoic eras), but the Jackson Dome is due at least partially to deep-seated igneous intrusion. Outcrop patterns of Tertiary deposits ranging from the Eocene to the Miocene clearly indicate the effects of the doming, but the movement does not appear to have significantly affected surface elevations or influenced the course of the Pearl River which developed across the feature in late Tertiary times.

A secondary structural feature that has strongly and extensively affected late Tertiary and Quaternary formations is the Wiggins Arch of extreme southern Mississippi and the Florida Parishes area of Louisiana. This feature has also been referred to as the Wiggins Uplift (Murray 1961) and the Southern Mississippi Uplift (Fisk 1944). It extends roughly east-west from the eastern side of the Mississippi alluvial valley south of Natchez to the vicinity of Mobile, Alabama. Originally believed to be a simple, arcuate anticline, the arch is now known to be an irregular, complex composite of several structural anomalies with evidence of igneous intrusion and faulting along its flanks.

Deep stratigraphy in the region indicates the Wiggins Arch is a comparatively young structure with little or no movement occurring before the early Eocene. However, considerable uplift occurred into the Miocene and has strongly influenced the surface distribution of both late Tertiary and Quaternary deposits. The crest of the arch is a topographically prominent broad, hilly ridge from which both Tertiary and Quaternary deposits slope gently to the north and quite steeply to the south into the zone of growth faults. Fisk (1944) demonstrated how the coastwise terraces of the Florida Parishes (the Prairie, Intermediate, and Upland complexes, Plate 2) have a steeper southward dip than those of similar age and relative location in southwestern Louisiana. The Wiggins Arch is also believed to be responsible for the narrowing of the alluvial valley south of Natchez and anomalous courses of rivers in extreme southern Mississippi east of the Pearl River.

Burnett and Schumm (1983) and Jurkowski, Ni, and Brown (1984) report evidence from precise leveling surveys (unadjusted) that the Wiggins Arch is currently experiencing uplift of several millimeters per year. This is thought to be a possible continuing crustal response to the high sedimentary loading occurring in the Mississippi Delta region.

In terms of rates of uplift and recentness of activity, the Lake County Uplift of extreme southeastern Missouri and western Kentucky and Tennessee is in a class by itself. This relatively small feature, which includes the Tiptonville Dome and Ridgely Ridge, is unique and of entirely different geologic origin than the other uplifts: it is associated with active faulting and modern seismicity in the New Madrid Earthquake Zone (Russ 1982) and likely is no older than the early Holocene. This feature is discussed in detail in Chapter 8.

One significant effect of the development of the uplifts mentioned above was the creation of two shallow basins. The North Louisiana Salt Basin (and the adjacent, northwest-southeast trending North Louisiana Syncline) formed between the Monroe Uplift and the Sabine Uplift of northeast Texas and northwest Louisiana (Figure 6). To the east, the Mississippi Salt Basin formed between the Jackson Dome and Monroe Uplift to the north and the Wiggins Arch to the south. During Jurassic times, these two shallow depressions experienced the deposition of salt that reached a thickness of 3,500 ft (Ingram 1991). As a consequence of loading from thousands of feet of overlying sediment, the salt began migrating in the late Mesozoic and has produced rather uniformly spaced, parallel, linear ridges (salt anticlines) and lines of salt "pillows" that roughly parallel the trends of the basins. In a few cases, the salt has moved up to within 1,000 ft of the surface and has formed surface uplifts (e.g., the Vacherie Dome in Winn Parish, Louisiana, and the Glass Dome in Warren County, Mississippi). All total, there are about 80 discrete and named domes in the two basins.

Salt domes principally occur in the Lower Mississippi Valley area in the Gulf Basin beneath and offshore from coastal Louisiana. There they are simply referred to collectively as Gulf Coast Salt Domes (Figure 6). The onshore domes are actually the northern part of a broad zone that extend over much of the northern and western Gulf of Mexico and that is as much as 200 mi wide (north-south), extending from about the latitude of Baton Rouge southward to the edge of the Sigsbee Escarpment. In this zone, salt (the Louann formation of middle Jurassic age) was originally deposited to a thickness of about 13,000 ft. Onshore and beneath the continental shelf, the domes resulting from subsequent sedimentary loading are mostly isolated diapiric structures. In the deep offshore area, they are mostly irregular, tongue-like masses. Beneath coastal Louisiana, the "mother bed" of the domes lies at a depth of about 40,000 ft. Salt is still rising (albeit episodically) as spires in some salt domes at rates that classify as rapid in geological time frames. The estimated actual rates are probably on the order of 0.01 in./year or less.

The Gulf Coast Salt Dome area is essentially coincident with the area of South Louisiana Growth Faults. This area includes at least ten separate and named fault zones, each of which consists of a series of *en echelon* normal faults. The zones trend roughly east-west, are about 8 to 20 mi apart, and can be traced for distances of 100 mi or more (Murray 1961). Displacements along the faults are principally but not exclusively down-to-the-basin, took place contemporaneously with deposition, and generally increase with depth. Dips of the faults average 50 deg or more, and displacements in the deeper sedimentary sequences of the Gulf Basin are on the order of several thousand feet. In general, fault movements began in the Paleocene or Eocene in the northernmost zones but did not start until the Miocene in the southernmost zones.

Several lines of evidence suggest that most of the fault zones have had some noticeable but geomorphologically unimportant effect on nearsurface deposits of Pleistocene age. Only the Baton Rouge Fault Zone has had any major geomorphic impact and is known to be currently active. Extending eastward from the city of Baton Rouge along the northern side of the Pontchartrain Basin to the mouth of the Pearl River, that fault zone contains well-documented evidence of surface escarpments and offset stream channels. The fault zone is discussed in more detail in Chapter 8.

Excluding those in the New Madrid Seismic Zone (also discussed in Chapter 8), the only other major fault zones in the Lower Mississippi Valley area are South Arkansas Fault Zone and the Pickens-Gilberton Fault Zone (Figure 6). These two zones are closely related and part of a long, nearly continuous trend of 5- to 8-mi-wide *en echelon* grabens extending from northeast Texas to southern Alabama. Fault displacements are sometimes visible at the surface, increase with depth, and often exceed 1,000 ft at depths of 5,000 to 10,000 ft. Surface effects of the faulting include topographic lows and stream anomalies, but this writer is not aware of any documented cases of Holocene effects or modern fault movement.

Formation of the Mississippi Valley

Development of the Mississippi River system is a consideration entirely separate from the development of the Mississippi alluvial valley, and the two should not be confused. On a continental scale, geological and physiographic conditions allowing southward drainage between the Rocky Mountains on the west and the Appalachians on the east to enter the newly formed Gulf Basin had developed by either the Jurassic (Mann and Thomas 1968) or the Paleocene (Baker, Killgore, and Kasul 1991), depending on the interpretation. By that time, the connection between the Appalachians and the Ouachita Mountains had been broken by the continental rifting and downwarping that created the Mississippi Embayment. Drainage from the north was the principal contributor of sediment to the northern end of the embayment. During the several transgression-regression cycles of the Tertiary, there was progressive filling of the embayment and a net gulfward movement of the shoreline. Throughout that time the ancestral Mississippi system maintained a course probably along the general trend of the embayment axis while the lower portions of the ancestral valley were alternately alluviated and drowned. The central part of the embayment would have been a topographic low that existed between a broad alluvial apron of fluvial clastic sediments eroded from the Ouachita Mountains to the west and a similar apron radiating out from the Nashville Dome (Figure 6) and the Southern Appalachians to the east.

By the end of the Tertiary, most of the embayment had been filled with sediment, and the ancestral Mississippi was probably established in a well-defined but rather narrow and shallow valley. That marks the first manifestation of a Mississippi alluvial valley. However, the river system was much smaller than at present (much lower discharge) since major tributaries to the Mississippi did not drain directly into the valley. Instead, it is believed that the upper and middle Missouri River system drained northward into Hudson Bay (Baker, Killgore, and Kasul 1991), some of the Ohio River system drained into the Atlantic Ocean via the St. Lawrence River (Fisk 1944), and much of the drainage from the southern Great Plains may have directly entered the western Gulf of Mexico.

The precise location of the late Tertiary valley south of Cairo cannot be determined, but unquestionably it was between the Ozark Escarpment and Crowley's Ridge to as far south as Helena. Drainage from the east, including the ancestral Ohio and Tennessee rivers, probably occupied a small valley somewhere in the area of the present St. Francis Basin. The two valleys joined somewhere south of Helena. Since these two valleys north of Helena are asymmetrically situated within the embayment, it is hypothesized that their positions were influenced by structural control, including faulting and down-warping in the New Madrid Seismic Zone (VanArsdale et al. 1992).

During the onset of the first Pleistocene glaciation in the Early Pleistocene stage (Figure 4), falling sea level caused the entrenchment and deepening of the lower portion of the ancestral Mississippi valley. As Fisk (1944) pointed out, this downcutting worked headward resulting in steeper stream gradients and formed probably the first manifestations of a hilly topography in this portion of the Coastal Plain. Simultaneously, the advancing ice sheet caused a complete disruption of northward flowing streams in the upper Midwest (i.e., the eastern Great Plains, Great Lakes, and south-central Canada) and diversion of the flow southward into the Mississippi River system.

Waning of the first glaciation about 2 million years ago must have produced significantly increased discharges in the Lower Mississippi Valley area. Glacial meltwater was augmented with runoff from a greatly enlarged drainage network. Waning glaciation is believed to have been accompanied by major valley widening and deepening and probably marks the *effective* beginning of the Mississippi alluvial valley. However, the valley was smaller than the one that exists today. By the first major interglacial stage, the valley for the first time probably was characterized by an alluvial plain underlain by a thick coarse-grained substratum of glacial outwash deposits. The elevation of the Early Pleistocene alluvial plain was 100 ft or more above the level of the present (Holocene) floodplain.

It should be kept in mind that the Mississippi and Ohio rivers occupied separate valleys to at least as far south as Helena throughout the majority of the Quaternary period. Diversion of the Mississippi River into the upper end of the St. Francis Basin to join the Ohio River did not take place until very late in the Wisconsin stage.

5 Landforms and Geomorphic Processes

Erosional Landscapes

Surface weathering and erosion

In the Lower Mississippi Valley, essentially only the Holocene alluvial valley and the deltaic and chenier plains are primarily depositional landscapes; in all other areas, the deposits and formations lie above the base levels of erosion and hence are being degraded. Rates, patterns, and processes of degradation vary greatly from area to area depending on a variety of factors.

Processes and patterns. As defined by geomorphologists, surface (subaerial) weathering includes the decomposition and disintegration of rocks or parent material by chemical, physical, and biological processes. If these processes continue for an extended period of time, soil is formed. To engineers, soil is any accumulation of unconsolidated materials. However, according to earth scientists (and as used herein), soil is a deposit with distinct horizons caused by weathering processes that can be described in terms of a profile. The properties of the soil are measurably different from the rock that it is derived from.

Chemical weathering includes such processes as oxidation and reduction, solution, hydrolysis, and ion exchange. Physical weathering includes thermal expansion, frost effects, hydration, and other processes. In a humid, temperate climate such as has characterized the Lower Mississippi Valley, even during glacial stages, chemical weathering processes are dominant. Percolating water is the dominant factor in soil formation, and rates of weathering obviously are related to such factors as climate, topography, the mineral composition of the parent material, and time.

Erosion involves the movement of soil or decomposed parent material by a geomorphic agent such as water, wind, waves, and gravity. Water is overwhelmingly the dominant agent in the Lower Mississippi Valley area and involves waves and currents in the coastal area, and overland flow (wash) and the actions of rivers (fluvial erosion) in the interior. Raindrop erosion, often an important process in some areas, is not important in this region because of the prevailing heavy vegetative cover on most landscapes. Mass movements, including slumps and landslides, are not areally widespread but are important erosional processes along bluffs and escarpments.

In general, rates of erosion in the region are high because of the large extent of unconsolidated parent material and the nature of the climate. As is typical of most humid areas with such materials, hill slopes tend to have rounded profiles and are concave in their lower portions. Steep slopes or vertical faces are limited to areas of rock outcrops, or where undercutting of slopes by lateral stream migration or entrenchment has taken place. Drainage networks are well developed and integrated (although not necessarily efficient), and enclosed depressions are almost nonexistent.

Weathering and fluvial erosion have produced four primary landscape types in the Lower Mississippi Valley area. These are definable primarily on the basis of parent material, geologic structure, and the length of time that the area has been subjected to weathering and erosion.

The most intensive weathering and erosion have taken place in the bedrock of the Paleozoic uplands of Arkansas and Missouri. The determining factors have been the long period of time (250 million to more than 400 million years) that the area has been uplifted and continuously subjected to weathering and erosion, and the susceptibility of the limestones and dolomites to solution processes.

The uplands are characterized by maturely developed and deeply incised drainage networks with thick, cherty, red clay residual soils capping interfluve areas. Local relief is often more than 100 ft. While slopes in interfluve areas are gentle, they are quite steep with vertical faces and steep bluffs tens to hundreds of feet high common along streams. Lower slopes of hills are composed of thick masses of stony, clayey slope wash (colluvium), and stream valleys have narrow floodplains overlying alluvial fills consisting of sands, gravels, and cobbles. In the area of Ordovician-age rocks, karst features such as caves, rock shelters, springs, and sink holes are common occurrences.

Coastal Plain uplands are the second erosional landscape type and are composed of Tertiary-age deposits of clays, silts, and sands. The uplands occur primarily in northeastern Louisiana, southeastern Arkansas, and east of the loess belt in western Mississippi, and exhibit a moderately dissected landscape of rounded hills with relief measurable in terms of tens of feet. Drainage is efficient, networks are typically dendritic, and individual stream valleys have relatively flat floodplains underlain by alluvial fills. In comparison to the Paleozoic uplands, colluvial slopes are relatively less extensive, and alluvial fills along streams are much more extensive. Soils are moderately well developed in interfluve areas and closely resemble the parent materials.

Pleistocene terraces, such as the Grand Prairie region in Arkansas (Plate 1), the several terrace complexes of southern Louisiana (Plate 2), and the Wisconsin-age valley trains (e.g., Macon Ridge), constitute the third landscape type. In these areas, erosional landscape patterns are more strongly influenced by the age of the deposit rather than the nature of the parent material. Older terraces are more maturely and uniformly dissected, relief is higher, and relatively flat interfluve areas are rare. Drainage on younger terraces is rather poorly developed, interfluve areas are flat to undulating with very shallow slopes, soil profiles are relatively weakly developed, and stream valleys are broad and shallow. Dissection is generally well developed only along terrace margins where gullying is extensive and streams are actively eroding headward and extending into inland areas because of sharp changes in elevation.

The degree of erosion and dissection of the older terraces, such as the Upland complex, is high enough that there is seldom any reflection of primary depositional features in the drainage patterns. However, with decreasing age and immature drainage system development, the influence of such features as abandoned channels and point bar accretion topography becomes much more evident. For example, on the Wisconsin-age valley train surfaces such as Macon Ridge and the Western Lowlands, the drainage networks are strongly dominated by the pattern of relict braided-stream channels that carried the glacial outwash. Because of low gradients, surface runoff is quite inefficient and is via small, underfit streams that are within the confines of the larger relict features.

The fourth erosional landscape type includes the thick loess deposits that cap the uplands on Crowley's Ridge and east of the Mississippi alluvial valley. In those areas, and especially near the bluffs, the degree of erosion and dissection has been intense. The high susceptibility of the unconsolidated silt deposits to erosion by overland flow (slopewash) has led to the formation of "badlands topography" characterized by steep gullies separated by irregular, knife-like ridges. Stream gradients are sufficiently steep, and the silt so easily moved by running water that most of the eroded loess has been transported from the uplands into the alluvial valley.

Historic period changes. However intensive erosion has been in the Lower Mississippi Valley area, it increased by orders of magnitude during the historic period. Beginning in the early part of the 19th century, land clearing for agriculture and timber clearcutting caused widespread erosion and gullying in viturally all upland areas. It is reasonable to conclude that there has been as much erosion in the last 150 years as there was in the preceeding several thousand years if not several tens of thousands of years. Fortunately, increased forest cover and improved agricultural practices, as well as sediment control dams by the Soil Conservation Service, have led to a decline in erosion rates during the last several decades.

As would be expected, erosion has been most extreme in the loess uplands. Barnhardt (1988) has recently documented a 20-fold increase in rates of sedimentation in western Tennessee gullies directly related to loess erosion, and

Neitzel (1983) describes an archeological site on a terrace along a small creek near Natchez that was veneered with as much as 6 ft of silt since 1700 A.D. The term postsettlement alluvium generally is used to describe these materials. Silt deposition was a direct result of loess erosion caused by land clearing in the creek watershed. Even in nonloess Coastal Plain upland settings, erosion has been intense. Schumm, Harvey, and Watson (1984) estimated soil erosion rates in northern Mississippi increased from 0.008-0.02 mm/year prior to cultivation to as much as 8.2 mm/year during the peak of cotton cultivation, an 820-fold average increase.

In the Paleozoic uplands, historic period erosion caused by widespread timber cutting has resulted in the stripping of vast amounts of residual soil. Clays and silts from the eroded soils have been swept downstream and out of the uplands, but huge quantities of chert gravels have remained to choke stream valleys (Saucier 1983) and cause significant changes in stream behavior.

Stream entrenchment

All landscape degradation in the Lower Mississippi Valley is not simply a product of subaerial weathering and erosion. In geologic time frames, considerable degradation is taking place as a result of stream entrenchment, including both uplands and the Holocene areas. This process is relatively unimportant at present but was much more widespread at various times during the Quaternary.

Entrenchment mechanisms. With climate and sea level *relatively* constant, and excluding the effects of man during the historic period, most streams in the area have reasonably been in a state of dynamic equilibrium during the Holocene. Equilibrium is a balance between form and process and, in fluvial systems, involves an adjustment between velocity, discharge, slope, and load. In a state of dynamic equilibrium, stream channel depth, width, and slope are relatively constant, but channel migration continues in one of several regimes such as meandering. Under these conditions, floodplains experience neither net degradation nor aggradation, but this does not mean that stream valley enlargement (both deepening and widening) is not taking place.

Irrespective of regime, a laterally shifting stream will occasionally impinge against the valley walls and erode them by direct solution or abrasion but mostly by toe scouring and the mass failure of upper banks by slumping. A shifting stream will also scour its bed during flood stages and will erode the underlying deposits or formations to depths well below low-water-stage channel elevations. It is an accepted rule-of-thumb that the Mississippi River will deepen its channel a foot for every foot of stage increase during floods, especially in sharp bends.

Careful examination of the planform of the bluffs bordering the Mississippi alluvial valley and the margins of smaller stream valleys reveals scalloped configurations representing the cumulative effects of multiple episodes of channel impingement. The amplitude and length of the scallops are indicative of the size of the stream involved and, in the case of the Mississippi alluvial valley, indicate lateral erosion and valley widening that were caused by both meandering and braided streams but primarily the latter. This is logical in view of the much greater instability and much higher erosive power of streams flowing in a braided regime.

A state of equilibrium in a fluvial system will be interrupted by any significant change in base level. In the Lower Mississippi Valley area, such changes have frequently occurred, primarily as a consequence of stream-course shortening or lengthening due to diversions and, of course, sea level changes. When the base level is lowered at the mouth of a stream of modest size (such as a Mississippi Valley tributary), a knickpoint or oversteepened segment of stream channel will develop and migrate headward in the fluvial system (Kesel and Yodis 1992). Following passage of a knickpoint, lateral channel migration will respond to the lower water level. Scouring will take place to greater depths, and floodplain construction (e.g., point bar and natural levee deposition) behind a migrating channel will attain a lower elevation. Hence, with time, a net degradation of the floodplain will result, and there will be stream entrenchment.

Sea level lowering during times of waxing glaciation has been the dominant process leading to stream entrenchment. However, before discussing this in more detail, two other processes are worthy of mention. First, in the upper portions of the Western Lowlands (such as in the Advance and Drum Lowlands, and the St. Francis Basin), some entrenchment must have occurred in waning glacial stages because of the tremendous scouring effect of the huge quantities of meltwater and coarse-grained outwash that flowed with high velocity through these areas. There is reason to believe that the scour was episodic and perhaps concentrated around several brief, catastrophic flooding events caused by the bursting of ice-margin lakes (Saucier, in press). Second, there are apparent cases of the rapid entrenchment of the lower segments of upland streams due to shifts in the course of the Mississippi River and the creation of new meander belts during the Holocene. For example, Delcourt and Delcourt (1977) and Alford, Kolb, and Holmes (1983) call attention to the formation of stream terraces along small streams (e.g., Bayou Sara) in the Tunica Hills of Louisiana (north of Baton Rouge) caused by base level lowering and entrenchment brought on by the shift of the Mississippi River from the western side of its valley to its present course along the eastern side. Saucier (1987) similarly discusses evidence for the response of the lower portions of streams in the uplands of western Tennessee (e.g., the Obion and Forked Deer rivers) to base level changes in the adjacent Mississippi alluvial valley. Historic period stream entrenchment due to channel cutoffs has also occurred (Kesel and Yodis 1992) but is not discussed herein.

Effects of sea level changes. Harold Fisk, more than any other individual, is responsible for hypothesizing and dramatizing the effects of sea level variations on the Lower Mississippi Valley. He envisioned that during times of maximum sea level lowering, the entire Mississippi alluvial valley area was swept clean of alluvium and deeply entrenched as far north as Cape Girardeau. Perhaps the most widely used illustration from his 1947 report is a perspective block diagram showing a deeply entrenched canyon-like valley.

Fisk was correct in recognizing the causal relationship between sea level variations and major base level changes, but he misinterpretated the precise response mechanisms. He interpreted the highly irregular base of the alluvium (the suballuvial surface) that was encountered in borings to be characterized by an integrated, dendritic drainage network, the trunk stream of which was graded to the maximum low stand of sea level during the last (Late Wisconsin) glaciation. Furthermore, he portrayed the incised network by detailed contour maps (Fisk 1944). While he did not elaborate on the processes involved, Fisk implied that vertical stream incision was principally responsible, probably by braided streams.

Durham (1962) was the first to challenge Fisk's views and argued that the suballuvial surface beneath the alluvial valley was the product of lateral planation by both braided and meandering streams. This interpretation was generally accepted in concept (Saucier 1974), but proof was lacking and it has not been used as the basis for subsequent contouring for practical and "mechanical" reasons (e.g., Saucier 1964, 1967; Smith and Saucier 1971) until the present time. During the time of the preparation of the folios of quadrangle maps for major basin areas, it was not appropriate to introduce "in midstream" a new concept that could not be applied to the entire valley area. As a consequence of this conservative approach, for more than 15 years there has been incompatibility between accepted process theory and interpretation of the resulting form. This writer now believes that the suballuvial surface is the net result of multiple episodes (glacial cycles) of both lateral planation and vertical incision by meandering and braided streams and, as such, is a composite of multiple discontinuities that form innumerable irregular depressions, troughs, and ridges.

Proof of the new concept has emerged as new subsurface data have become available. Developed originally to dramatize the situation,[1] Figure 7 illustrates for the Mossy Lake quadrangle (scale 1:62,500) in west-central Mississippi why, with significantly increased data, it is necessary to approach contouring in an entirely different way. The 1959 interpretation was made using 18 borings, whereas the 1981 interpretation was made using 70 borings. It is now clear that an integrated drainage network is not present on the surface. It is also apparent that the suballuvial surface is consistently flatter than what was indicated in previous mapping. This has also been found to be the case elsewhere in the alluvial valley area where data from closely spaced borings (e.g., every 500 to 1,000 ft) have become available. As a consequence, and taking advantage of the present opportunity to synthesize knowledge for the entire Lower Mississippi Valley area, a completely new interpretation of

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[1] Personal Communication, 1990, Lawson Smith, Geologist, U.S. Army Engineer Waterways Experiment Station, Vicksburg, MS.

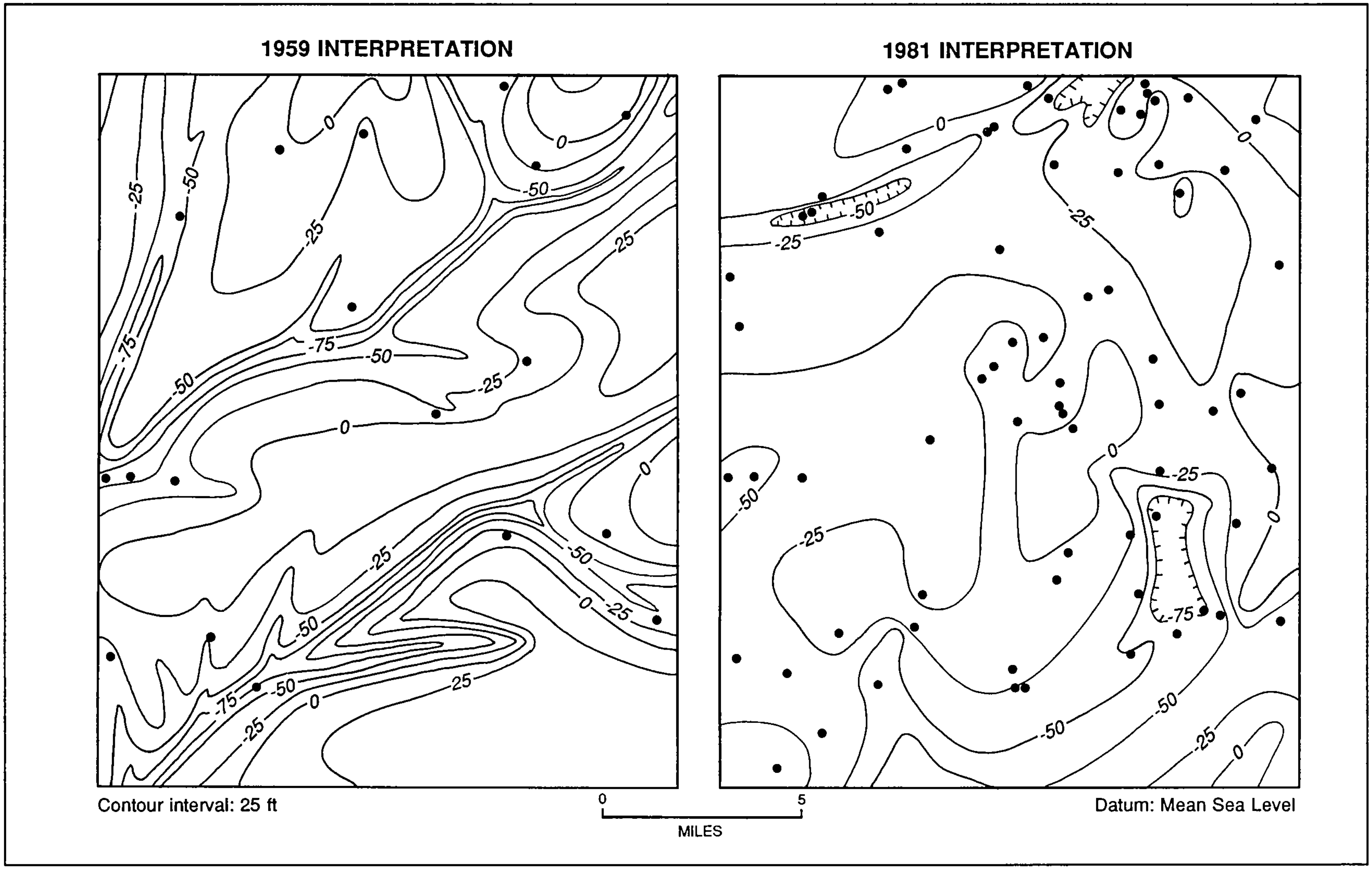

Figure 7. Comparative interpretations of the configuration of the entrenched suballuvial surface in a portion (Mossy Lake quadrangle) of the Yazoo Basin of west-central Mississippi (from Kolb et al. 1968, and subsequent revised editions)

the configuration of the suballuvial surface has been made using computer technology and is discussed in detail later in Chapter 6.

Whereas engineering geological mapping by and for the Corps of Engineers perpetuated the ideas of Fisk in its products, some other scientists and engineers have never accepted them. Groundwater hydrologists and geologists with the U.S. Geological Survey have produced numerous publications in which the thickness of the alluvial aquifer in various portions of the alluvial valley area have been portrayed by contouring data from water wells (e.g., Whitfield 1975, Dalsin 1978, Broom and Lyford 1982, Luckey 1985). None of their interpretations are based on the assumed presence of an entrenched drainage system beneath the alluvium, and indeed their raw data do not show that one is present.

Durham (1962) recognized that vertical stream incision was not a major factor in the alluvial valley area and presented physiographic evidence to substantiate his view, but it was considerably later before apparently supporting stratigraphic evidence emerged. Saucier (1981) made the observation that if sea level was the dominant control that affected the entire valley, then postglacial sea level rise should have resulted in a steady and uniform thinning of the Holocene overbank alluvial sediments (backswamp sequence) between the Gulf of Mexico and the head of the alluvial valley at Cape Girardeau. These deposits represent materials deposited during the last 12,000 years and hence represent a significant time interval during which there was a significant and rapid sea level rise. They are a direct measure of the longitudinal response of the valley to entrenchment and subsequent filling. If sea level change was the dominant process, then there should have been a uniformly decreasing upvalley response of base level. However, this is not the case. Rather, the backswamp sequence rapidly thins from a thickness of over 100 ft in the lower Atchafalaya Basin area to about 60 ft at the latitude of Natchez and from there northward thins very slowly to about 40 ft at the latitude of Memphis. Hence, there is a hinge line in the Natchez area which is not caused by a change in lithology or structural control. Taking into consideration the inherent difference between the geometry of the alluvial valley sedimentary sequence and that of the deltaic sequence, and allowing for the effects of subsidence, this writer has concluded that the areally uniform alluvial valley degradation and subsequent aggradation north of about Natchez was primarily in response to a climatically-induced regime change rather than sea level fluctuation. South of the Natchez area and out onto the continental shelf, however, the Mississippi River drainage incised deeply as a direct consequence of the last glacial-stage low stand.

This writer adds the caveat that his view of the restricted extent of direct sea level response may be applicable only to those glacial cycles of short duration such as the Late Wisconsin. In those cases, limited entrenchment was no doubt at least partially influenced by the fact that sea level was significantly lower than at present for only a few thousand years. If sea level remained well below its present level for tens and even possibly hundreds of thousands of years in older glacial stages, there would have been ample time for the entire alluvial valley area to respond directly to eustatic controls.

An argument can thus be made against sea level change as a major factor in at least the last episode of alluvial valley entrenchment. However, this is by no means the case in the areas of the present deltaic and chenier plains. In these areas, there is abundant evidence, primarily from very large numbers of borings, for well-developed, dendritic, shallowly incised drainage systems on the surface of the now-buried late Pleistocene deposits that were exposed at the surface during the last glacial-stage low stand. Each of the streams that now drain the Prairie and older terrace complexes of southeastern and southwestern Louisiana (e.g., the Amite River) extended their courses as sea level fell and new tributaries developed. The resultant patterns are discussed and illustrated later in this report.

It is important to note that the entrenchment that occurred was not nearly to the extent that Fisk believed and illustrated in his 1944 work. For example, maximum entrenchments in the greater New Orleans area are only a little more than 100 ft below present sea level rather than over 300 ft (Kolb, Smith, and Silva 1975) as Fisk inferred on the basis of much less subsurface information. Fisk failed to take into consideration the appreciable lengthening of stream courses that took place as the shoreline retreated seaward during waxing glaciation. This course lengthening, as much as 100 mi off southwestern Louisiana, meant that stream gradients did not steepen sufficiently to allow deep incision. High-resolution seismic profiling in the northern Gulf of Mexico has produced profiles of buried valleys, such as those of the Sabine River between Texas and Louisiana, that closely resemble those across the currently exposed valleys tens of miles inland (Berryhill, Trippet, and Mihalyi 1982).

Boring control is far more abundant in the deltaic plain area than in the alluvial valley area, but there are still large areas (tens of square miles or more) where there are little data on the configuration of the shallowest Pleistocene surface. Consequently, it is unavoidable that contouring of entrenchments in these areas portrays them to be relatively planar, V-shaped features, whereas in reality, they are known to be otherwise. In the relatively few instances where lines of closely spaced borings are available, such as along interstate highways or expressways in urban areas, they reveal that the entrenchments have narrow, sinuous floodplains with channel-lag deposits, point bar accretion areas, backswamp areas, and even terraces. Unfortunately, features of this scale cannot be portrayed by contouring except in rare instances.

The Mississippi Valley suballuvial surface

Techniques used in reevaluation. As discussed in the last section, previous Corps-sponsored attempts at contouring the suballuvial surface beneath the alluvial valley were biased by the interpreted presence of an integrated, dendritic drainage pattern. The attempts sought to refine the pattern as more information from deep borings became available. In hindsight, they perpetuated what is now considered to be a nonviable concept and led to numerous discontinuities or mismatched interpretations because not all quadrangle-scale maps were revised at the same time.

For purposes of a complete recontouring in this synthesis of the entire alluvial valley area, it was decided that the best approach would be a non-biased, literal interpretation of the available data such as could be accomplished by computer. To this end, basic data used in all previous delineations of the suballuvial surface (Fleetwood 1969; Kolb et al. 1968; May et al. 1984; Saucier 1964, 1967, 1969; Smith and Russ 1974; Smith and Saucier 1971) were assembled and digitized according to location (using Universal Transverse Mercator Grid coordinates) and elevation (relative to mean sea level). This consisted of 5,964 control points, the distribution of which is shown in Plate 15A. Included were any borings that encountered the suballuvial surface beneath Sangamon-stage or younger terraces, the Holocene floodplain, and that portion of the deltaic plain overlying the main entrenched valley. Excluded were areas in southwestern and southeastern Louisiana where the suballuvial surface consists of Prairie Complex deposits that were entrenched a single time in direct response to sea level lowering during the last glacial stage and dendritic patterns do occur (see following discussion).

Plate 15A indicates that the distribution of control points in the alluvial valley area is highly irregular. The greatest concentrations of data are in the Grand Prairie region of Arkansas and the central and southern Yazoo Basin of Mississippi where numerous water wells have been drilled to the base of the Holocene alluvium. In those areas, there are sometimes more than 10 control points per square mile. In contrast, there are relatively few control points in the St. Francis, lower Tensas, and upper Atchafalaya basins, and some areas as large as 100 sq mi have no information whatsoever. This obviously makes valley-wide contouring difficult and of limited value.

The digitized database of nearly 6,000 control points was augmented with a file of 900 points providing the boundary conditions which represent either the valley wall *per se* or arbitrary lines drawn across the mouths of tributaries (or at the limits of available data). To avoid mechanical problems in contouring, control points located very close to the edge of the alluvial valley (within a quarter mile or less of the valley wall) were omitted since relatively few data of this type exist in this zone where the suballuvial surface rises abruptly to and above the floodplain surface into the uplands. Closely spaced and uniformly distributed control would be essential to accurately portray that zone. Therefore, the contouring described below delineates only what may be considered the floor of the entrenched valley rather than its abrupt margins.

Contouring was accomplished using the commercially available *Surfer for Windows* (Golden Software, Inc.) software run on a personal computer using a 486 processor. After considerable experimentation, it was determined that the optimal balance between output resolution and the amount and distribution of input control points was achieved using cells measuring 2,000 by 2,000 m and a contour interval of 25 ft with smoothing interpolators. The gridding algorithm was minimum curvature, and grid file smoothing was achieved using the spline method with two node-insertion default. Cells larger than 2,000 by 2,000 m resulted in excessive generalization in some areas while smaller cells resulted in contour spacing too close to be presented in Plates 17 through 27 and not realistic in terms of the amount and spacing of control points.

In the interpretion of contour maps, it is important to note that all control points more than 2,000 m apart were considered by the computer in processing the data. However, when several control points occur within a single cell, the depth values are averaged. Thus, the resultant contours in areas of closely spaced data do not reflect the total database and are indicative only of the general form of the surface rather than its precise configuration. At the scale of Plates 17 through 27 (i.e., 1:250,000), use of all data in areas of dense control would result in contours too confusing to be intelligible. Readers should also be aware that, for sake of clarity, hachures were used only on alternate contours within enclosed depressions.

Alluvial valley area. In addition to the large-scale plates, the results of the processing of the database are shown in synoptic form in Plate 15B which is a computer-generated, color-coded, hypsometric map of the "floor" of the sub-alluvial surface. The overall pattern of the contours and the color coding graphically illustrate several important characteristics of the surface.

Of greatest significance, the contouring shows no indication whatsoever of the presence of linear entrenchments such as would have been caused by the downward cutting of single stream channels during a single episode of valley degradation. Rather, the overall configuration of the suballuvial surface is one of numerous, irregularly spaced and sized knolls and shallow depressions, suggesting an almost random pattern of scouring which is compatible with current concepts of valley formation. Also of significance is the fact that the surface is much flatter than has traditionally been believed. The flattest areas, and also some of the relatively shallower areas, are largely coincident with the subcrop pattern of the Jackson Group formations (Plate 2), indicating a significant influence of lithology on the formation of the suballuvial surface. Correspondingly, the areas of noticeable hummocky terrain, such as in the lower Yazoo Basin area, are largely coincident with the subcrop pattern of the more easily eroded Claiborne Group formations. Lithologic control is further apparent in the relatively shallow and flat suballuvial surface in the Western Lowlands area where Paleozoic formations resisted intense scouring during the multiple episodes of glacial outwash flow by the ancestral Mississippi River.

Lithology is also a possible factor in the rather abrupt and appreciable deepening (downvalley steepening) of the suballuvial surface between the latitudes of Natchez and Baton Rouge, but structural control is probably a more viable explanation (see Chapter 8). The sudden steepening of the surface north of Baton Rouge closely coincides with the zone of coast-parallel growth faults, and the large knolls and depressions that characterize the surface south of Baton Rouge strongly suggest the effects of diapiric salt domes. If the suballuvial surface began forming well back into the Quaternary as is presently hypothesized rather than during the last glacial stage, it is entirely logical that such tectonic processes are strongly manifest on the surface.

In Plate 16, modified from Autin et al. (1991), seven typical cross sections show the general lithologic characteristics of the Quaternary units and the configuration of the suballuvial surface as taken from the contouring described above. A comparison of the present interpretation with that used in the original publication (based on the concept of an entrenched, integrated, dendritic network) reveals some important differences. As expected, the surface is now depicted by much smoother profiles. However, other differences are apparent that are not readily explainable. For example, whereas other than for some smoothing, the profiles in Sections A-A' to D-D' in the two interpretations are actually very similar, but they are significantly different in Sections E-E' and F-F'. In Section E-E', the latest interpretation shows the average cross-valley elevation of the suballuvial surface to be about 100 ft lower than that in the original interpretation. The smoothed profile generally coincides with the deepest parts of the entrenchments shown in the latter. In contrast, in Section F-F', the converse is true. The average elevation of the surface in the latest interpretation is 50 to 100 ft shallower than in the original interpretation with the smoothed profile generally coincident with the highest points on the surface rather than the deepest ones. It is suspected that this may be an artifact of (or at least enhanced by) the computer contouring of the data, but the reasons for that effect are unknown. Once again, it is important to emphasize that a precise, literal interpretation of either the contours or the cross-valley profiles should be avoided.

Deltaic and chenier plains. Several attempts have been made since Fisk (1944) to contour the surface of the shallowest Pleistocene deposits (Prairie Complex) beneath all or part of the deltaic and chenier plains using large amounts of new data each time (Coleman 1966b; Kolb 1962; Kolb, Smith, and Silva 1975; Kolb and VanLopik 1958; May et al. 1984; Nichols 1959; Orton 1959; Saucier 1963, 1969; Saucier et al. 1984, 1991b; Schultz and Kolb 1950; Smith, Dunbar, and Britsch 1986). For purposes of this synthesis, the efforts were combined to produce the most accurate delineation of that surface currently available (Plates 25 to 27). As can be imagined, considerable judgment had to be exercised to achieve a matching of the various contouring attempts.

Three types of problems had to be addressed in contouring the suballuvial surface in these areas. First, the contouring had to be subjectively merged with the computer contouring of the main entrenched valley at the margins of the latter. Some arbitrary shifting of contours was required to create a realistic situation. Second, on Plates 25 and 27, it was necessary to highly generalize the contouring in the New Orleans-Lake Pontchartrain areas where vast amounts of data are available and contouring is actually possible using a 10-ft interval (see Chapter 7 and Figure 48). A related problem, made critical by the scale of the mapping, was how to depict the narrow zone in which the present channel and relatively recent abandoned distributaries of the Mississippi River have deeply scoured into the Pleistocene deposits. Whereas these steep-sized entrenchments have been portrayed in larger-scale mapping (e.g., Saucier et al. 1984, 1991b), it was decided that any attempt to portray them by contours would be impractical. Consequently, contours showing the older entrenchments (caused by the incising of local drainage) are broken where interrupted by the younger scoured channels. Readers needing more detail in these areas are referred to the original sources. Third (and in striking contrast), because of the extreme paucity of data beneath the southern and eastern portions of the deltaic plain (Plate 27), contours in those areas are highly generalized and are little more than form lines. In response to this situation, a 50-ft contour interval is used rather than a 25-ft one.

A modest amount of data (control points) is available from several studies in portions of the chenier plain, and the small number and smoothness of the contours in that area (Plate 26) do not reflect a lack of information. Rather, it reflects the extreme flatness and low slope of the Pleistocene surface which is generally shallower than 50 ft below sea level at the present shoreline. Away from the major streams (e.g., the Calcasieu and Mermentau rivers), entrenchments apparently are at most only a few tens of feet deep and only a few hundred feet wide. Otherwise, relief and slopes on the buried surface are only a few feet per mile or less.

Depositional Landscapes

Terraces and terrace formation

Fluvial terraces. In terms of surface morphology, a terrace is a relatively linear level to gently inclined surface that is bounded on one side by a steeper ascending slope (e.g., a dissected upland) and on the other by steeper descending slope (a scarp) to a lower level (e.g., a stream floodplain). In the Lower Mississippi Valley area, terraces occur both as narrow benches along modern stream valleys and as broad, coast-parallel plains. Geologists typically attempt to correlate terraces from one area to another by considering such parameters as absolute elevation, slope, surficial features, and degree of dissection.

Terraces in this region were first recognized and described in coastal Louisiana well back into the 19th century. They were correlated with a sequence of coast-parallel terraces that are well developed along the Atlantic and Gulf Coastal Plains south of New York and extending into Mexico. Until nearly the beginning of the 20th century, this sequence was believed to be of marine origin; i.e., the terraces were created by either wave-cutting (planation) or wave-building (Shaw 1918). Cyclical sea level variations were assumed to be the reason for the presence of discrete terrace levels separated by erosional scarps.

About the same time, both Russell (1938) and Fisk (1939) advanced the idea that the coast-parallel terraces of Louisiana were of freshwater, fluvial origin rather than marine origin and were actually relict alluvial and deltaic plains that had been uplifted inland and downwarped Gulfward. Along inland streams such as the Red River, they argued that terraces were not "cut terraces" due to the lateral planation of older formations by stream migration as had been traditionally believed. They demonstrated that the terrace surfaces had more than just a thin veneer of sediment and were underlain by complete fluvial sequences consisting of a fine-grained topstratum (overbank deposits) and a coarse-grained substratum (channel deposits). Hence, Russell and Fisk can be credited with originating the concept of Lower Mississippi Valley terraces as terrace formations--a morphologic terrace plus its underlying alluvial deposit rather than just the terrace surface.

Moving forward with the idea that the Louisiana terraces each represented a complete alluvial valley/deltaic plain sequence attributable to an interglacial stage, Krinitzsky (1949a) made the first attempt to correlate terraces along inland streams and along the margins of the alluvial valley with those in Louisiana. Thus, all terrace exposures were correlated with the Prairie, Montgomery, Bentley, or Williana coast-parallel terraces.

Even though these terraces characteristically are distinctive physiographic units, they are not always unequivocal, and there is considerable opportunity for different interpretations. For example, Doering (1956) agreed with Fisk's view of terrace formation but never accepted his correlations and strongly pursued an alternative interpretation. As pointed out by Frye and Leonard (1954), ample areas of disagreement in correlations occur because of problems caused by colluvium from older surfaces, dissection of terrace margins due to erosion, and loess veneers. In each instance, the original terrace topography will be altered, making correlations on the basis of elevation or surface features quite tenuous.

In this synthesis, the writer accepts the now well-established concept of a terrace as being the geomorphic surface associated with a definable sedimentary body. Under this concept, it would appear feasible to correlate terraces on the basis of the lithology or internal structures of the sedimentary body rather than relying solely on surface morphology. However, this is not always possible or advisable. It is true that the terraces are all underlain by a sedimentary sequence with a topstratum and a substratum, but from one area to another the topstratum may consist of deposits of widely varying composition laid down in any of the several distinctively different environments of deposition (see following discussion) that characterize an alluvial valley or a deltaic plain. Thus, lithology may vary from area to area within a given terrace and from one terrace to another. In sharp contrast, the substratum in all terraces is so similar in lithology that there is no sound basis for differentiation.

The fact that downvalley terrace profiles are not always parallel to each other or to the adjacent Holocene floodplains further complicates terrace correlation along stream valleys. Relative topographic relationships may change significantly from area to area. For example, whereas a terrace may occur as a true topographic terrace in the upstream areas of a drainage system, the same terrace may occur at floodplain level farther downstream or even plunge into the subsurface.

It has been recognized for several decades that fluvial terraces in the Lower Mississippi Valley area include deposits laid down in widely differing environments. This has complicated and confused the correlations. However, with only minor questioning (Durham, Moore, and Parsons 1967), the basic sequence advocated by Fisk was perpetuated until only recently as being the best interpretation. Finally, however, new problems and issues have emerged that have necessitated, as a minimum, a change in nomenclature. Specifically, the Prairie terrace of Louisiana was found to consist not only of deposits laid down in different environments but also of deposits representing at least two major interglacial stages: the Middle Wisconsin or Farmdalian and the Sangamon shown in Figure 4 (Saucier 1977b). In reality, the Prairie terrace consists of two alluvial sequences of significantly different age that morphologically form a single terrace. To avoid further confusion and establish a better framework for future studies, Autin et al. (1991) advocated, and it is accepted herein, that the term "complex" be substituted for "terrace" for each of the Quaternary alluvial sequences in the Lower Mississippi Valley area. Also contributing to this decision was the growing realization that the two oldest terraces of the Fiskian concept (the Bentley and Williana terraces) may be a single, pre-Quaternary depositional sequence of non-Mississippi River origin.

Readers should note that in this synthesis, the terms terrace and complex are not used to refer to the areas of Wisconsin-age glacial outwash (valley trains) or the various levels therein. Entire features such as Macon Ridge meet most of the morphologic criteria of a terrace as do the several levels that constitute its surface, and some workers have referred to them as braided-stream terraces. However, they are considered herein as landforms within an alluvial sequence rather than a sequence *per se.*

Lacustrine plain terraces. It has been discovered in the last several decades that not all of the terraces in the Lower Mississippi Valley area are of fluvial origin. Detailed mapping and interpretations of depositional environments in the Ouachita River area of northeastern Arkansas and southeastern Arkansas (Fleetwood 1969) revealed that a low terrace contained beaches and barrier bars similar to those found previously along the lower Ohio River (Finch, Olive, and Wolfe 1964; Olive 1966). In that area, the features were interpreted as evidence of a former extensive lake that were created by impoundment of river water by glacial outwash. After careful mapping of the accordant crest elevations and field verification of the lacustrine origin of the features in the Ouachita River area, Saucier and Fleetwood (1970) presented evidence that during the Early Wisconsin stage, impingement of a mass of glacial outwash (Macon Ridge) against the uplands south of Monroe caused impoundment of the river and the creation of what was designated as Lake Monroe. The process that created the lake has been long recognized (Shaw 1915) and was referred to by Russell (1938) as alluvial drowning. Saucier and Fleetwood (1970) found that terraces with the beaches and bars were actually relict lacustrine plains and assigned them to the Deweyville terrace sequence (a confusing and misleading assignment, as will be discussed in Chapter 7). The plains consist both of areas where the underlying Quaternary or Tertiary deposits have been planed by wave action (erosional terraces) and areas where lacustrine sediments have accumulated (depositional terraces).

Discovery of a lacustrine plain terrace in the Ouachita River area sensitized this writer to the possibility that terraces elsewhere along the margins of the alluvial valley may be of similar origin. While mapping fluvial deposits along the Obion, Hatchie, and other rivers in western Tennessee and Kentucky in the 1980s, the writer observed two terraces (the Finley and Hatchie terraces) that exhibited nil or even reverse downstream slopes in the lower reaches of the stream valleys. This was interpreted as evidence of alluvial drowning along these rivers by glacial outwash deposition in the Mississippi alluvial valley. More careful examination of the terraces revealed that they also contained beach and gravel bar features along their upland margins, and hence they have been described as being of lacustrine origin (Saucier 1987) and correlated with two episodes of outwash deposition in the alluvial valley. Terraces of similar age and origin may occur along streams such as the Coldwater and Tallahatchie in northwestern Mississippi, but terrace mapping in that area has been precluded by the creation of large reservoirs.

Concept of allostratigraphy

Fisk (1944) and others working in the Lower Mississippi Valley area regarded fluvial terraces as formations. However, they do not meet the criteria for formations as set forth in the stratigraphic code, and they have never been formally accepted as such. Because of the highly variable nature of alluvial sediments and the frequent lateral and vertical unconformities, terraces cannot be considered as lithostratigraphic units. Moreover, as determined by recent work, terraces sometimes consist of units of substantially different age or are time transgressive; hence, they cannot be considered as chronostratigraphic units. Substitution of the term "complex" for "terrace or terrace formation," as discussed earlier, should be considered only an interim measure since it mitigates but does not solve the terminology problem. Allostratigraphy (North American Commission on Stratigraphic Nomenclature 1983) is an entirely appropriate scheme for defining heterogeneous geologic units that are genetically related (Autin 1992), but this approach has not yet been systematically applied in the Lower Mississippi Valley area.

An alloformation, the fundamental allostratigraphic unit, is a mappable unit of sediment that is definable and identifiable on the basis of its bounding discontinuities rather than its age or character. Discontinuities may be erosional unconformities, such as the base of channel scouring during lateral stream migration, wave-cut surfaces, and paleosols, or they may be constructional surfaces.

As a test of the applicability of allostratigraphy to a Quaternary fluvial terrace, Saucier (1988) considered the entire Prairie Complex in the Lower Mississippi Valley and was able to identify and tentatively delineate 13 alloformations. These include seven different basic depositional environments (e.g., nearshore marine, deltaic, meander belt) ranging from Sangamon to Late Wisconsin in age. Although suggested names were assigned to the 13 alloformations, they are not used in this synthesis since they have not been formally described and properly introduced in the literature. The test demonstrated that it is feasible and eventually it may be desirable to subdivide all Quaternary terraces (both fluvial and lacustrine) into possibly as many as several dozen alloformations--a step that would greatly facilitate the correlation of isolated terrace units. This would also provide a framework for future more detailed terrace mapping wherein specific features, such as abandoned channels within meander belts, could be designated as allomembers.

Concept of depositional environments

Geomorphologists are concerned with processes and landforms, while sedimentologists are concerned with the nature and origin of sedimentary sequences. Depositional environments are to the sedimentologist what facies or facies models are to the "hard rock" geologist--a means of classifying a modern or ancient sedimentary sequence according to its origin and characteristics. In both cases, parameters such as grain-size distribution, internal structures, and faunal (or fossil) content are used to infer from modern analogs what the physical, chemical, and biological conditions were like where the sediments were laid down. The knowledge of these "environmental" conditions and the geomorphic processes at work thereby makes it possible to correlate and predict the range and distribution of soil types and their physical properties. Hence, the recognition and delineation of depositional environments is the fundamental classification scheme long adopted for use in the engineering geologic study of alluvial and deltaic sequences. To geologists, depositional environments are geographically restricted complexes that are generally described in geomorphic terms. Reconstructions of environments of deposition by sedimentologists are analogous to the building of facies models by geologists.

Classifying sedimentary sequences according to environments of deposition is particularly feasible when dealing with geologically young situations where the geomorphic processes responsible are not different from those operating somewhere in the area at present. In this synthesis, reference is occasionally made to the environments of deposition of Sangamon-stage or older deposits and landforms, but they do not form the basis for classification. In contrast, depositional environments are the primary means of classifying and subdividing the Holocene and Wisconsin-stage deposits of the alluvial valley and deltaic and chenier plains.

Reconstruction of depositional environments for alluvial sequences is a valuable tool in geologic and geomorphic interpretations, but it is absolutely essential in the study of deltaic sequences. As demonstrated so well by Lankford and Shepard (1960), it is possible to correlate subsurface facies in, for example, a line of borings using only sedimentary properties like grain size, consistency, or water content. However, the combination of these properties with others like mineralogy, faunal content, and sedimentary structures makes it possible to infer depositional environments and thereby reconstruct the cycles of delta lobe growth and deterioration.

Environments of deposition have been the basis for the detailed, quadrangle-scale mapping of the entire Lower Mississippi Valley area over the last several decades and there are dozens of publications that deal with the origin, distribution, and geotechnical properties of deltaic environments. Nevertheless, there is no generally accepted classification. Workers have variously regarded and assigned the deltaic environments which, depending on viewpoint, can be considered as terrestrial or marine, or a separate transition category between the two. For this synthesis, the writer has devised the following breakdown, relying heavily on sources such as Fairbridge and Bourgeois 1978, Kolb and VanLopik 1958, and Miall 1987:

A. Terrestrial
    1. Fluvial
        a. Alluvial fan/alluvial apron.
        b. Valley train (braided stream).
            (1) Channel.
            (2) Island/braid bar.
        c. Meander belt.
            (1) Natural levee.
            (2) *Crevasse splay.*
            (3) Distributary.
            (4) Point bar.
            (5) Abandoned channel.
            (6) Abandoned course.
        d. Backswamp/flood basin.
    2. Lacustrine
        a. Lacustrine.
        b. Lacustrine delta.
    3. Eolian
        a. Loess.
        b. Sand dune.
B. Paralic
    1. Deltaic
        a. Inland swamp.
        b. Interdistributary/intratidal marsh.
        c. *Intradelta.*
        d. *Prodelta.*
        e. *Mudlump.*
    2. Deltaic Marine
        a. *Bay-sound.*
        b. *Beach/barrier.*
        c. *Reef.*
        d. Chenier.
        e. *Nearshore Gulf.*

All of the depositional environments listed above are described in this synthesis, but not all have been mapped and shown in Plates 4 through 14. Those indicated in *italics* either are too limited in size or extent to be portrayed on the maps (e.g., mudlumps), or they occur only in the subsurface (e.g., prodelta). A few minor environments such as deltaic tidal channels and mangrove swamp that have been mapped by others have been omitted from consideration herein to avoid excessive detail.

Fluvial Environments and Processes

Upland graveliferous deposits

Upland formations of Tertiary age are unconformably overlain by a mostly continuous blanket of graveliferous deposits on Crowley's Ridge and in a 10- to 50-mi-wide band east of the alluvial valley from western Kentucky to southeast Louisiana. In Plate 2, this unit is designated the Upland Complex (Autin et al. 1991) and constitutes the most widespread Quaternary unit of single origin. Typically, the deposits consist of brightly colored masses of sands and gravels from a few feet to several tens of feet thick that immediately underlie loess deposits on hills and ridges but that are absent on lower slopes and in creek bottoms. On a regional basis, the deposits are relatively uniform in thickness, but well-defined channels several miles wide and several tens of miles long have been observed where the deposits reach thicknesses in excess of 100 ft. The trends of the channels are often evident by the locations of gravel pits or quarries where the thicker gravel deposits are being exploited for road metal or construction purposes.

Upland Complex deposits have received considerable attention with dozens of papers in the literature over the last 80 years discussing their age and origin. In western Kentucky and Tennessee, they have been designated the Lafayette gravel (Potter 1955a), but elsewhere they have generally been designated as the Citronelle formation (Matson 1916; Doering 1956, 1958). They have been correlated with stratigraphic equivalents that extend from Virginia to Texas.

All investigators of the last several decades agree that the deposits are fluvial sediments laid down primarily by braided streams; however, the source of the large volume of clastic materials has been argumentative. Self (1993) has recently articulated the views of several workers that the gravels were derived from Mississippian and Cretaceous formations that were eroded and transported outward as a broad alluvial fan as a result of uplift and rejuvenation of the Nashville Dome (Figure 1). This model is certainly compatible with the mineralogy and petrology of the deposits that indicate an Appalachian provenance in western Kentucky and Tennessee and a continental interior origin for those on Crowley's Ridge (Potter 1955b). The postulated uplift event is believed to have started in the late Tertiary, but the time of arrival of the graveliferous materials into the Lower Mississippi Valley area is variously estimated from late Miocene (May 1980) through the Pliocene to the early Quaternary (Doering 1956).

Fisk (1939) refused to recognize the Citronelle as a Tertiary/Quaternary continental, fluvial blanket of eastern source, proposing instead that all the deposits in Louisiana were Pleistocene terrace deposits. In the Florida Parishes area of that state, based on physiographic data from topographic profiles, he designated the graveliferous deposits as the Williana and Bentley terraces and assigned them to the Aftonian and Yarmouthian interglacial stages (Early Pleistocene shown in Figure 4). Krinitzsky (1949a) later applied this concept to all graveliferous deposits in the Lower Mississippi Valley area.

Subsequent work by Cullinan (1969) and Rosen (1969) have convincingly refuted the views of Fisk, and the concept of the Citronelle as a late Tertiary, braided alluvial fan is once again widely accepted. However, there are complications. Terrace levels, although not of the type described by Fisk, do exist as physiographic elements in the landscape, and some of the deposits appear to be of unquestionable glacial origin (and hence are of Quaternary age) rather than an alluvial fan. No deposits of glacial outwash origin have been identified in the Florida Parishes (Campbell 1971); however, large numbers of cobbles and boulders of chert, quartzite, and sandstone have been found elsewhere in Louisiana (Woodward and Gueno 1941) and in Arkansas and Missouri which are believed to represent ice-rafted materials carried downvalley during outwash events.

Attempting to reconcile the inconsistent evidence, this writer has proposed an alternate hypothesis (Autin et al. 1991). The Upland Complex throughout the area is considered to be primarily a nonglacial, Plio-Pleistocene fluvial deposit of eastern (Appalachian) and continental interior origin, but the various terrace levels that are present (especially the higher ones) are regarded as being *dominantly* erosional rather than depositional as envisioned by Fisk. The levels may represent local reworking and partial redistribution of the original alluvial fan deposits to base levels that were cyclically lowered during the Early Pleistocene. Locally, they may incorporate topographically indistinguishable inclusions of Pleistocene outwash, especially along the valley margins.

Alluvial fans/alluvial aprons

Within the Mississippi alluvial valley, there are four depositional environments that were significantly underemphasized or not mentioned by Fisk (1944) who presumably considered them to be of secondary importance. They are the crevasse, distributary, sand dune, and alluvial fan environments. Although they once may have been deemphasized since few engineering projects were in their areas of occurrence, in the broader context of contemporary engineering investigations (e.g., basin planning, environmental evaluations), they are now considered equally as important as "major environments" and are the subject of full discussion herein. The alluvial fan environment, which can also correctly be regarded as a landform, is the most important of the four.

Based on the observations that were made by this writer over the last decade and are described in the following paragraphs, it is possible that alluvial fans contain a wealth of unexploited chronostratigraphic evidence that could make major breakthroughs in reconstructing area geologic and climatic history.

Simply defined, an alluvial fan is a low, gently sloping mass of fluvial sediment shaped as an open fan or segment of a cone that is deposited where a stream discharges from an upland into a basin or onto a plain (Figure 8A). In humid areas, the cause of deposition is not a sudden reduction in stream gradient, as is commonly believed, but rather the ending of the constriction of the stream valley. In many respects, it is analogous to a delta. Where streams are relatively closely spaced such as along a bluff line or escarpment, alluvial fans coalesce to form an alluvial apron (Figure 8B).

Alluvial fans have been recognized and investigated by geologists and geomorphologists since the mid-nineteenth century, but mostly in arid climates where fans are exceptionally well developed. In the Lower Mississippi Valley area, fans have been delineated as part of the systematic program of quadrangle-scale mapping and in other studies, but Smith (1983) has been the only investigator to comprehensively examine their geomorphic development. His consideration of alluvial fans in the Yazoo Basin is a major contribution to understanding geomorphic processes and fan stratigraphy in a humid area.

The distribution of alluvial fans of moderate to large size in the Lower Mississippi Valley area is shown in Plates 4 through 11. They occur primarily at the base of the uplands of Tertiary or older age that border the eastern side of the alluvial valley between Baton Rouge and Cairo, along both sides of Crowley's Ridge between Cape Girardeau and Helena, and at the base of the Ozark Escarpment north of the mouth of the White River. Alluvial fans are absent or poorly developed where the alluvial valley is bordered only by formations or deposits of Quaternary age (e.g., the Prairie Complex). Cumulatively, alluvial fans constitute an area of about 810 sq mi, or about 2.4 percent of the alluvial valley. Individually, they range in size from less than 0.1 sq mi to approximately 40 sq mi. There is no accurate count of the number present in the alluvial valley area. Smith (1983) studied 34 in just a small portion of the Yazoo Basin, and McCraw (1991) estimated that there are over 200 present just along the eastern valley wall between Louisiana and Kentucky. The writer estimates there are at least 100 along the sides of Crowley's Ridge. In some cases (Plates 5 and 6), they coalesce to form essentially continuous, several-mile-wide aprons that are tens of miles long.

Alluvial fans typically occur at the mouths of drainage basins too small to support perennial base flow of any consequence and experience growth only during times of flash flooding. However, they also occur, and reach their best development, along permanent streams as large as the Current and St. Francis rivers.

Smith (1983) concluded that there is a definite relationship between alluvial fan size and slope and drainage basin characteristics. Fans are primarily

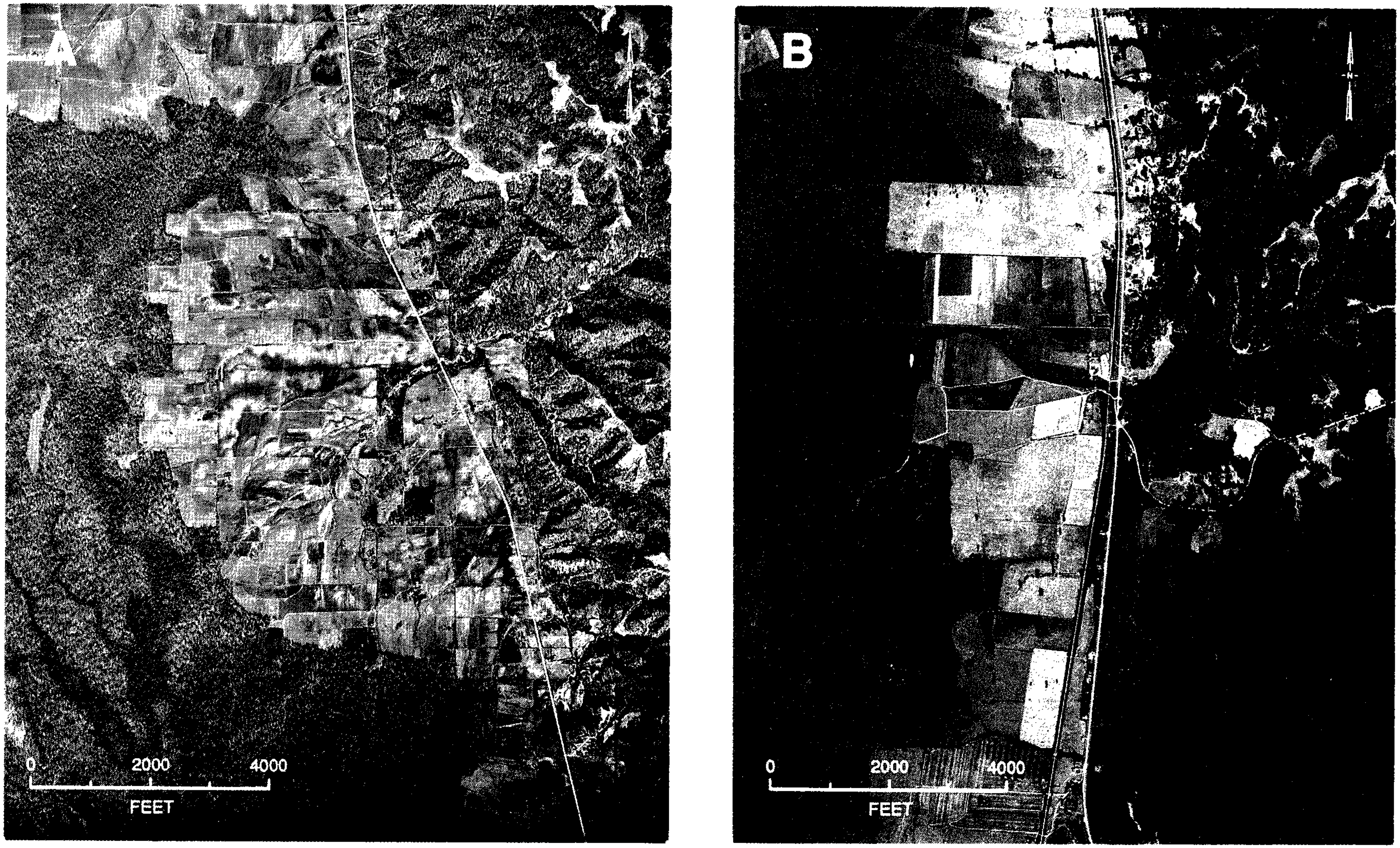

Figure 8. Alluvial fans. A: Small discrete symmetrical fan of unnamed creed near Leverett in Tallahatchie Co., MS (Plate 8); B: Coalesced small fans forming alluvial apron (in cultivation) west of U.S. Hwy 61 near Yokena in Warren Co., MS (Plate 9)

controlled by the geology (lithology) of the source area, the climate of the region, and the time available for fan development. Although there is a regional trend in the amount and character of precipitation in the Lower Mississippi Valley, for all practical purposes climate can be considered a constant. Source-area lithology and time are highly variable factors and essentially govern the size and distribution of alluvial fans.

The largest alluvial fans and aprons occur where loess deposits are thickest. This thickness relationship is highly logical considering the highly erodible nature of those deposits. Secondarily, there is a correlation with the distribution of the sands and gravels of the Upland Complex and the more readily erodible deposits of Tertiary age such as the Sparta Sand. These correlations do not occur, however, with regard to the fans of unusually large size on the rivers that discharge from the Ozark Plateau.

The degree of alluvial fan development strongly reflects the age of the underlying fluvial deposits. Along streams of a given drainage basin size and lithology, fans are significantly larger where they formed on valley trains of Early Wisconsin age than they are on ones of Late Wisconsin age. In turn, fans are poorly developed or absent where they overlie Holocene meander belt deposits. This relationship is clearly apparent along the eastern side of Crowley's Ridge as shown in Plate 6. Since time has been shown to be an important factor, it is logical that fans that have had more than 12,000 years to develop will have a larger average size than those that have been able to develop for only a few thousand years. The only known significant departure from this relationship is in the New Madrid Seismic Zone where McCraw (1991) has shown that as a result of earthquake-induced subsidence, fans adjacent to Reelfoot Lake which overlie valley train deposits are anomalously small.

Figure 9 illustrates three types and sizes of alluvial fans in the alluvial valley area and the appreciable variation in channel patterns on their surfaces. Figure 9A portrays the fan of Teoc Creek which is located just northeast of Greenwood in the east-central Yazoo Basin (Plate 8). That fan is representative of those that occur east of the alluvial valley and along Crowley's Ridge. It is symmetrical and classically shaped with an area of about 6.5 sq mi and a height (between apex and distal margin) of about 40 ft. As described by Smith (1983), this type of fan is created by the lateral coalescence and vertical buildup of sand and gravel bars, channel bed deposits, intersection point lobes, and sheetflood deposits.

Figure 9B portrays the fan of the Castor River which lies in the Advance Lowland where that stream discharges from the Ozark Plateau (Plate 4). This fan has an area of about 28.5 sq mi but much shallower slopes with a difference in elevation of only 15 ft between apex and distal margins. It has developed across the entire width of the Advance Lowland and its shape is strongly influenced by existing landforms, but a radial pattern of abandoned channels (distributaries), each with a low natural levee ridge, is clearly evident. The overall configuration of the fan is similar to that of the adjacent St. Francis

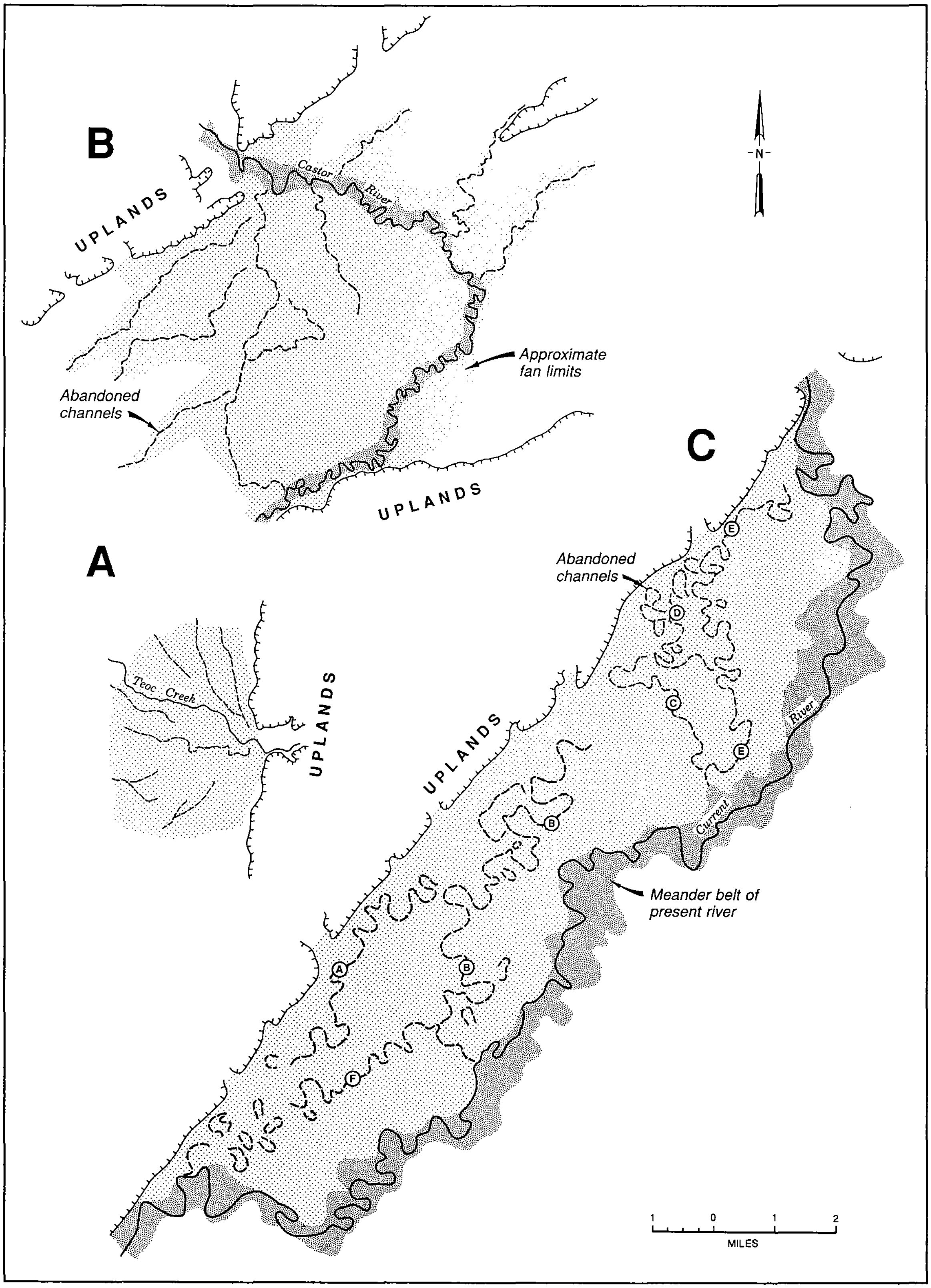

Figure 9. Configurations of and channel patterns on three alluvial fans representing a range of types and sizes: A=Teoc Creek, B=Castor River, C=Current River

River fan and resembles a digitate (bird-foot) delta which is indicative of deposition under very low-gradient conditions. Unfortunately, there are no clear cross-cutting relationships that would allow reconstruction of a sequence or chronology of abandoned channels on the fan surface.

The largest and most extraordinary alluvial fan in the alluvial valley area is that of the Current River (Figure 9C). It is a highly elongate, 18-mi-long feature that is bounded by the Ozark Plateau on the northwest and the present meander belt of the Current River on the southeast (Plate 5). It has an area of about 52 sq mi (including the present meander belt of the Current River) and declines in elevation only about 30 ft between its apex and southwestern margin. The most unusual aspect of this alluvial fan is the sequence of six well-preserved abandoned channels (designated A the youngest through F the oldest, Figure 9C) and their associated abandoned channels (cutoffs). Rarely do cross-cutting relationships provide such an unambiguous picture of relative ages, and rarely does channel geometry provide such an interesting indication of variations in stream discharge. The fan has been the subject of an intensive archeological survey, and estimates of channel ages have been made based on numerous sites. However, the channel sequence has not been studied geomorphologically. The prospect that this may eventually be done was greatly diminished about 10 years ago when the already relatively flat fan surface was land-leveled for intensive agricultural development.

Valley trains (braided stream)

Fisk (1944) is generally regarded as being the first to recognize the presence of valley train deposits within the Lower Mississippi Valley and certainly was the first to understand their origin and significance. He referred to them as alluvial fans and concluded they were laid down by the Mississippi, Ohio, and Arkansas rivers while flowing in braided regimes. He categorically included them as dissected alluvial plains that were of different origin and age from the "Recent" alluvial plain.

During the several decades of quadrangle-scale mapping and interpretation of depositional environments in the region, beginning with Saucier (1964) and Kolb et al. (1968), the valley trains were consistently referred to as braided-stream surfaces, braided-stream terraces, or braided-relict alluvial fans. In 1991, Autin et al. recommended adoption of the more generic term valley train as being more diagnostic of the mode of origin and not simply descriptive of surface features.

According to Fisk (1944), all valley train deposits were laid down in a single episode of outwash deposition attributable to waning of the Late Wisconsin ice sheet. Hence, they were coincident with the last stages of rising sea level and the early stages of standing sea level. That view was accepted without question until Saucier (1968) discerned stratigraphic evidence and obtained radiocarbon dates affirming that the Lower Mississippi Valley experienced two discrete episodes of valley train formation. He correlated those episodes with the waning of the Early and the Late Wisconsin glaciations, a view that has generally remained unchanged although a few small areas of valley train in the Western Lowlands area are now regarded as possible candidates for assignment to the late Illinoian or early Sangamon stages. This differentiation is consistent with and is a more logical explanation for significant differences in valley train surface morphology, relative elevations, and the degree of subsequent loess and alluvial fan deposition.

Since the 1940s, it has also become apparent that there is no convincing evidence for attributing Macon Ridge to the Arkansas River. Fisk (1944) concluded that the river was the source of the outwash based on the location and configuration of the deposits relative to the Arkansas Valley and the reddish color of its sediments. However, the geographic relationships may only be coincidental. The sediment color can be explained by the recently recognized greater age of the deposits, and studies in the Arkansas Valley west of Little Rock have failed to find evidence for outwash in that area (Autin et al. 1991). Therefore, it is currently believed that only the Mississippi and Ohio rivers were responsible for valley train formation. The Mississippi River was the sole source of the outwash deposits of the Western Lowlands, whereas all others in the alluvial valley area probably represent a mixture of both Mississippi and Ohio river sediments.

Channel patterns. The relict patterns of wide, frequently branching channels separating irregular braid bars and interfluve areas are the most apparent and diagnostic recognition criteria for valley trains. They are typically clearly discernible and often mappable using both aerial photos and topographic maps. However, the patterns are distinctively different on the valley trains of different age.

Valley trains of Early Wisconsin age of the Western Lowlands and Macon Ridge stand several tens of feet higher than the adjacent Holocene floodplains (the local base levels of erosion); consequently, local drainage is incised into the deposits. Since nearly all local drainage concentrated in the depressions of the relict braided channels, the present drainage pattern is essentially an underfit system within the confines of the last braided stream pattern that existed on the valley train surface (Figures 10A and 11A). While the present system mimics the trends of the braided channels, the degree of stream incision and erosion has been sufficient to remove or obscure evidences of the actual former bank lines. Therefore, the actual widths of the braided channels cannot be ascertained.

The pattern of braided channels on the valley trains of Late Wisconsin age, such as on the Malden Plain, is strikingly different in several respects (Figure 10B), including the significantly larger size (width and length) of individual channels. Because those surfaces lie at or even slightly below the level of the adjacent Holocene floodplains and there has been no incision of local drainage, the relict braided channel patterns are remarkably well preserved (see also Plates 5 and 6). Relief is extremely low (no more than several feet), but the former braided channel bank lines are clearly apparent on aerial photos

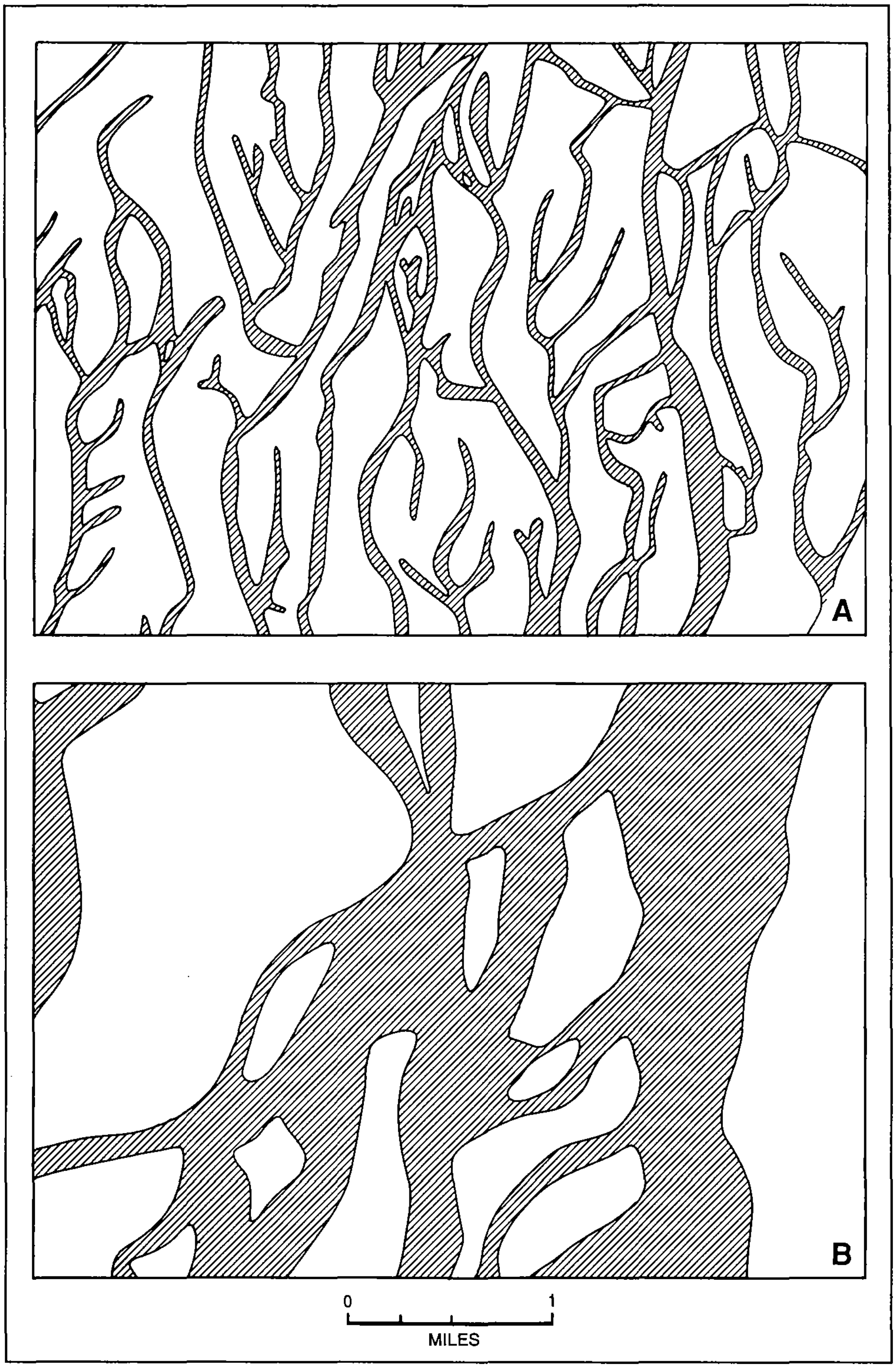

Figure 10. Typical braided-channel patterns on (A) Early Wisconsin valley train in Mangham quadrangle, Richland Ph., LA; and (B) Late Wisconsin valley train in Dee quadrangle, Poinsett Co., AR (adapted from Whitworth 1988, Saucier 1964)

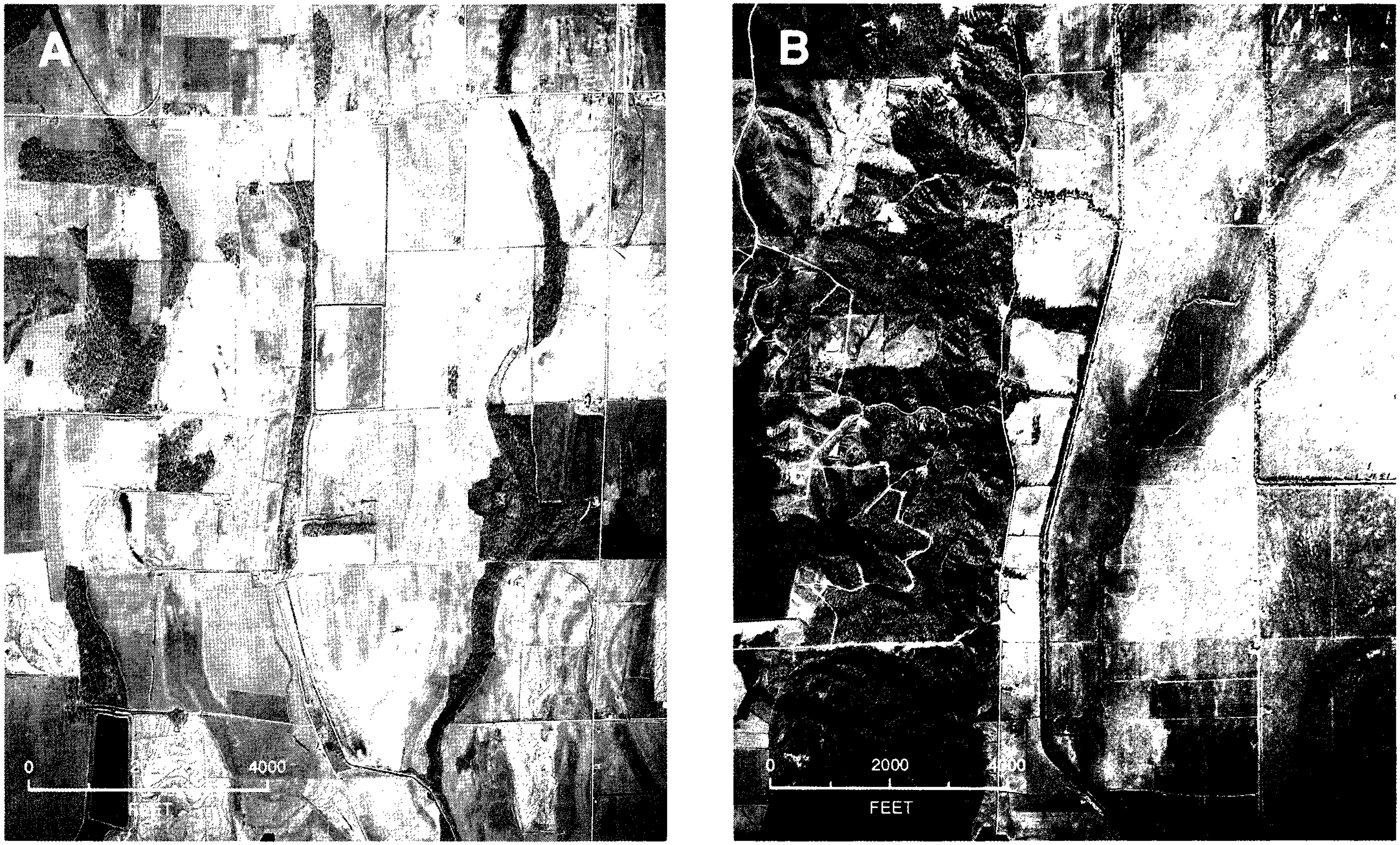

Figure 11. Relict braided channel patterns on valley trains. A: Early Wisconsin-age channels near Weiner in Poinsett Co., AR (Plate 6) marked by entrenched local drainage; B: Late Wisconsin-age relict channel east of Crowley's Ridge near Harrisonburg in Poinsett Co., AR (Plate 6) marked only by darker soil tones (between arrows)

because of soil changes (Figure 11B). On the youngest and lowest Late Wisconsin valley trains, such as in the Morehouse Lowland, the relict channels can be detected but not mapped in detail because of a veneering of the surfaces with backswamp deposits derived from the Holocene meander belts.

It has long been recognized that valley trains are the result of the deposition of coarse-grained glacial outwash by stream carrying copious quantities of meltwater from receding continental glaciers. Until several years ago, however, little thought was given to the precise mechanisms involved. This writer became curious about the reason(s) for the formation and unusual size and preservation of the braided channel pattern on the Malden Plain, a situation that easily could be interpreted as evidence of an extremely brief depositional event and sudden stream abandonment or avulsion. The question has not been resolved, but evidence suggests that the braided patterns may have formed during episodes of near-catastrophic flooding caused by the draining of ice-margin lakes in the upper Midwest (Saucier, in press). It has become apparent that outburst floods of enormous magnitude through the Lower Mississippi Valley and into the Gulf of Mexico may have been a rather common occurrence during periods of glacial recession. Teller (1990) has estimated that during late-glacial time when the valley trains were forming, the *average* discharge through the Lower Mississippi Valley exceeded that of present by a factor of five. During brief outburst-flood events, the discharge would have greatly exceeded even that high level.

Terrace levels. Surfaces in each of the major valley train areas--Macon Ridge, the Western Lowlands, and the St. Francis Basin--exhibit as many as five well-defined terrace-like levels that are vertically separated by 10 ft or more. In Plates 4 through 10, the levels are numerically designated for identification purposes, with level 1 being the lowest and youngest. Readers are cautioned that the relative numbering sequences are unique to the major basins or outcrop areas and do not imply chronologic or stratigraphic equivalents. For example, there may be no relationship between level 3 on Macon Ridge and level 3 in the Western Lowlands, even though both valley trains are of the same approximate age.

Most of the valley train levels are separated by relatively straight to slightly sinuous scarps that were caused by the truncation of earlier deposits by the lateral migration of braided channels. However, the reason for the vertical separation of levels is progressive (but probably episodic) downcutting or degradation. During the time of maximum glacial outwash deposition, which probably coincided with the early to midstages of waning glaciation, the entire basin areas must have been filled to the approximate elevation of the highest valley train level. As glaciation continued to wane and the ice margin retreated northward, it is hypothesized that downcutting of the braided outwash plains resulted from an increase in stream competence. The increase would have been caused by a decrease in the sediment load (outwash) relative to the amount of meltwater. There can be little doubt that the major episodes of downcutting are directly correlated with the recession of major glacial lobes, but temporal correlations currently are impossible due to a lack of chronologic control in the alluvial valley area.

Valley train surfaces in all parts of the alluvial valley are underlain by massive amounts of coarse-grained outwash, the vertical extent of which is presently impossible to determine. In some cases, the outwash of a particular stage (e.g., the Early Wisconsin) may extend to the base of the alluvial sequence (the suballuvial surface), whereas in other cases, the outwash may only be a thin layer overlying lithologically similar outwash from an earlier glacial stage. Significant volumes of outwash have been removed only beneath the Holocene Mississippi River meander belts where the deposits have been reworked to depths averaging about 100 ft (the average depth of channel scouring in the meandering river). Otherwise, outwash underlies all areas mapped as backswamp on Plates 6 through 14. Throughout most of the alluvial valley area, the backswamp deposits are tens of feet thick, but they are much thinner in the south-central portion of the Yazoo Basin (Plates 8 and 9). In that area, backswamp is a thin, progressively southward thickening veneer of Holocene overbank deposits that overlies Late Wisconsin-stage outwash. The actual division between the valley train and the backswamp environments shown in Plate 8 is arbitrary and based on an approximation of where the backswamp deposits are sufficiently thick to obscure the braided channel pattern on the valley train surface.

Meander belts

A river will develop a meandering regime and a characteristic sinuous pattern when it has a relatively low gradient, a high suspended load/bed load ratio, is bordered by cohesive bank materials, has a relatively steady discharge from year to year, and is responding to a relatively constant base level. The Mississippi River has met all of these requirements when it has not been required to carry pulses of meltwater and coarse-grained glacial outwash. Although a large meandering river must be considered a dynamic system, the change is more orderly and predictable than in a braided system.

As a meandering river shifts laterally over time, it establishes a complex zone in which sediments are laid down in a series of active and abandoned channel environments and proximal (nearby) overbank environments. Since sedimentation rates are highest near the active river channel, the net result is a *meander belt*, an alluvial ridge that develops to an elevation higher than the more distant floodplain. Fisk (1947) developed a simple illustration of a meander belt ridge that has become a classic reference (Figure 12A). From this illustration, it can be easily envisioned how meander belt ridges have created the numerous small lowlands or depressions that constitute the major subdivisions of the basin areas (Plate 1). As pointed out by Russell (1957), once a meander belt ridge forms, most local drainage thereafter is directed away from the river channel into the lowland areas rather than into the channel. Only major rivers such as the Arkansas or Red are able to breach the ridge and enter the Mississippi River as tributaries.

The Holocene floodplain of the Mississippi alluvial valley contains the meander belt of the present course of the river and up to five abandoned meander belts that were created at various times in the past because of avulsions. North of the latitude of Vicksburg, the present meander belt is the largest, having a width up to 20 times that of the river channel itself. Before a discussion of interesting regional variations and possible causal factors in meander belt morphology and geometry, individual depositional environments are described below.

Natural levees. Natural levees are the most conspicuous landforms of meander belts and the primary reason for their topographic prominence. Further, natural levees are without doubt the most significant landforms of the alluvial valley and deltaic plain areas from both geological and cultural points of view. In both regions, natural levees overwhelmingly have been the dominant factors in the patterns of human settlement in both prehistoric and historic times. Their distribution has strongly influenced the locations of settlements, transportation routes, agriculture, and industrial development.

In geomorphic terms, a natural levee is a low, broad ridge that slopes gently away from the parent channel to the level of the adjacent floodplain or backswamp (Figure 12B). It results from the deposition of the relatively coarse (silts and sands) fraction of a stream's suspended load as floodwaters overtop the streambanks. Relatively coarser sediments and the largest volumes of sediment are deposited closest to the channel and decrease toward the floodbasin because of a decrease in the velocity and turbulence in the overbank flow. The latter are strongly influenced by the vegetative cover. Overbank flow may be in the form of either sheet flow or locally channelized. Where channelization becomes well developed, a crevasse splay may result (see the following discussion).

Natural levees develop incrementally, and consequently they increase in both height and width as a function of age. At any given point along a river, the levees are relatively higher on the cutbank (concave) side of a bend where the river is cutting into older deposits. On the point bar (convex) side, preexisting levees have been recently destroyed by lateral channel migration, and new natural levees are just beginning to develop (Figure 13).

The height and width of natural levees are a direct function of the size and volume of the suspended sediment in the parent channel. Where sediments are relatively coarse (silts and sands), the levees tend to be relatively high but narrow (hence steeper). Conversely, where the sediments are primarily silts and clays, the levees are lower and broader. Along the Mississippi River, natural levee crests average about 15 ft above the level of the adjacent floodbasin, and throughout most of the alluvial valley area, they average 2 to 3 mi wide. However, because of the progressive downvalley decrease in the average particle size of the suspended load and the relative young age of some of the channels, natural levees significantly decrease in size below Baton Rouge in the deltaic plain. Along the lower reaches of the present river and along

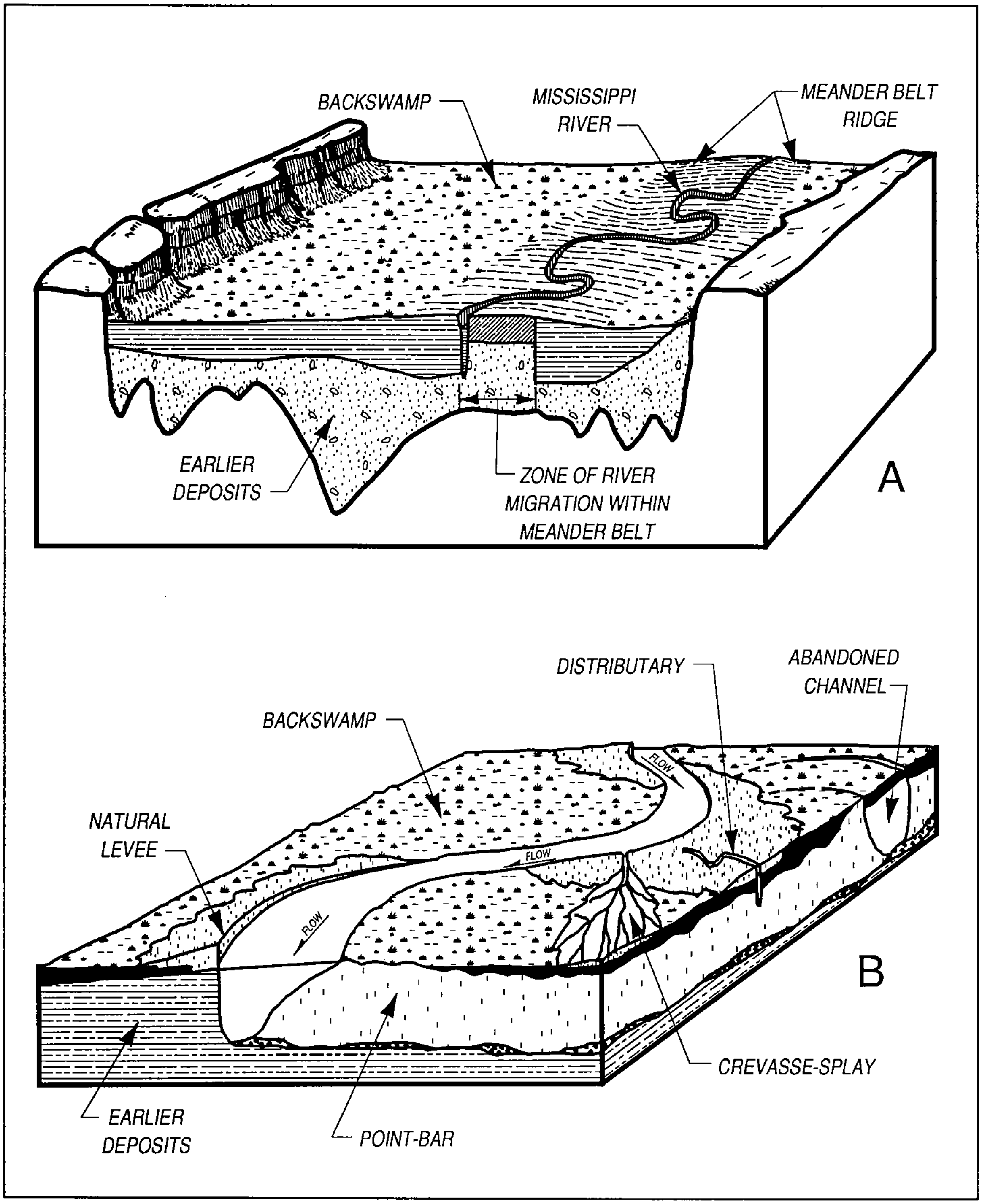

Figure 12. A typical meander belt ridge (A) within the context of the Mississippi alluvial valley (modified from Fisk 1947), and an illustration (B) of the principal depositional environments involved (modified from Allen 1964)

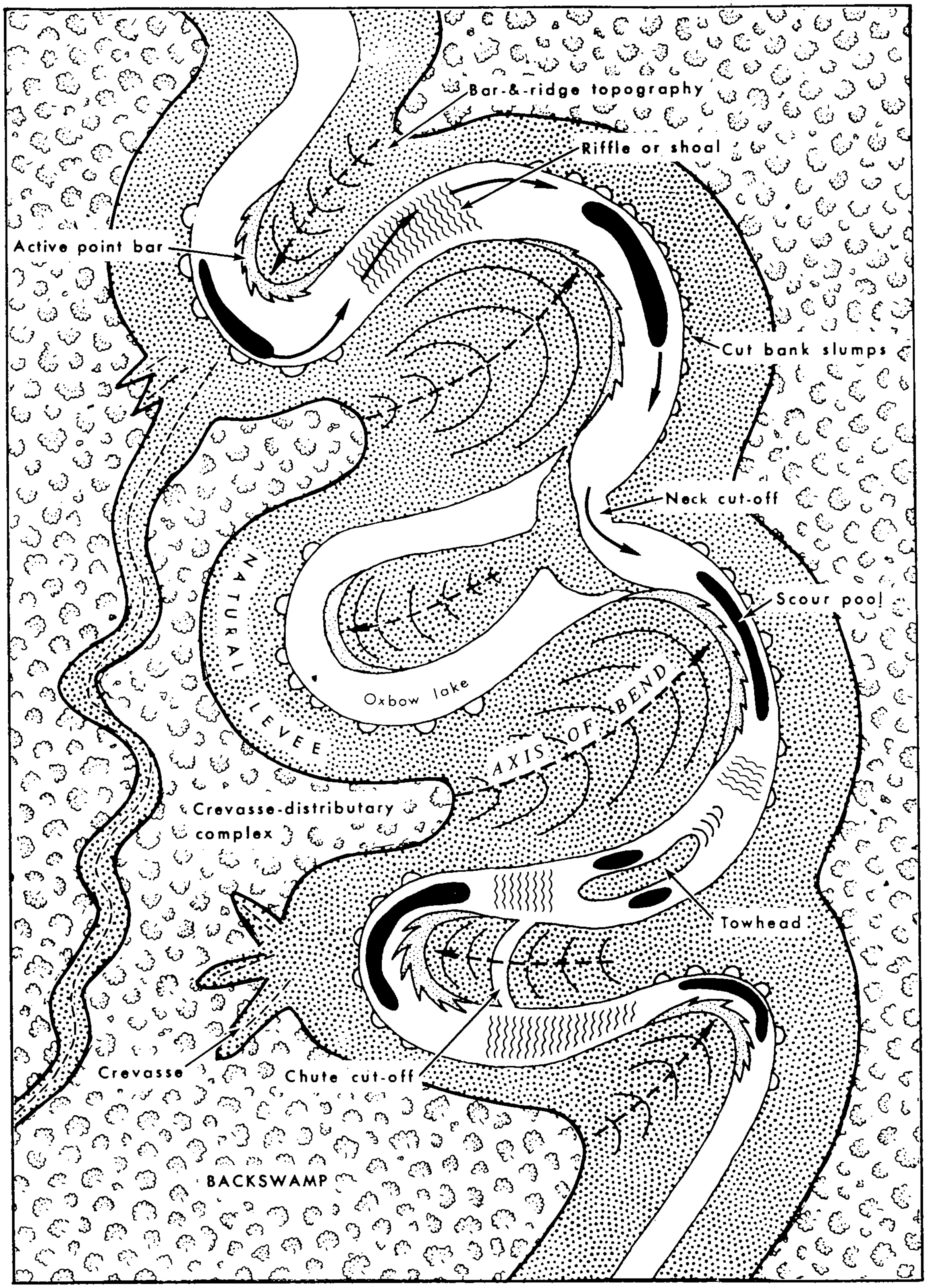

Figure 13. Primary and secondary depositional environments and related channel features of a typical meander belt (from Gagliano and van Beek 1970)

abandoned distributaries, discernible natural levees may only be a few feet high and only several hundred feet wide.

In this synthesis, natural levees are delineated as a depositional environment only in the deltaic plain area (Plates 12 to 14). Natural levees in this area are usually clearly distinguishable from adjacent environments by topography, soils, and vegetation (Figure 14A). However, in the alluvial valley area, meander belts typically include complex spatial relationships of several generations of abandoned channels, each with their own natural levees. The resultant pattern of levees thus far has not been delineated except in large scale mapping. In much of the alluvial valley area, natural levees exist more as discontinuous sheet-like deposits of locally highly variable thickness and geometry. Point bar accretion areas normally exhibit some degree of levee development and are entirely absent only over some (but not all) abandoned channels and along fresh point bar accretion along the active river channel.

From the standpoint of surface morphology, many natural levees are characterized by narrow, shallow, linear crevasse channels that are an integral element in levee formation. Typically, these ephemeral features occur in parallel series that trend perpendicular to the parent channel and extend from the levee crest to the distal margin and sometimes beyond into the floodbasin (Figure 14B).

Crevasse splays. The lateral coalescing of numerous small crevasse deposits is a normal process in natural levee formation; however, occasionally an unusually large or persistent crevasse (or series of closely spaced crevasses) will form and create a distinctive feature called a crevasse splay (Figure 12B). This feature is a discrete minidelta or thin lobe of sediment deposited on the distal side of a natural levee adjacent to where the levee crest has been breached by a crevasse. Typically, the splay extends into the floodbasin area beyond the general trend of the levee margin and hence, over a period of time, is a means by which levees increase in width (Figure 15A).

A crevasse splay may form during a single, severe flood event, but more often it forms and grows incrementally during a series of floods extending over a period of years. In any event, however, a splay is a short-lived phenomenon since crevasse channels naturally tend to fill and become inactive after a short period of time. Because a splay becomes a slight topographic feature, floodwaters from subsequent crevasses tend to avoid rather than reoccupy the splay. Crevasse splay deposits tend to be distinctively coarser than the average natural levee deposits because channelized floodwaters have a higher sediment transport capacity than those occurring as sheet flow.

Crevasse splays generally have been underestimated and underemphasized by sedimentologists, geomorphologists, and engineers and are a more important depositional environment than generally recognized. Unfortunately, their rather small size (at most a few square miles in extent) prevents their separate delineation on Plates 4 through 14. However, larger crevasse splays are portrayed on the quadrangle-scale maps (e.g., Saucier 1969) from which the plates

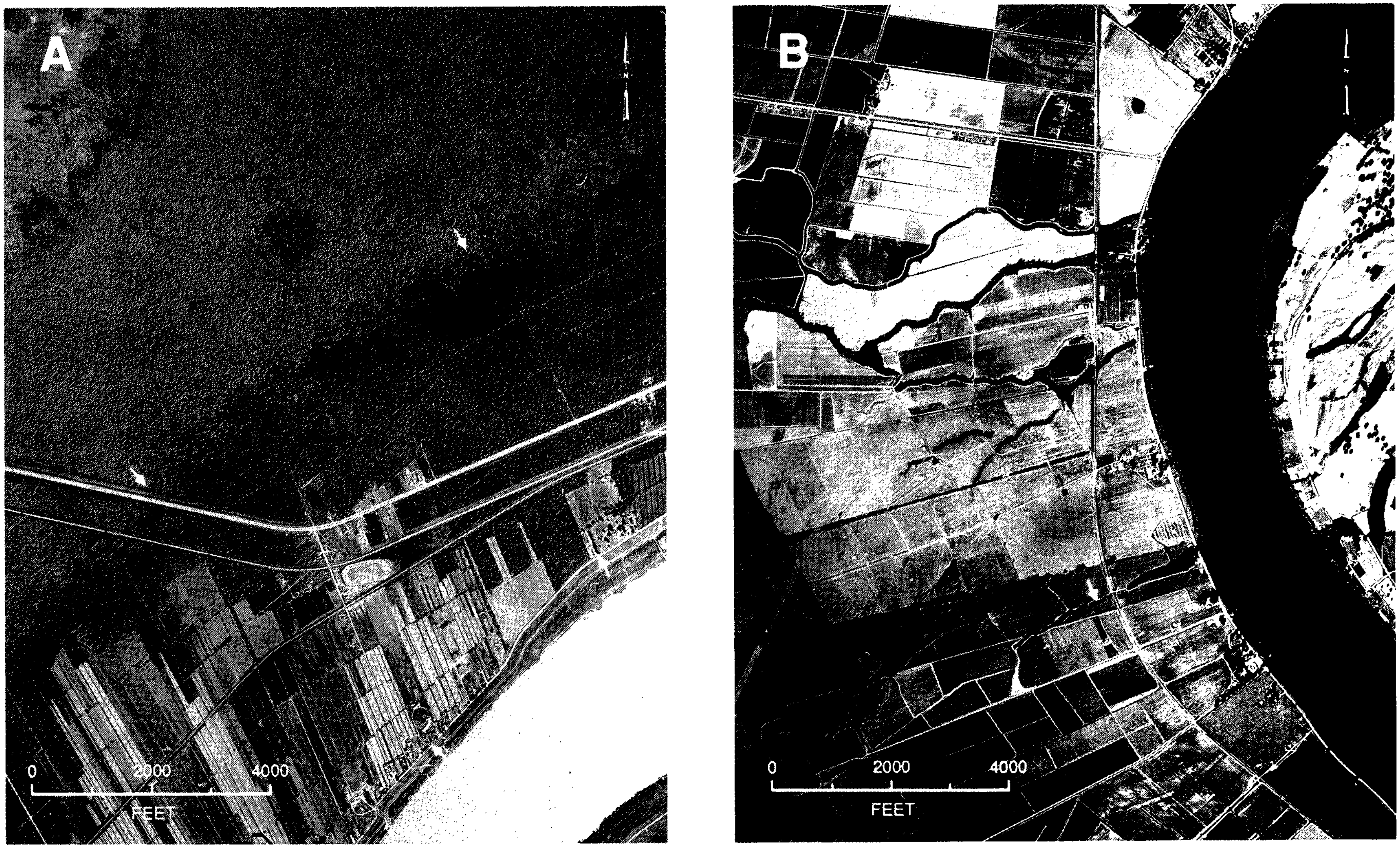

Figure 14. Natural levees. A: Natural levee, clearly evident by distribution of cultivated fields and hardwood forest vegetation (between arrows), along the Mississippi River near Gramercy in St. James Ph., LA (Plate 12); B: Crevasse channels (marked by arrows) on natural levee flanking Lake Providence, an abandoned channel in East Carroll Ph., LA (Plate 9)

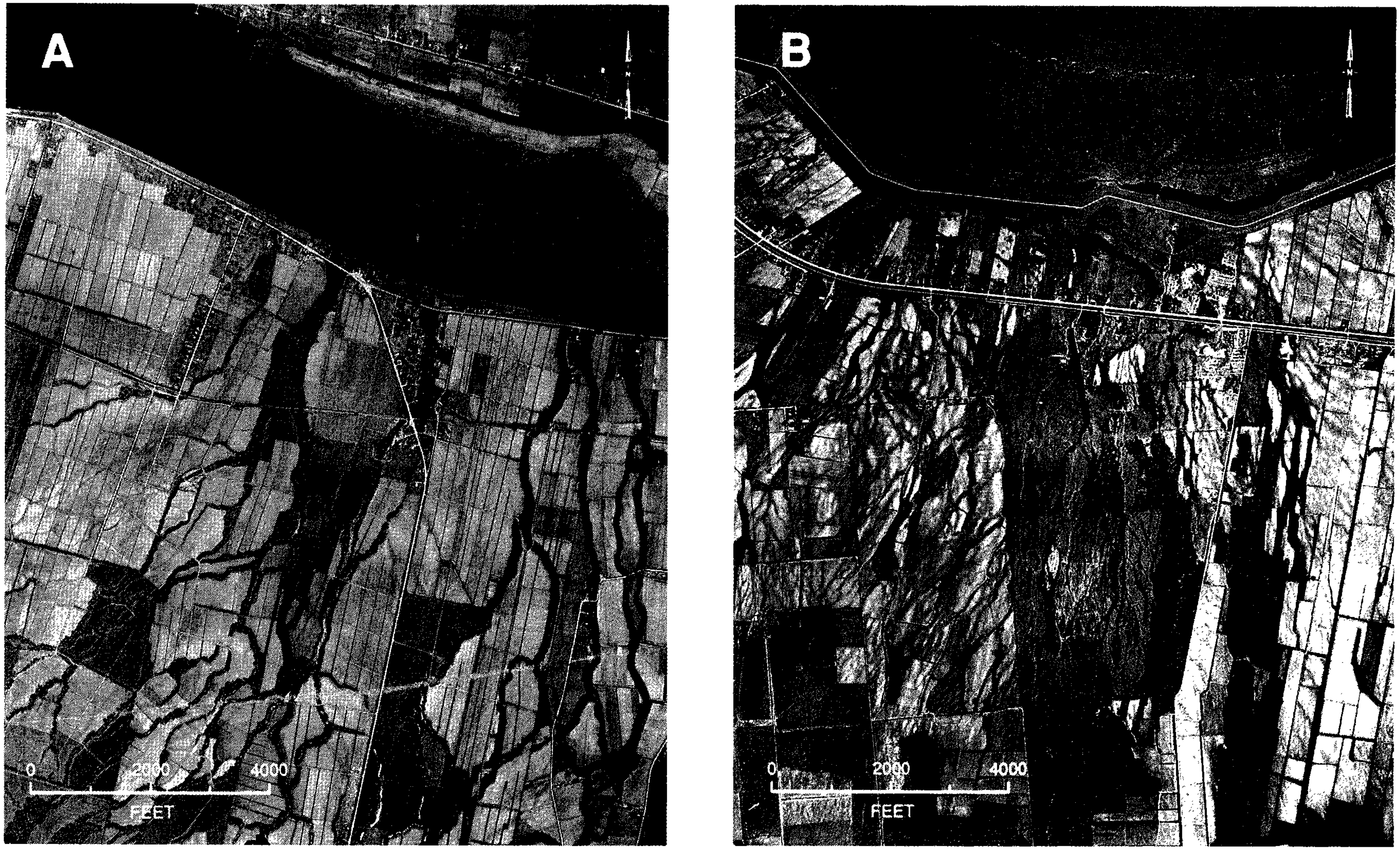

Figure 15. Crevasse splays. A: Anastomosing channels from a large, natural crevasse that formed along the south side of False River in Pointe Coupee Ph., LA (Plate 11) when it was the active course of the river; B: Crevasse splay resulting from a failure of an artificial levee (Morganza Crevasse of April 1890) near Morganza in Pointe Coupee Ph., LA

were derived and a few have been the subject of special studies (Farrell 1987). Perhaps the most significant investigation, which combined elements of geomorphology and archeology, has been that by Britsch and Dunbar (1990) of the Davis Pond area, an unusually large and complex splay located near Luling in southeastern Louisiana (Plate 14).

Crevasse splays that form under natural conditions are similar to but should not be confused with those of much larger extent that formed as a result of crevasses or major breaks in artificial flood control levees. Because of the near-catastrophic release of water that may be 10 ft or more above the level of the natural levee, splays formed under these conditions are coarser textured and more irregular in topography (hummocky terrain) and shape (Figure 15B). They typically lack a well-defined, feeder crevasse channel and instead lie immediately adjacent to a deep scour pool often called a "blue hole" that forms at the point of levee break.

Distributaries. The division of a river's flow into multiple distributary channels, several of which may be active at the same time, is a common behavior in the deltaic plain. Deltaic distributaries have been extensively studied and are the subject of a voluminous literature; however, there is essentially no discussion of the occurrence and characteristics of distributaries in the alluvial valley area. This is a major void in the literature of both alluvial morphology and sedimentology since detailed mapping has revealed them to be both numerous and extensive.

A distributary is a depositional environment/landform that is intermediate in scale and life span between a crevasse splay and a meander belt. It develops when flood flows through a crevasse are of sufficient volume or persistence to scour a channel that eventually is capable of also carrying flows at moderate to low stages. While there is no definitive information on the actual chronology of development of typical Mississippi River alluvial valley distributaries, their size and extent suggest that the larger ones persisted over periods of time measured in terms of centuries. A fundamental characteristic, however, is that none of them ever functioned to carry more than a relatively small percentage of the stream's total discharge and consequently never seriously threatened to affect a stream diversion. Notwithstanding, distributaries occur in a wide range of sizes and morphologies, indicating variable lengths of occupation/activity.

Distributaries occur in all parts of the alluvial valley and formed along all of the major river systems, but those of the Mississippi River differ in several respects from those of other streams. One major cluster of Mississippi River distributaries occurs in the lower portion of the St. Francis Basin where they are occupied in part by the St. Francis River and the Left Hand Chute of Little River (Plate 6). A second major cluster occurs in the Yazoo Basin where five are of sufficient size and importance to be designated as major physiographic features (Plate 1 and 7 through 9). Along each of these distributaries, there is a well-developed natural levee ridge, and a single, relict meandering channel tens of miles in length is clearly evident (Figure 16A). Flow through the

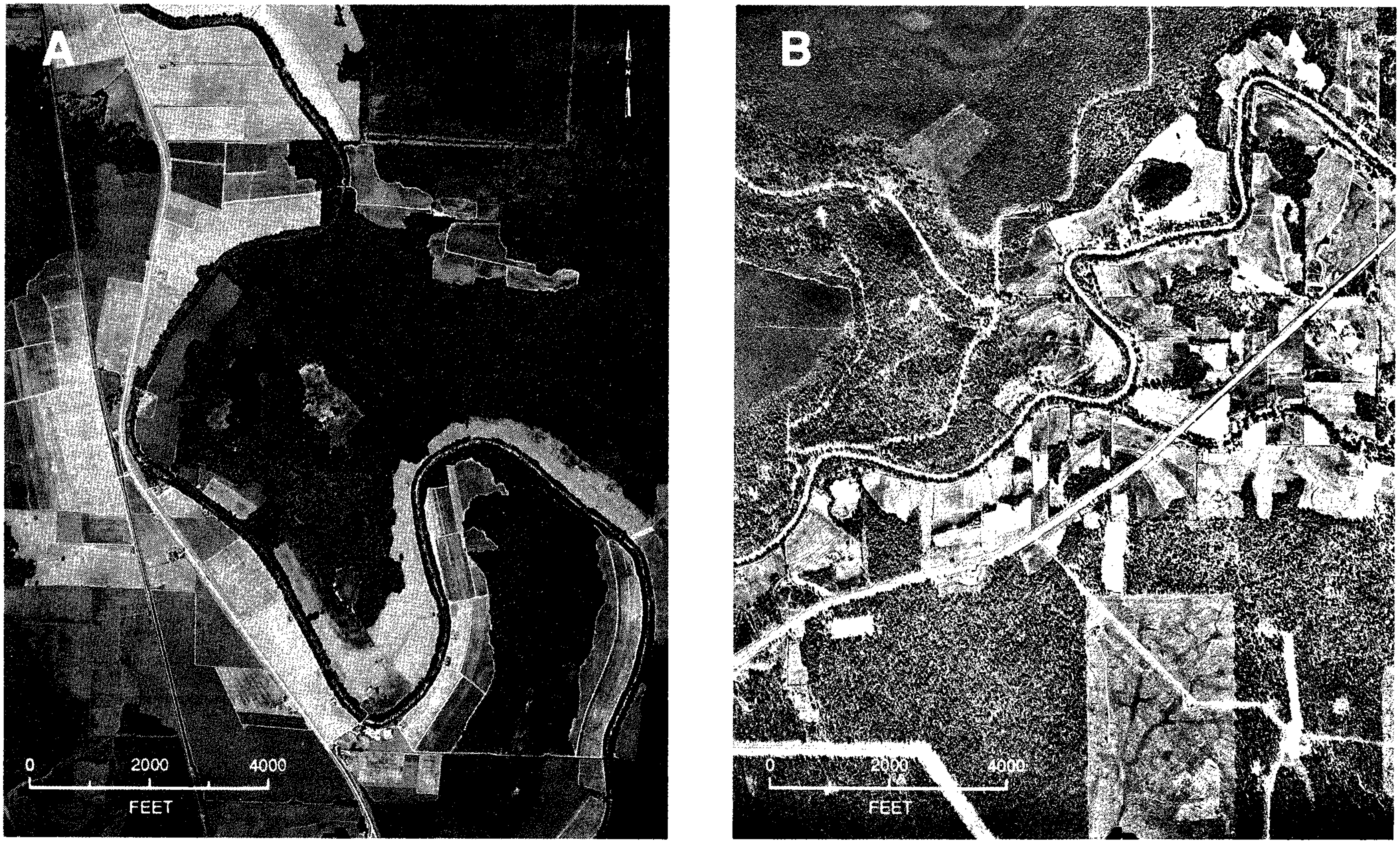

Figure 16. Distributaries. A: Abandoned channel and natural levees (cultivated areas) of the Deer Creek distributary near Valley Park in Issaquena Co., MS (Plate 9); B: Main channel and branches (arrows) of the Little River distributary near Walters in Catahoula Ph., LA (Plate 10)

distributaries was of such duration that an appreciable number of cutoffs took place, leaving easily recognizable abandoned channels or oxbow lakes.

All of these distributaries can be traced to their points of origin which are abandoned channels of the Mississippi River; however, their precise relationship to the active river at the time of activity is ambiguous. This writer believes that the distributaries originated as crevasses along concave banks of the active river channel and that flow persisted until the bends on which they were located were abandoned by the river. Abandonment and the accompanying sedimentation in the channels may have been the process that marked the end of flow in the distributaries and prevented what might have eventually been river avulsions and new meander belt formation. The possibility cannot be completely dismissed, however, that the crevasses formed along what were already abandoned channels that were interconnected with the active river channel.

During early stages of development, the distributary channels must have lengthened rapidly, following slight gradient advantages and by occupying existing small streams and water bodies on the floodplain. Distributary development was not always into backswamp areas, since it is apparent that some developed across topographically low valley train areas and others occupied relict channels in older meander belts. In essence, the development probably consisted of little more than the progressive confinement of floodwaters by natural levees and accompanying channelization.

In the southern part of the alluvial valley, the Little River is a major crevasse system that originated from an abandoned course rather than an abandoned channel. It originated from the Walnut Bayou Meander Belt (M2C, Plate 1) at Jonesville, Louisiana (Plate 10), discharging water and sediment into the Catahoula Lake Basin. For about 10 mi, the distributary occupied a single, rather steep channel; but in proximity to Catahoula Lake, it formed a series of branches, including Old River and French Fork, and built a small delta in the basin (Figure 16B). The main channel of the system and each of the branches has very well-developed natural levee ridges, but the system apparently was not occupied long enough for meandering to begin.

Distributaries of major tributaries such as the Arkansas, Red, and White rivers exhibit a much more branching network than those of the Mississippi River, perhaps because of differences in the size and character of their sediment loads. Figure 17 shows the pattern of development of two unusually large but otherwise typical distributaries that formed along two separate Arkansas River meander belts. In both cases, the channel networks can be traced for many tens of miles, and each branch has small but discernible natural levees that decline in size downstream. Because of their complexity, these distributaries are referred to as systems and were given proper names in a previous study. At the distal ends of the networks, the levees are only a few hundred feet wide and only a few feet higher than the floodbasin elevation. However, even natural levee ridges this small apparently were influential in prehistoric settlement patterns (Kidder and Saucier 1991).

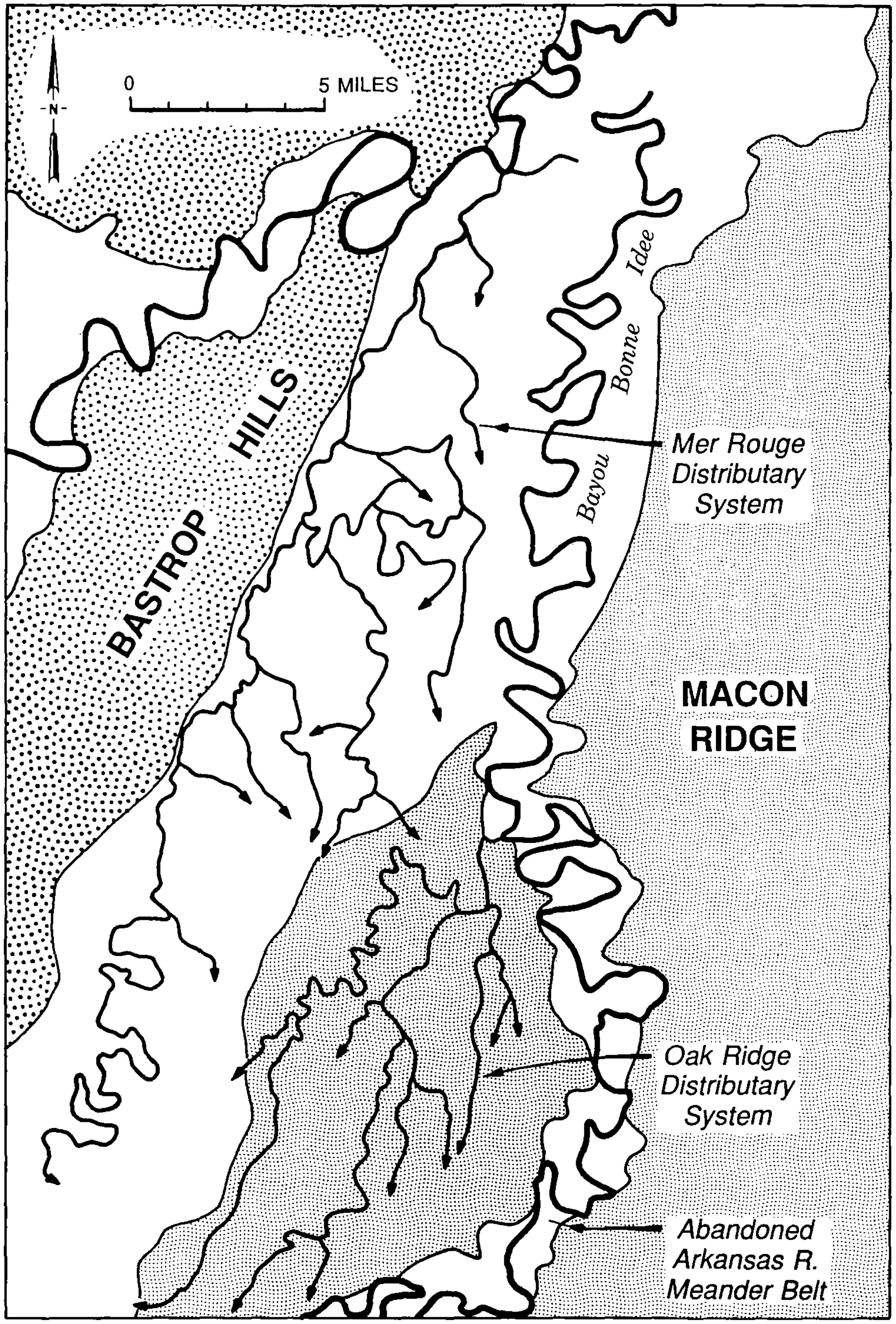

Figure 17. Two large distributary systems originating from crevasses along Arkansas River meander belts northeast of Monroe in northeastern Louisiana (from Kidder and Saucier 1991) (meander belt details and traces of older meander belts omitted to avoid confusion)

Point bars. No depositional environment of a meandering stream has been more extensively investigated by sedimentologists than the point bar. In simple terms, a point bar consists of relatively coarse-grained deposits (mostly silts and sands) that are laid down during higher stream stages in a zone of relatively low turbulence and velocity along the convex side of a migrating stream bend (Hickin 1974). Bar development is a means by which a meandering stream strives toward equilibrium by compensating for channel widening caused by bank caving. Point bars would not fully develop without appreciable stage variations on the stream (e.g., annual floods) and easily erodible banks. Each major high-stage event is accompanied by a new increment of bar development from the stream's bed load, much of which may have come from material eroded from the cutbank side of the river immediately upstream from the bar.

In more complex terms, a point bar is a composite of sediments that are transported as underwater dunes in the stream channel. Because of helical flow, the sediments are moved into shallower water and deposited as transverse bars and sand waves (Figure 18A). The bars and waves typically begin forming just below reaches (straight channel segments between bends) and progressively develop downstream around the convex bends as arcuate ridges. Before a ridge develops completely around a convex bend, one or more new ridges are beginning to form near the head of the bar and accrete in a downstream direction; hence, bar formation is a continual process.

Rates of point bar formation differ greatly from area to area within a given river system and even more so from one river system to another. Bar formation will be most rapid during and immediately following an unusually large flood event, but local factors are more important in dictating local variations. For example, bar formation will be unusually rapid immediately downstream from a migrating bend that is encountering a zone of easily eroded deposits (e.g., an older point bar sequence), and it will be extremely rapid below a recently formed cutoff because of the temporarily increased gradient and greater scouring capacity of the river.

Such local variations in sediment supply and energy levels cause significant differences in bar size, spacing, and continuity. While it is typical for new bars to be separated from previous ones by narrow, shallow, linear swales, periodically swales of unusual width and length will form and are preserved in an accreting bar sequence. As will be discussed later, unusually large swales may become the routes of subsequent chute formation and channel cutoffs.

Cumulative point bar development results in the formation of characteristic point bar ridge and swale sequences (sometimes referred to as meander scrolls or scroll-bar sequences). These sequences faithfully record the directions of bend migration and produce the highly diagnotic and easily recognizable landscape patterns that are the "trademark" of a meandering stream (Figure 18B). In a fully developed meander belt with a series of cutoffs, the pattern of point bar ridges and swales can be complex with numerous cases of abrupt truncations and changes in trends (Figure 19A).

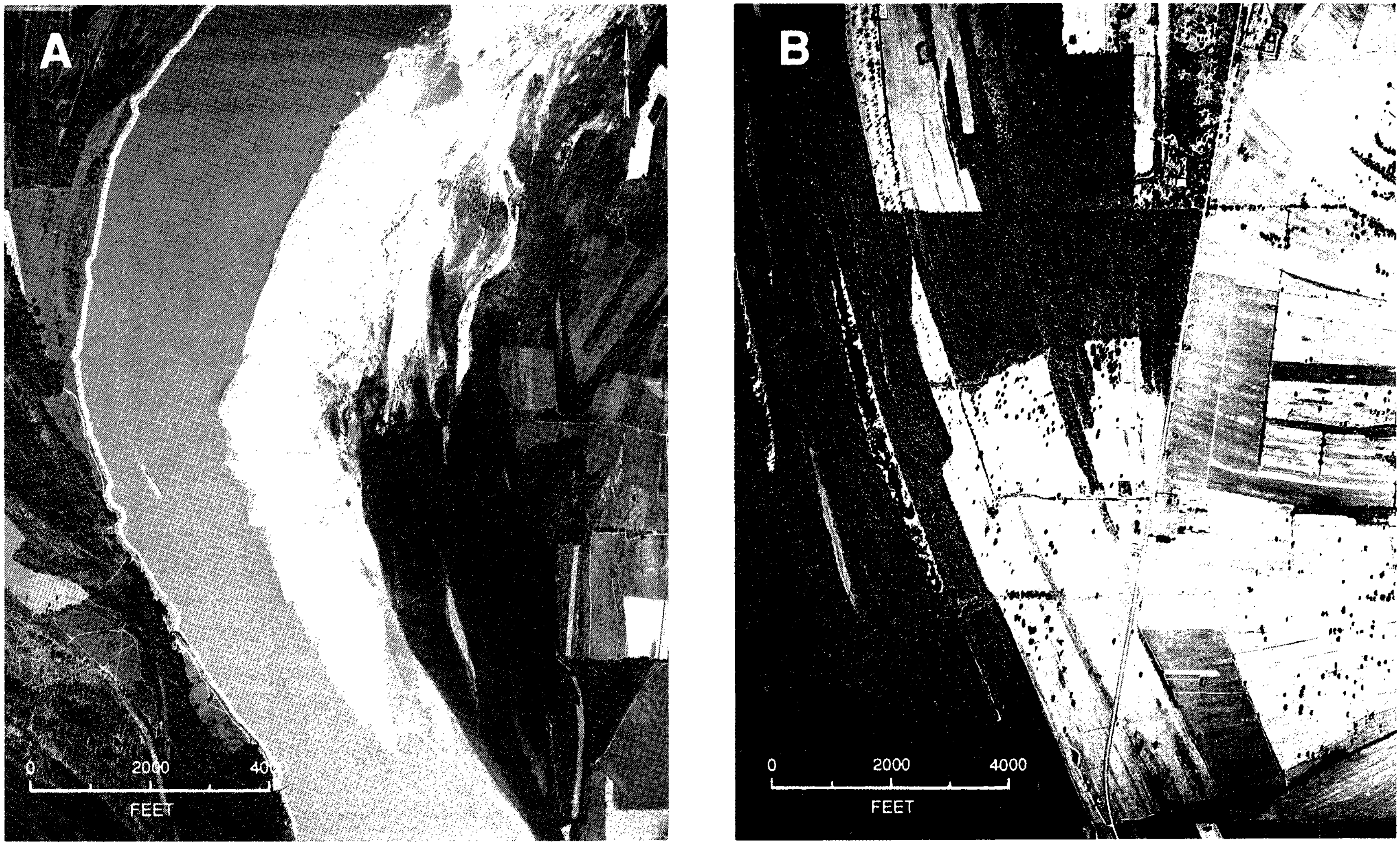

Figure 18. Point bar environment. A: Sand waves being deposited on an actively accreting point bar along the Chute of Island No. 35 bend of the Mississippi River in Tipton Co., TN (Plate 6); B: Point bar ridges and swales of one meander sequence truncating another east of the Tensas River near Ferriday in Concordia Ph., LA (Plate 10)

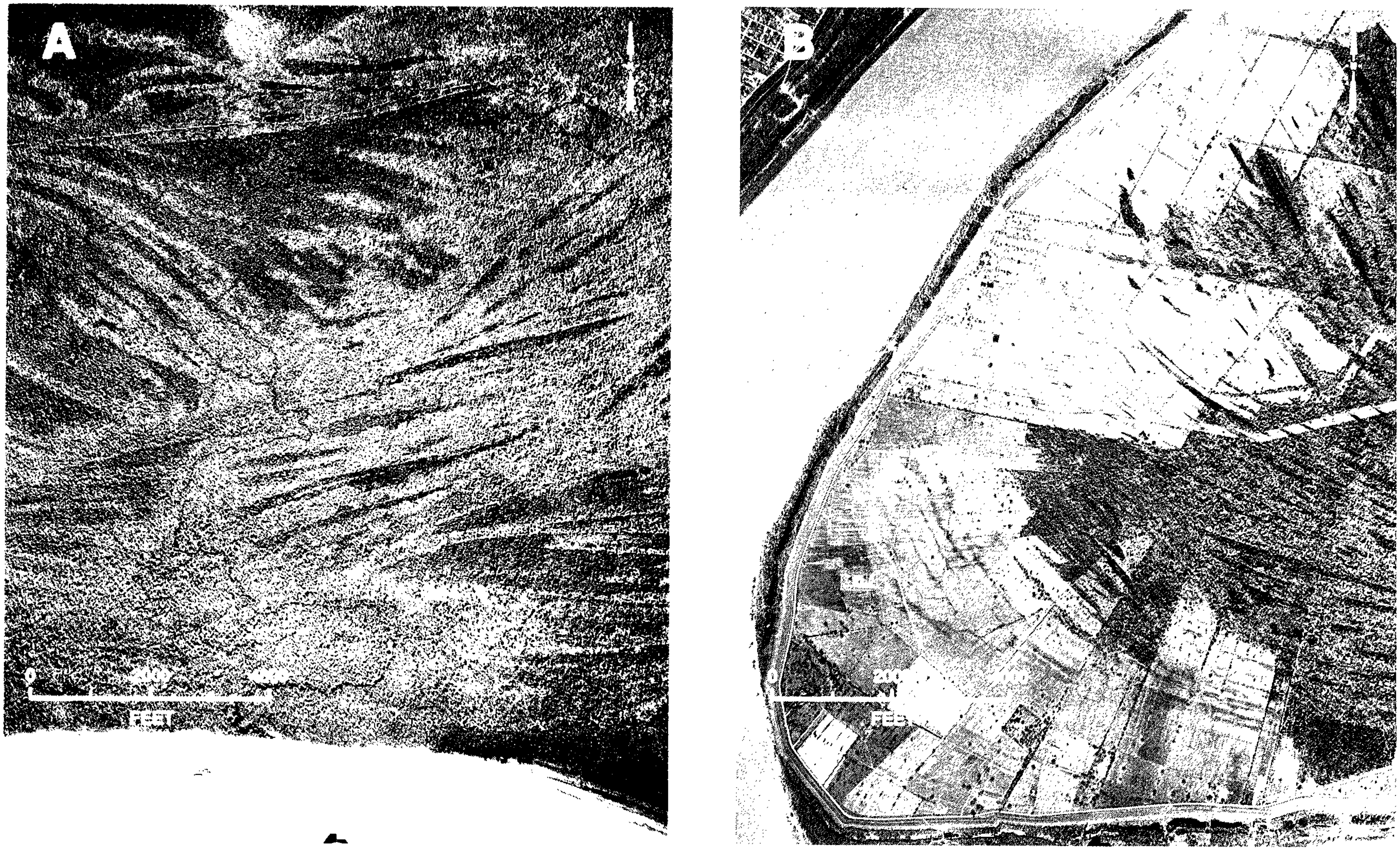

Figure 19. Point bar environment. A: Complex accretion topography indicating movements of several different Mississippi River bends west of St. Francisville in West Feliciana Ph., LA (Plate 11); B: Point bar ridges and swales veneered with natural levee deposits near the present river channel at Duncan Point south of Baton Rouge in East Baton Rouge Ph., LA (Plate 11)

A point bar sedimentary sequence is therefore primarily the product of channel processes and lateral accretion. Point bar deposits extend to the depth of greatest channel scour (the thalweg) in the migrating bends where they are separated from older underlying deposits by erosional unconformities. The upper limit of point bar development is variable and hard to define but probably approximates the bank-full stage of the parent stream. Once a stream meanders away from a given area and channel processes cease, thereafter the area receives sediment through the vertical accumulation of overbank sediments and a natural levee begins to form. Swales are initially filled and then the ridges are covered with thin veneers of overbank deposits (Figure 19B). If the process continues long enough, then fully-developed natural levees result and the pattern of point bar ridges and swales will be obscured. Normally in a sedimentary sequence, the uppermost point bar deposits are transitional with and hard to differentiate from the lowermost overlying natural levee deposits.

Point bar sequences are associated with the meander belts of all major and most minor river systems in the Lower Mississippi Valley area. Those along Mississippi River meander belts are most extensive since there is a correlation between the amount of meandering and hence the amount of point bar deposits and the discharge (size) of the stream. Volumewise, point bar deposits encompass the majority of Holocene alluvial sediments, especially in the upper and central parts of the alluvial valley. Only below the mouth of the Red River does the amount of point bar associated with the Mississippi River meander belts become noticeably and progressively smaller. This is in part due to the downstream decrease in the grain size and quantity of bed load of the river and the greater youth of the channels. However, more so it is a reflection of the greater cohesiveness (and hence decreased erodibility) of the materials comprising the bed and banks of the river.

In Plates 4 through 14 and the quadrangle-scale maps from which they were compiled, midchannel bars and islands, sometimes referred to as towheads or sandbars (Figure 13), are designated as point bar deposits. Their precise mode of origin is admittedly different from that of typical point bar ridge and swale sequences; however, the deposits are essentially the same in terms of source and composition and hence are included in this depositional environment. Moreover, it is not unusual for the islands to occasionally become incorporated into an accreting point bar sequence.

Abandoned channels. The natural levee environment may be the single most important one in the Lower Mississippi Valley area from several considerations, but the abandoned channel environment ranks a close second. In terms of engineering considerations, it may even be *the* most significant because of the typically soft and compressible soils that are often present (Fisk 1947).

In a meandering river regime, short channel segments may be abandoned in two ways as the stream constantly strives to shorten its course. If two bends migrate such that they intersect each other and the narrow neck of point bar is breached, streamflow will be short circuited and a channel segment will be abandoned. This process results in what is called a "neck cutoff." On the other hand, a "chute cutoff" occurs during a major flood when riverflow cuts across a point bar by occupying a major swale and scouring it into a major channel. A cutoff is affected if and when a majority of the flow is diverted through the chute. The process of channel shortening by chute cutoffs is relatively slow and may take tens of years to accomplish via progressive chute enlargement. In contrast, a neck cutoff is normally much more sudden and may be affected during a single major flood event. If two migrating bends intersect, the formation of a cutoff is inevitable: formation of a chute cutoff requires a favorable presence and alignment of swales and is by no means an assured event.

While no count has been made and would probably be impossible for several reasons, this writer estimates that neck cutoffs along the major streams in the alluvial valley area outnumber chute cutoffs by at least 20:1. This ratio increases in a downstream direction since chute cutoffs are relatively harder to affect where point bar deposits are finer grained and stream stage variations are smaller.

When a neck cutoff takes place, the abandoned channel segment typically undergoes a predictable life cycle that can be described in terms of a series of stages. Various models of abandoned channel evolution have been proposed (e.g., Gagliano and Howard 1984). Each stage involves the development of subenvironments, the locations of which have been important in influencing occupational patterns during both prehistoric and historic times (Weinstein 1981). It has been determined that there has been at least 431 neck cutoffs along the various Mississippi River meander belts (Smith 1989) and twice that number along streams such as the Arkansas, Red, and White rivers. The vast majority of the surviving abandoned channels fit into one of the six stages shown in Figure 20. The duration of a complete life cycle is highly variable because of local circumstances but probably involves at least several hundred years and apparently can be as long as several thousand years.

Immediately after a neck cutoff takes place, sand bars quickly form in the upper and lower arms of the abandoned stream bend and an oxbow lake is created (Figures 20B and 21A)). No river through-flow takes places, but the lake is not completely hydraulically isolated from the river. Small channels called batture channels form and maintain themselves through the sediment wedges in the arms (Figure 20C) and serve to allow overflow from the oxbow lake to enter the river at low stages and floodwaters to back up into the lake during high stages (Figure 21B). Because of this hydraulic connection, fine-grained suspended sediment (clays and silts) periodically enters and is deposited in the oxbow lake, causing it to slowly fill. As the lake shallows, the sediment wedges or plugs in the arms also expand at the expense of open water, but from deposition of clays and silts rather than sands (Figure 20D). The fine-grained channel-fill deposits constitutes what engineers call "clay plugs" (Fisk 1947) and are manifest at the surface by a flat, featureless fresh-water marsh or swamp.

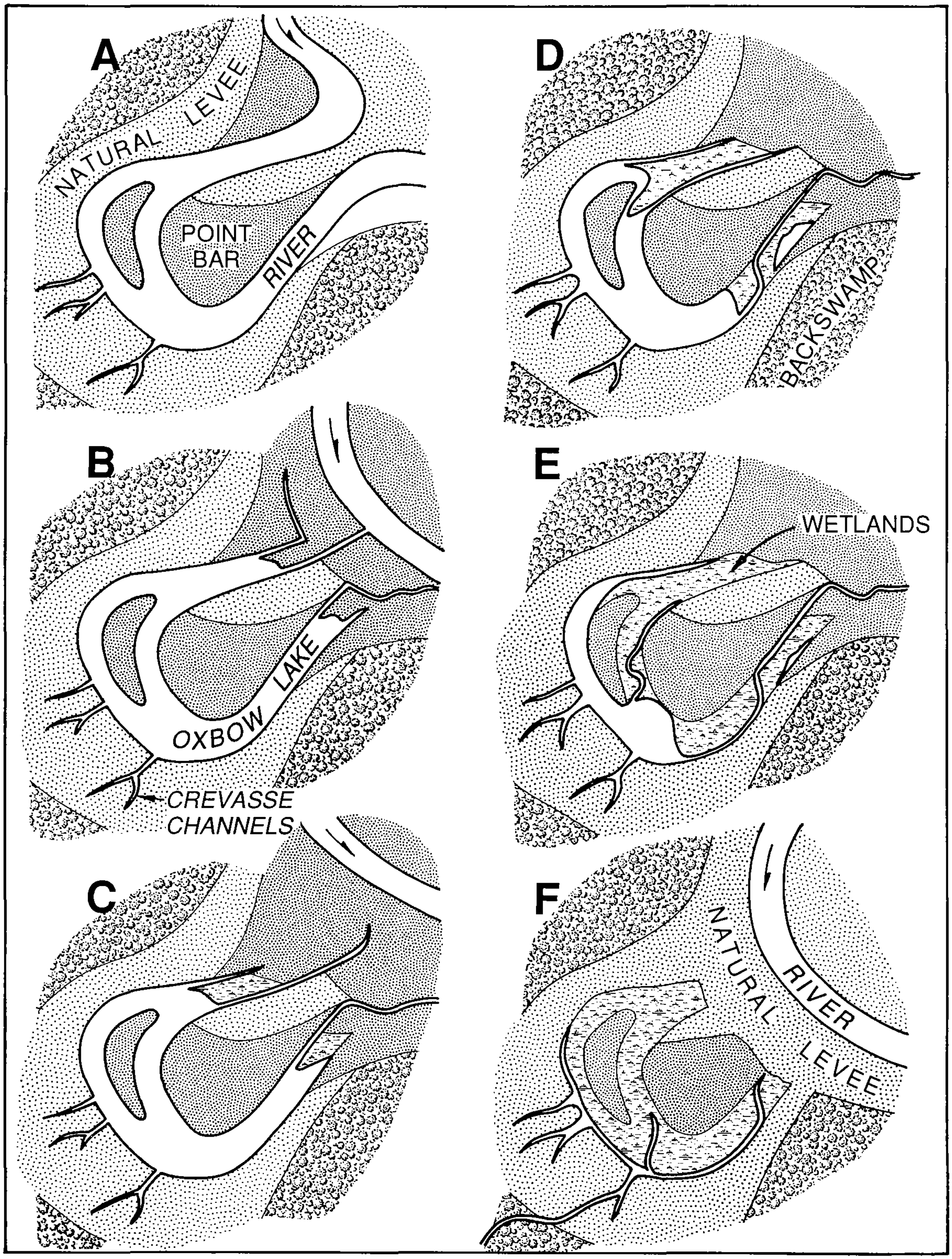

Figure 20. Characteristic stages in the life cycle of a typical neck cutoff along a major stream in the Mississippi alluvial valley

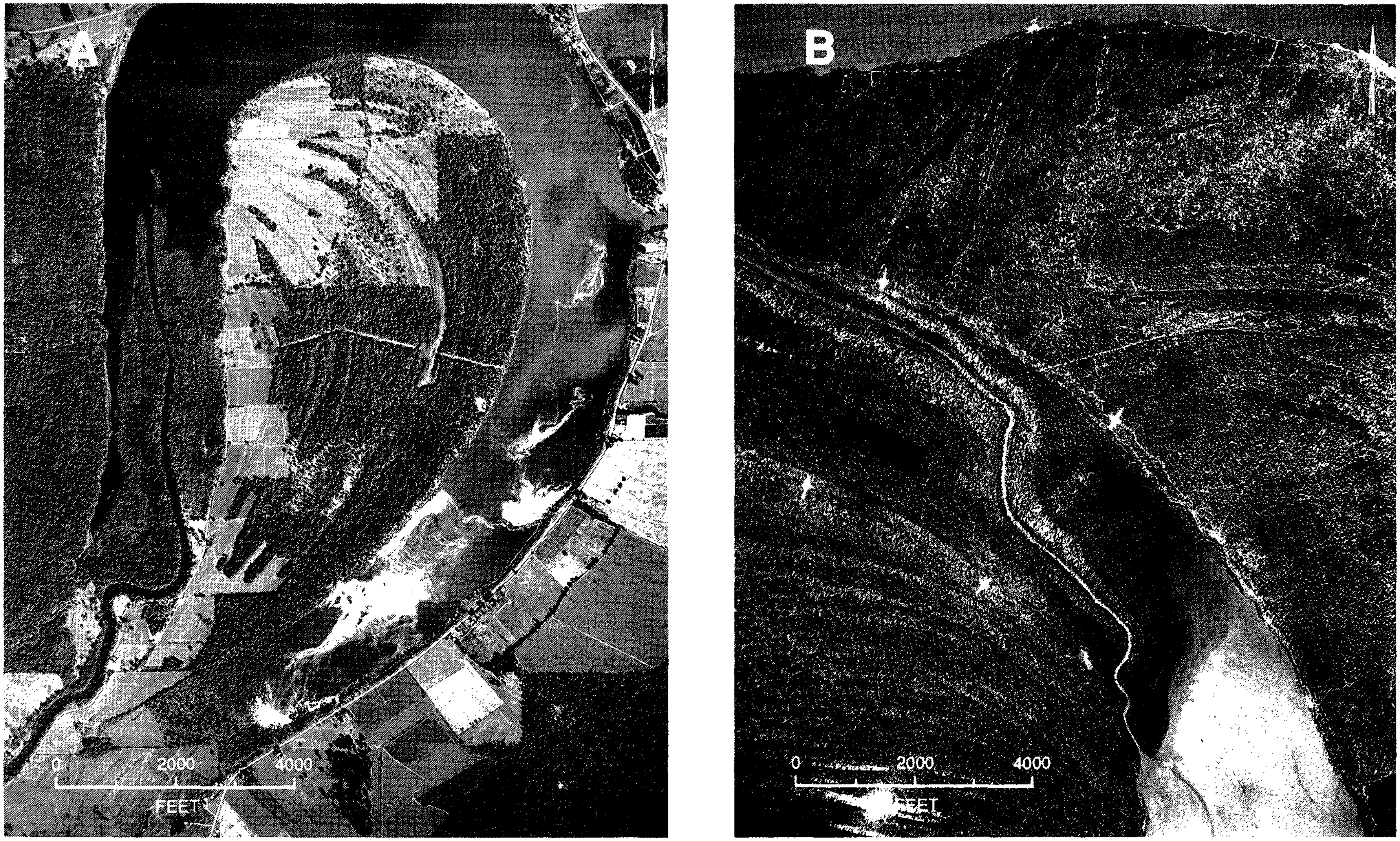

Figure 21. Abandoned channels. A: Cocodrie Lake north of Monterey in Concordia Ph., LA (Plate 10), a classic-shaped Mississippi River cutoff with oxbow lake and filling of arms below point of cutoff; B: Small batture channel connecting oxbow lake near Raccourci Old River in Pointe Coupee Ph., LA (Plate 11) with Mississippi River (channel filling located between arrows deposited between 1848, date of cutoff, and 1959, date of photo)

With continued sediment introduction into the channel, the wetland areas expand farther up the arms and the batture channels extend themselves accordingly (Figure 20E), sometimes terminating in a small lake head delta. For reasons not clearly understood, the batture channels characteristically hug the former channel bank lines and sometimes cut abruptly across the channel to the opposite side rather than occupying a midchannel position (Russell 1939) (Figure 22A). This phenomenon is significant because the positions of the batture channels often determine the courses of subsequent floodplain drainage in cases where the abandoned channels are covered and largely obscured by later sedimentation. In these instances, the presence of a buried abandoned channel often can be inferred from the distinctive pattern of local drainage.

The ultimate fate of an oxbow lake depends primarily on the behavior of the active river channel after cutoff takes place. If the river channel remains relatively nearby and there is an effective connection, the lake may fill completely and be characterized by a dense swamp forest (Figure 22B). Conversely, if the river channel meanders well away from the lake or occupies a new meander belt, the lake may persist for a long time as a relatively deep water body. More typically, however, the active river channel will eventually meander back toward the abandoned channel, constructing a natural levee over and filling and obscuring some portion of the feature (Figure 20F). Portions of the abandoned channel (or oxbow lake) not directly affected by natural levee development will persist thereafter for an unusually long period of time. That is because the natural levee ridge seals off the batture channels and forces local drainage to exit the abandoned channel via a different route, often into the floodbasin flanking the meander belt. Sediment introduction into the abandoned channel is thereby sharply reduced.

Fortunately, more than 90 percent of the abandoned channels that are portrayed in Plates 4 through 14 are conspicuous features that are identifiable in the field or on maps and aerial photos because of their characteristic drainage, topography, soils, and configuration (Figure 23A). Even when some of the diagnostic criteria are absent, the presence of an abandoned channel can often be inferred from the patterns of point bar ridges and swales and by reconstructions of the directions of channel meandering and bend evolution (Figure 23B). However, in a few cases where surface evidence is ambiguous or complicated by the behavior of small streams, the presence of all or a remnant of a buried abandoned channel has been detected by borings that encountered a large, thick mass of fine-grained sediments (a "clay plug" deposit).

Considering their small numbers and relatively variable morphology, no model has been attempted to describe the evolution of chute cutoffs. Since a much smaller segment of a bend is involved and most have a more arcuate than horseshoe shape, sediment filling is much more rapid and proportionately much more occurs as a sand bar or wedge rather than a clay plug. As would be expected, lakes generally do not form.

Abandoned courses. From a conceptual point of view, an abandoned course is easy to define, but in the Lower Mississippi Valley area, they have a

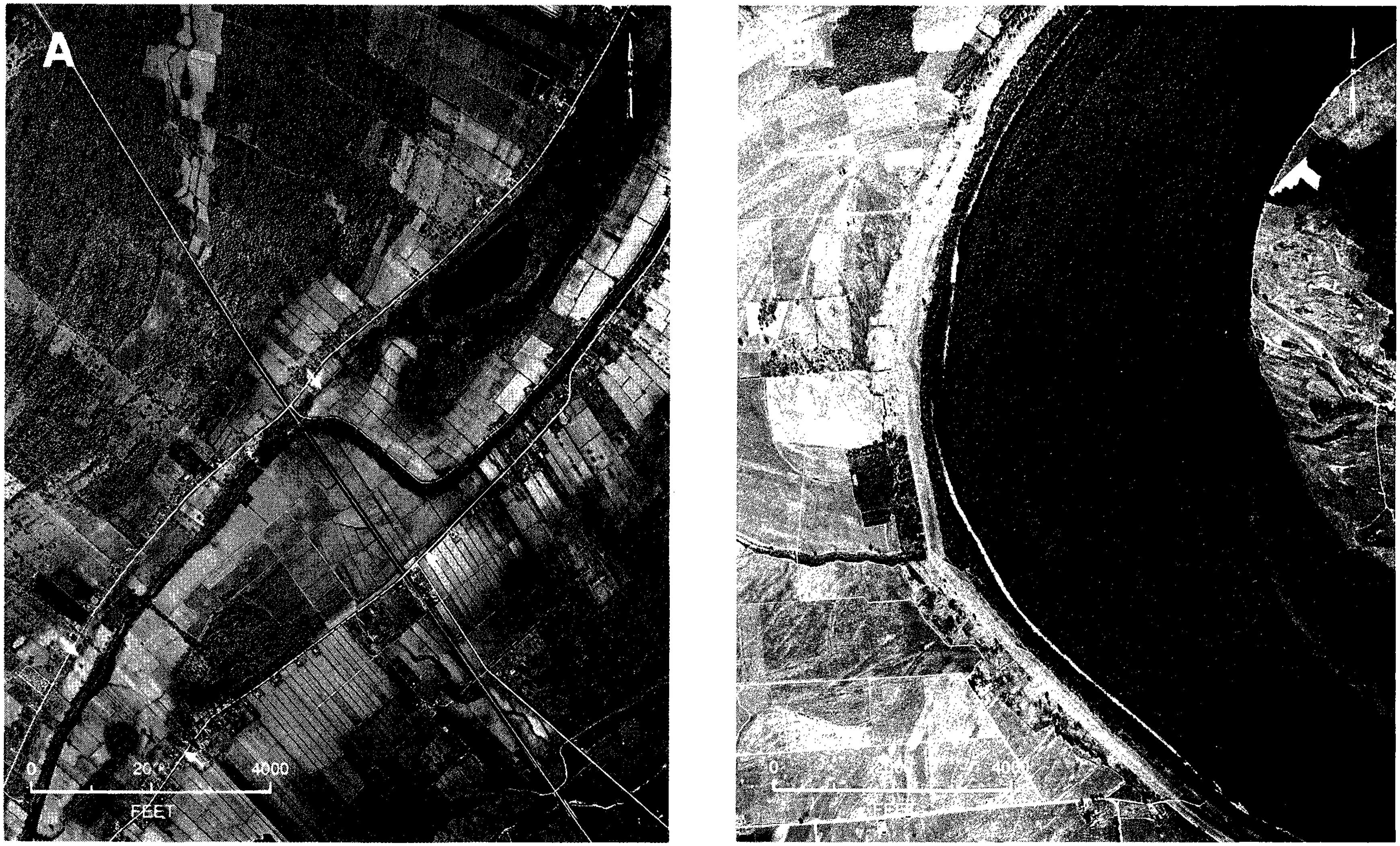

Figure 22. Abandoned channels. A: Typical batture channel pattern as exemplified by The Chenal in the lower arm of the False River cutoff in Pointe Coupee Ph., LA (Plate 11) (abandoned channel limits indicated by arrows); B: Essentially filled abandoned channel characterized by swamp forest vegetation west of Centennial Island in Crittenden Co., AR (Plate 6)

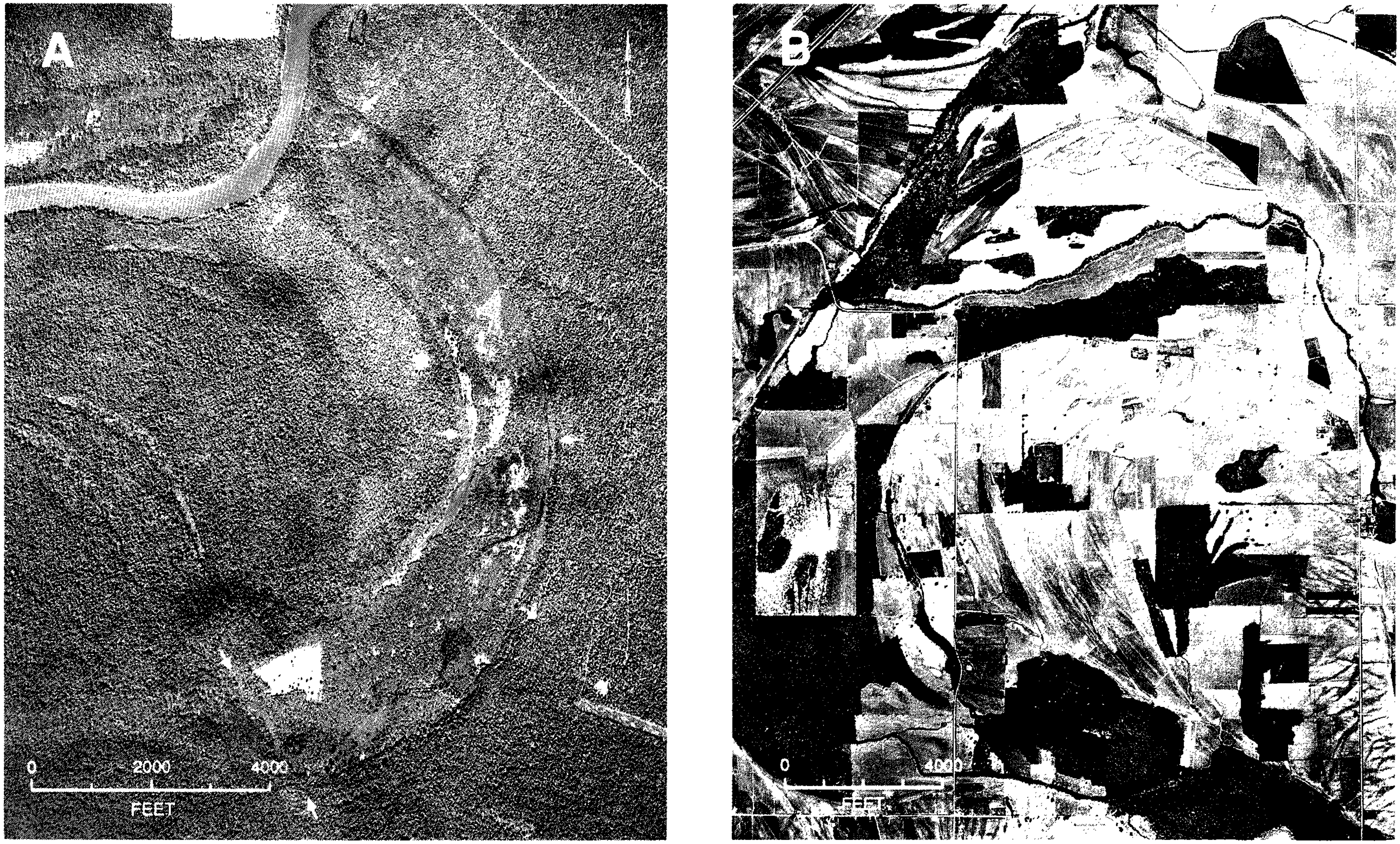

Figure 23. Abandoned channels. A: Filled abandoned channel detectable by characteristic vegetation assemblage and soils (between arrows) along the Tensas River northeast of Sicily Island in Tensas Ph., LA (Plate 10); B: Abandoned channel southwest of Hughes in Lee Co., AR (Plate 7) sufficiently filled and well drained to allow agriculture (limits located between arrows detectable by absence of point bar ridges and swales, vegetation, and darker soil tones)

highly variable morphology. No simple model explains their characteristics and composition.

An abandoned course is a lengthy segment of stream channel, more than a single bend and up to hundreds of miles long, that remains after a stream diversion (avulsion) to a new course and meander belt. The time involved in abandonment is not known but probably is measurable in terms of decades to a few centuries. During that interval, a sand wedge forms at the point of avulsion and slowly develops and thins downstream as flow progressively declines.

In the simplest case, the original course slowly narrows through lateral accretion to accommodate the reduced discharge. If the process of abandonment is rather slow, the extent of the former bank lines becomes indistinguishable, especially on convex bends, since the abandoned course fill closely resembles the adjacent point bar deposits in both morphology and lithology (Figure 24A). As flow declines, the underfit stream becomes progressively narrower but maintains the sinuosity of the original full flow channel (Figure 24B). In a more typical scenario, however, there is sufficient discharge and sediment load for the underfit stream to continue to meander as the discharge declines. When this occurs, the location and configuration of the original course during the last stages of full flow cannot be determined since the meanders of the underfit channel may greatly exceed the former limits. Point bar sequences of the underfit stream can usually be differentiated from those of the full-flow stream by the radii of curvature and spacing of the ridges and swales.

In still another case, a course is apparently abandoned rather rapidly and the sand wedge is relatively short, leaving a considerable length of stream channel as an open, slackwater stream carrying only local runoff. Geomorphic mapping suggests that when this occurred in the alluvial valley area, the relict course was eventually occupied in whole or in part by either a smaller river system or by a major distributary. As described above, in either event the net result was that the smaller system continued to meander.

Examples of the occupation of Mississippi River abandoned courses are numerous and include the St. Francis River in the St. Francis meander belt (M3A, Plate 1; Plate 6), the Yazoo River in the Yazoo meander belt (M2B, Plate 1; Plate 8), and several separate courses of the Red River in the Teche meander belt (M3D, Plate 1; Plate 11). In each case, the original course was significantly modified. In the case of the Arkansas and Red rivers, various segments of abandoned courses have been occupied by smaller streams, but the courses have not had the capacity to meander and are confined by the original channel banks.

When limitations in the knowledge of the geomorphic history of the Lower Mississippi Valley area and uncertainties in such determinations are recognized, there appears to be only one instance where a major river has reoccupied one of its former courses. This occurred where the Teche course

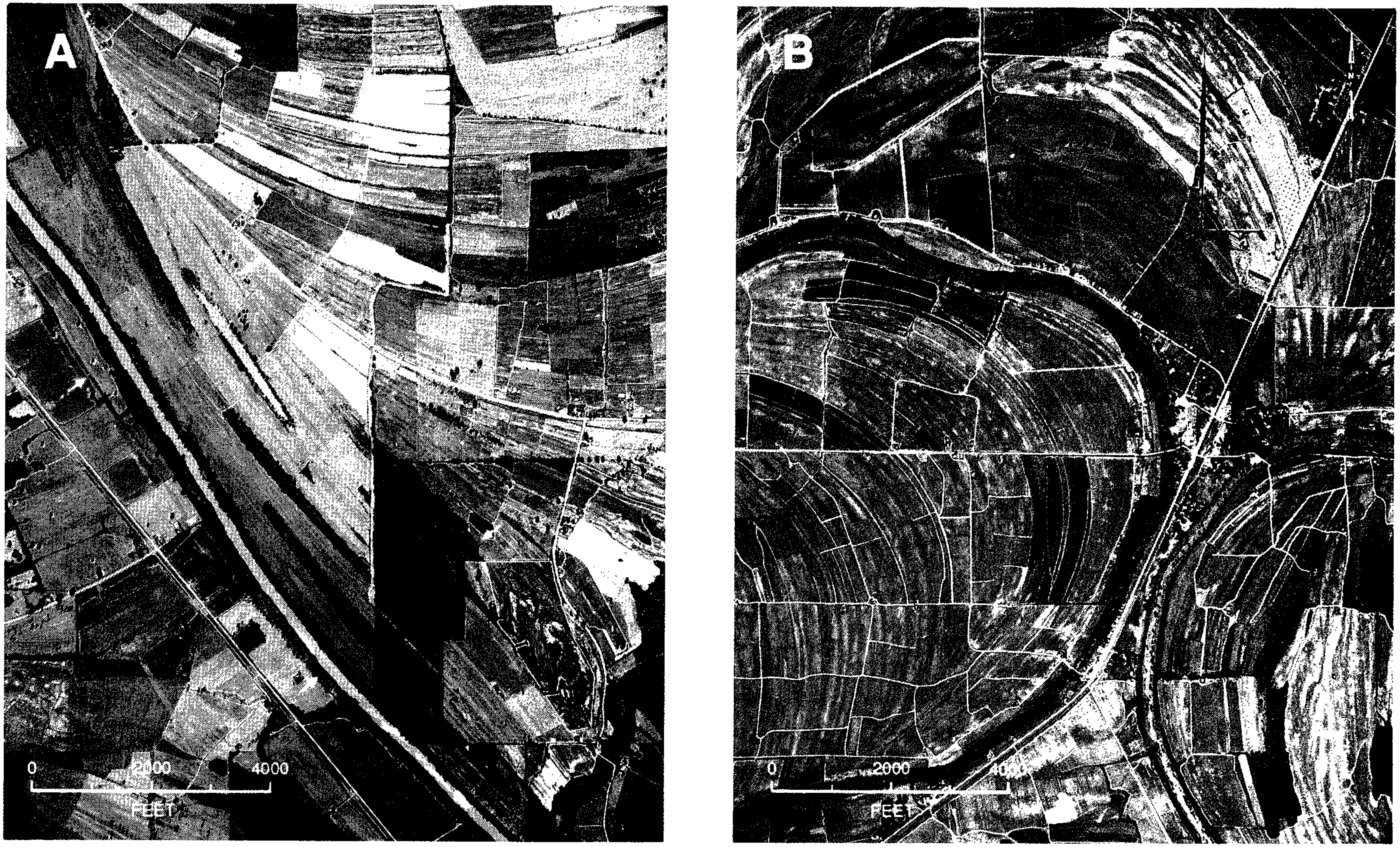

Figure 24. Abandoned courses. A: Mississippi River abandoned course now occupied by the Tensas River southeast of Sicily Island in Tensas Ph., LA (Plate 10) (estimated full-flow channel bank lines indicated by arrows); B: Underfit streams (Hopson and Cassidy bayous) that have exceeded and largely obliterated the limits of the former Mississippi River course northeast of Tutwiler in Quitman and Tallahatchie Cos., MS (Plate 7)

(M3D, Plate 1) reoccupied and developed a new meander belt in a portion of the older Bayou Portage course and meander belt (M4E, Plate 1; Plate 11). An intuitive explanation involves the infrequent nature of avulsions along the major rivers coupled with the fact that there have been large floodbasin areas where flow could be diverted without encountering older meander belts and courses. The fact that smaller rivers more frequently occupied abandoned courses of the larger rivers may only be a reflection of the fact that their flow entered the larger river anyway (as a tributary) and simply remained after an upstream avulsion diverted the flow of the larger system.

Meander belt morphology

Knowledge of Lower Mississippi Valley geomorphic history and processes primarily has been gained from the study of specific environments or the detailed mapping of small areas. There have been only four major attempts to step back and take an integrating overview--Fisk (1944), Saucier 1974, Autin et al. (1991), and the present synthesis. In each case, the opportunity "to see the forest and not just the trees" has proven invaluable in formulating new ideas and hypotheses, both good and bad, about the sequence of events that have led to the present landscape. This type of holistic perspective has been especially valuable concerning the origin and evolution of Holocene Mississippi River meander belts. There are several aspects of their overall size, configuration, symmetry, and downvalley trends that appear to be potentially more valuable clues to area geomorphic history than any aspect of individual environments or features. Most of the resulting observations have not been published or widely discussed.

The first of the significant observations based on a "mesoscale analysis" was that of Fisk (1944), wherein he recognized appreciable differences in the overall width and size of individual landforms (e.g., abandoned channels) within what he interpreted as five meander belts of the Yazoo Basin (six in this volume). These differences are clearly evident in Plate 8 and Figure 25A. He correctly surmised that the size variations were caused by significant differences in total discharge through the systems. His explanation for the variations was that some of the meander belts were formed by the Ohio and Mississippi rivers at times when these two streams flowed in separate channels (and meander belts) to points as far south as a few miles below Natchez. This concept (and its elaboration and illustration) became one of the more famous aspects of his 1944 report. Unfortunately, however, it has proven to be incorrect.

By the early 1960s, it became apparent from the developing chronology of outwash deposition that there could not have been a time when, simultaneously, the Mississippi River flowed in a braided course and the Ohio River in a separate meandering course below Cairo as postulated by Fisk. Moreover, detailed mapping of environments of deposition in the St. Francis Basin (Saucier 1964) indicated it would have been impossible for the two rivers to have flowed side-by-side in meandering courses. The extent and continuity of

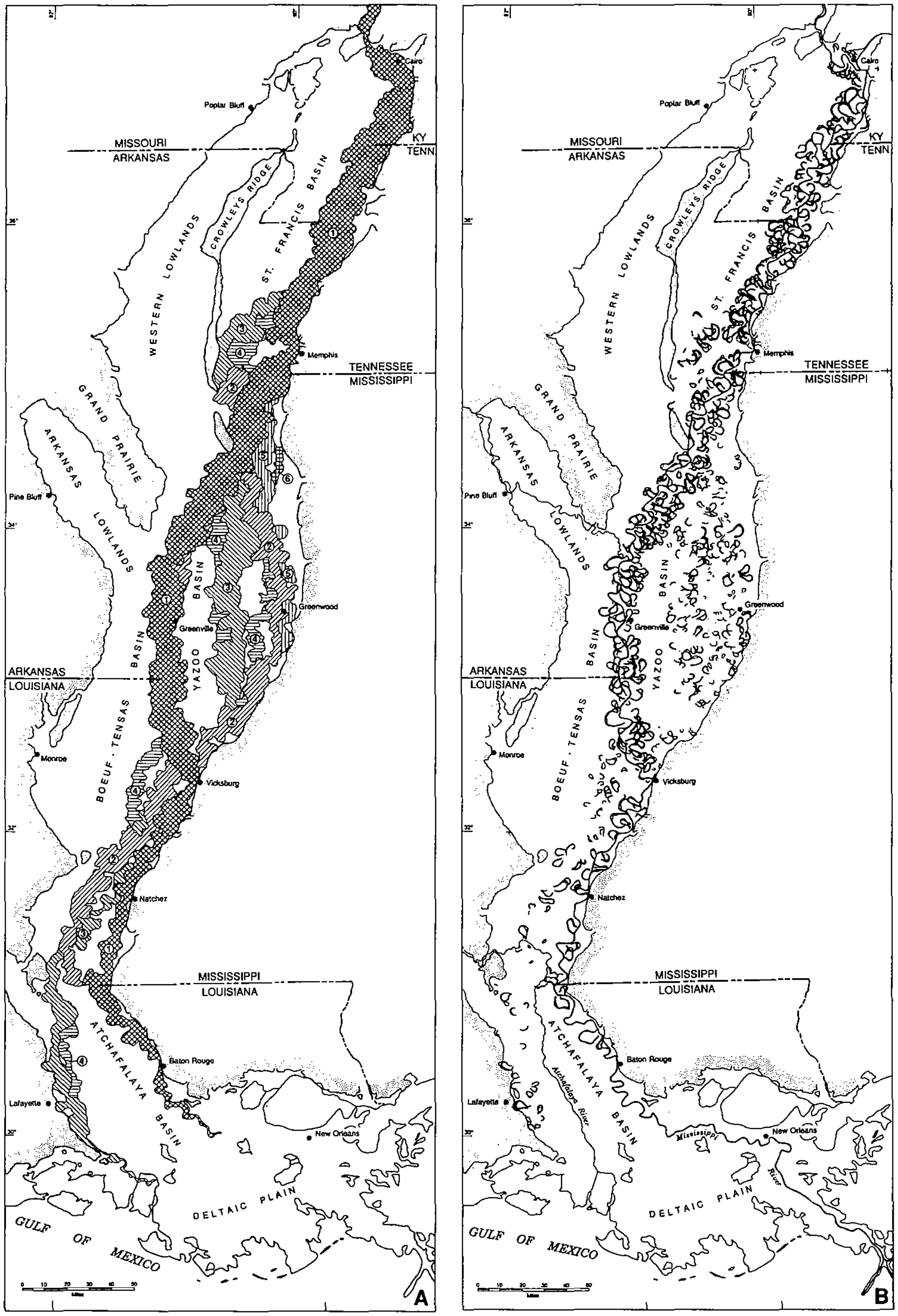

Figure 25. Configurations of Mississippi River meander belts as defined by the extent of the point bar environment (A), and distribution of neck and chute cutoffs (B)

valley train deposits in the area east of Sikeston Ridge (Plate 5) is such that two meander belts could not have been present in the narrow gap between those deposits and the valley wall. Saucier (1974) pointed out nearly 20 years ago that Fisk's interpretation was untenable, but a satisfactory alternative has been extremely elusive despite considerable deliberation.

As mentioned earlier (Chapter III), this writer first tried without success to establish a causal relationship between the size differences in the meander belts and climatically induced variations in discharge (Saucier 1985). Despite the attractiveness and logic of this explanation, there are unreconcilable conflicts between the probable ages of the meander belts in question and the timing of known regional climatic episodes (e.g., the Hypsithermal) that could have been responsible. Correlations of meander belt *formation* with the onset of episodes of either increased or decreased storminess are more promising but, with one possible exception (discussed below), do not explain the differences in size between meander belts.

More recently, this writer compiled for the first time a single small-scale map showing the distribution of all known Mississippi River abandoned channels (both neck and chute cutoffs) in the entire Lower Mississippi Valley area (Figure 25B). The results of this exercise are intriguing!

If the best current estimates of the ages of the Holocene Mississippi River meander belts are even reasonably correct (Autin et al. 1991, and Chapter 8), all of the meander belts were active for a comparable length of time (i.e., from about 1,700 to 2,800 years). Consequently, all other factors being equal, they should contain a comparable number of abandoned channels (cutoffs), but a glimpse at Figure 25B indicates that this is far from being the case. For example, the older meander belts of the Yazoo Basin contain a disproportionately small number of cutoffs. In light of this situation, and in consideration of an interesting discernible pattern in the size variation in the cutoffs, this writer has developed and offers herewith for consideration a tentative new model of meander belt formation that is a major departure from traditional views.

It is proposed that the variations in meander belt size and numbers of cutoffs within the meander belts are attributable to the fact that with the exception of the present meander belt, none carried the total discharge of the Mississippi River for more than a very short time (at most a few hundred years?) and some never did at all. Under this scenario, the Yazoo (M2B, Plate 1; Hpm2, Plate 8), Bear Creek (M4B, Plate 1; Hpm4, Plate 8), and Tensas (M4D, Plate 1; Hpm4, Plate 9) meander belts perhaps more appropriately should be regarded as major distributaries. Careful examination of the distribution of cutoffs of various sizes reveals there are some that apparently formed during periods of waxing flow and distributary growth and others that formed during periods of waning flow and distributary decay.

Reflecting on the processes and conditions that could have been responsible for this postulated situation, this writer believes that throughout most of the Holocene, it was unusual for the Mississippi River to occupy a single course and meander belt for any appreciable length of time. Rather, it was more typical for the river to be dividing its discharge between two or more courses: while one meander belt (distributary?) was forming, another was being abandoned. If this is correct, the present and relatively long-lived meander belt could be considered an anomaly.

This hypothesis, if true, has serious implications regarding reconstructions of alluvial valley history and chronology. With the current state of the art, it is difficult to establish the order in which the meander belts formed assuming they did so in a simple sequential manner: if two or more were active at a given time, the problem becomes extraordinarily complex.

As has been pointed out, most of the information regarding meander belt chronology has been derived from archeological evidence. Locally determined or extrapolated dates on cultural deposits, especially mound sites, have been the basis for estimating that the present Mississippi River meander belt between Memphis and Vicksburg was initially occupied about 2,800 years ago (Autin et al. 1991). With the assumption that multiple meander belts in the Yazoo Basin area may have been active at the same time, it becomes highly probable that at least a small meander belt occurred in the area of the present one at a much earlier date. There are cutoffs along the present meander belt whose size suggests association with an older, less than full-flow course, but there is no direct evidence, including archeological evidence, to suggest a greater antiquity. This writer is not concerned about the lack of evidence because older sites may be buried by more recent alluvium or have been found but not been properly interpreted out of an incorrect assumption (based on existing geomorphic studies) that the landscape is too young.

The new model of meander belt formation proposed herein is most compatible with observable evidence and explains the unusually small number of cutoffs associated with earlier meander belts. However, it does not explain the unusually large number and abnormally large size of cutoffs along the present meander belt between Cairo and Vicksburg (Figure 25). Neither does the model explain the very pronounced assymetry in cutoff distribution along long meander belt stretches. For example, a vast majority of the cutoffs occur west of the present river for a distance of about 75 mi north of Memphis, whereas most occur east of the river between Helena and Greenville (Figure 25 and Plates 6, 7, and 8).

Various explanations, including length of meander belt occupation and structural control, have been considered, but it has been tentatively concluded that the only positive correlation is with the lithology of the bed and bank materials. The distribution of the zones of unusually abundant and relatively small cutoffs appears to be causally related to shallow masses of relatively coarse glacial outwash (valley trains). For example, the zone of high cutoff frequency north of Memphis correlates with the axis of the Late Wisconsin valley train in the St. Francis Basin, and the zone between Helena and Vicksburg in the present meander belt coincides with the trend of the Early Wisconsin valley train which connects the southern end of the Western Lowlands with the northern end of Macon Ridge. Zones of low cutoff frequency correlate well with areas of thick, fine-grained backswamp deposits.

This explanation--a simple and obvious one--invokes essentially only the well-established principle that cutbanks in meandering rivers migrate more rapidly where they can erode into noncohesive sediments. Correspondingly, point bars develop further and more rapidly where there is an abundant supply of coarse-grained sediment. This situation of higher channel instability rather quickly leads to more frequent cutoffs and relatively greater channel widths. Whether or not the presence of abundant coarse-grained bed and bank materials is the explanation for meander belt assymetry is another matter, however. This writer believes that it is a contributing factor, but other processes such as neotectonics or the configuration of the suballuvial surface may be relatively more important. The impact that regional tectonic tilting can have on the symmetry of meander belts has been well documented in a recent study in Montana (Leeder and Alexander 1987).

Climate has been dismissed as an explanation for size differences between meander belts and cannot explain longitudinal variations in the frequency of cutoffs, but there is one aspect of overall meander belt geometry where climate may be a factor. Irrespective of numbers of cutoffs, the present meander belt of the Mississippi River and that of the Arkansas River are noticeably wider than any of the abandoned ones (e.g., see Plates 7 and 8), including those that carried full flow. While there is no definitive evidence, the possibility cannot be completely dismissed that this is a reflection of an increased frequency of storminess and flooding that some workers believe has occurred during the last several thousand years (Royall, Delcourt, and Delcourt 1991).

It has been suggested[1] that the change in the character of meander belts, including width and numbers of cutoffs, may be due to a natural evolution through the Holocene from simple to complex geometries. The evolution would be a reflection of adjustments to a progressively stabilizing base level. While this possibility appears viable for the southern part of the alluvial valley (e.g., south of Vicksburg), it does not appear to be a tenable explanation for the Yazoo Basin and areas to the north. In those areas, there has been relatively little change in base level, and furthermore, the observed changes in meander belt geometry have not been progressively in one direction over time during the Holocene. The meander belts with simple geometries are by no means always the relatively older ones.

Stream diversions and meander belt initiation. In the Mississippi alluvial valley area, crevasse splays outnumber distributaries by at least an order of magnitude, and distributaries outnumber apparent cases of successful diversions by a similar factor. Complete diversions require unusually favorable circumstances in both the parent meander belt and the receiving basin. Distributary development is an important step in the process of a successful stream

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[1] Personal Communication, 1994, Whitney Autin, Assistant Professor at Louisiana State University, Baton Rouge, LA.

diversion, but there appears to be a very small probability that even a large distributary will affect a stream diversion. In the southern part of the St. Francis Basin, the Mississippi River repeatedly formed well-developed distributaries in the Little River Lowland (Plate 6), but no complete diversion of the river took place. Similarly, in the Atchafalaya Basin area, numerous distributaries developed from both the present and the Teche meander belts (Plate 11), but the threatened diversion of the Mississippi River into the basin during this century is the only event of its type and magnitude to occur in thousands of years despite obvious channel gradient advantages.

As observed by Fisk (1952), diversions are known to have occurred along alluviating rivers in various parts of the world, but geologically they are infrequent events and have been little studied. Farrell (1987), however, contains an excellent discussion and illustration of the stages involved in the formation of a new meander belt (referred to therein as a channel belt) based on studies in the False River area of Louisiana (Plate 11). Developmental stages include deepening of an existing backswamp stream, initial natural levee development, further channel enlargement, initiation of meandering, rapid levee buildup, meander belt widening, and finally channel stabilization. The first four stages occur in all distributaries, but the last three stages occur only when a complete diversion takes place.

In a consideration of the development of a new meander belt, all attention should not be focused just on the newly forming channel. A major contribution of Fisk (1952) in his study of the possible diversion of the Mississippi River into the Atchafalaya Basin was his recognition of the role of the channel being abandoned. He estimated that when flow in the old channel fell below a critical level of about 60 percent of the total discharge, plugging with sediment effectively prevented the possible return of higher discharges. From that point on, a complete avulsion was inevitable.

Detailed mapping and interpretations of environments of deposition in the alluvial valley have revealed that small distributaries are present in unusual numbers along the flanks of abandoned meander belts just downstream from the points of diversion. This suggests that the diversions took place not just from the random, circumstantial enlargement of a single distributary, but rather they occurred at a point or in a small area where for some reason conditions were unusually favorable for crevassing and distributary development. Possible reasons could include meandering of the stream to near the edge of its meander belt, a favorable alignment and/or interconnection of cutoffs, headward erosion of backswamp drainage, faulting or structural control, and others. Whatever the reason(s), it is apparent that circumstances overrode the normal tendency of a stream to close off or "heal" a crevasse soon after it formed.

Backswamp/floodbasins. A backswamp is a simple and easily defined depositional environment, but it is more complex than has been interpreted and delineated in the Lower Mississippi Valley area and requires explanation. In purely geomorphic terms, a backswamp is a flat, shallow, poorly-drained, typically swampy or marshy floodplain depression bounded by natural levees or other uplands (Figure 26A). The term "flood basin" is often used synonymously with backswamp, and the term "rim swamp" is sometimes used when the area lies between a natural levee of a meander belt and the valley margin (either dissected upland or terrace) (Figure 26B).

The backswamp environment is characterized by the incremental accumulation of fine-grained sediments during periods of overbank flooding. Sedimentation rates are the lowest to occur on the floodplain since backswamps lie beyond the limits of natural levee development. Backswamps typically are poorly drained with small, low gradient streams flowing in chaotic or anastomosing patterns. Depending on the degree of ponding of water, backswamp subenvironments can be classified as lakes, poorly-drained swamps, or well-drained swamps (Krinitzsky and Smith 1969). In the Lower Mississippi Valley area, cypress and tupelo typically are the only forest species that can tolerate prolonged flooding and soil saturation. Virtually all backswamps experience significant seasonal water level variations: a swamp that might contain 5 to 10 ft of standing water for several months in the spring of the year may be completely dry and easily negotiated during the late summer and fall. Only relatively small areas of deep or poorly-drained swamp have permanent standing water.

When writing about drainage conditions and vegetative communities of backswamps, in most of the alluvial valley area it is more appropriate to use the past tense. Because of flood control, channelization, artificial drainage, and land clearing, most of the former extensive tracts of backswamp vegetation no longer exist. Significant stands of cypress swamp or bottomland hardwood forest exist only near the mouth of the Arkansas River, in the lower Yazoo Basin, in portions of the Tensas Basin, and in the Atchafalaya Basin. Elsewhere, especially in the northern part of the alluvial valley, they have been cleared for agriculture.

The distribution of the backswamp environment as shown in Plates 6 through 14 is based more on engineering considerations than on geomorphic ones and is of greater extent than if the definition given above had been strictly adhered to. Backswamp has been mapped primarily on the basis of the nature of the sedimentary sequence and, in reality, includes all areas not affected by lateral channel migration (point bar areas). Hence, areas delineated as backswamp include natural levees that flank the outer margins of meander belts and that occur beyond the limits of point bar areas along distributaries. As a generalization, the backswamp environment includes areas of thick, massive sequences of fine-grained overbank deposits as opposed to areas of thick, relatively coarse-grained point bar deposits. However, backswamp deposits must not be considered as uninterrupted sequences of pure clays and silts. Borings substantiate that all backswamp areas include frequent lenses or ribbons of noncohesive sediments representing minor crevasse splays and distributaries that are either too small or too deeply buried to be separately delineated.

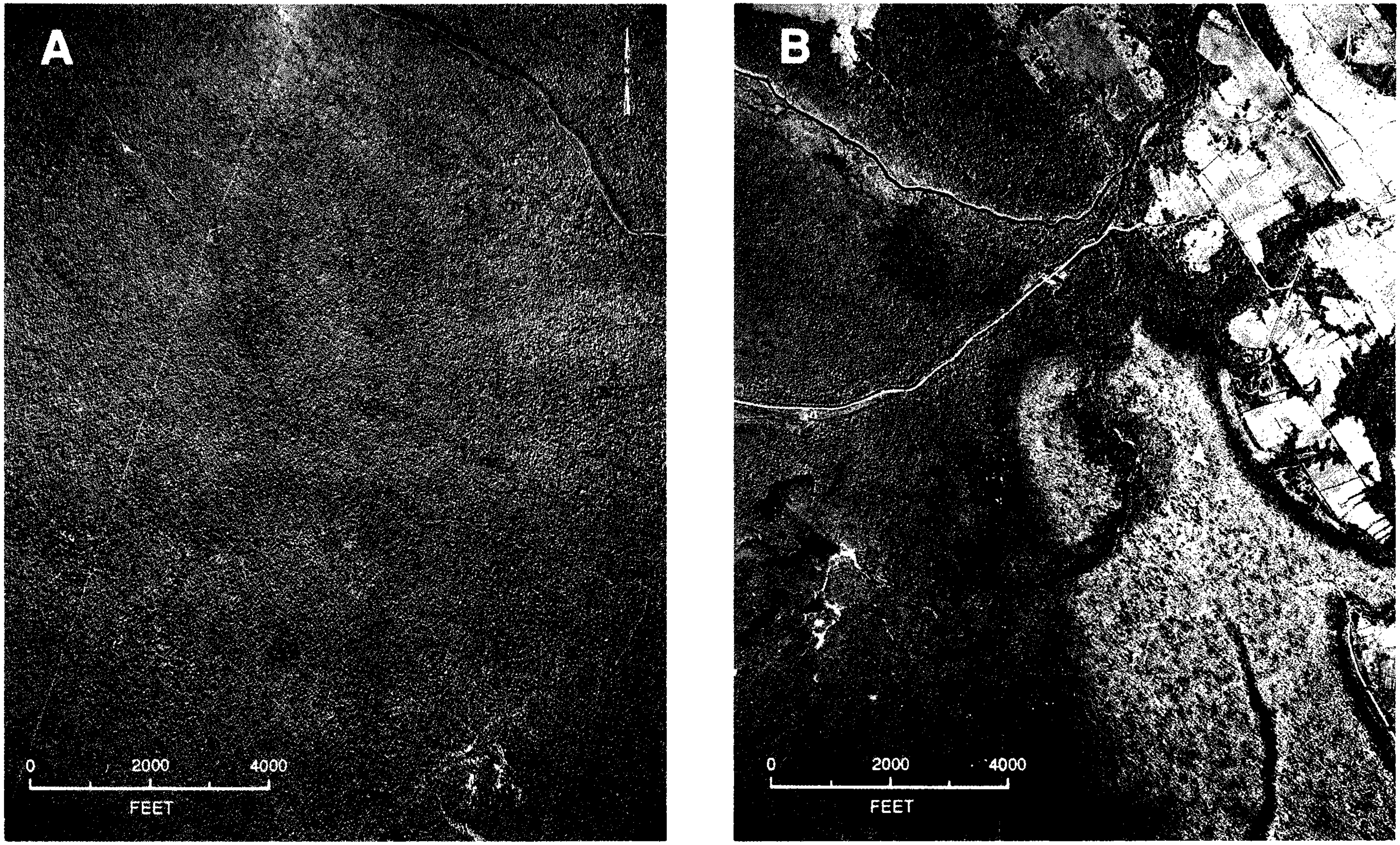

Figure 26. Backswamp. A: Nearly featureless, forested backswamp tract in the Atchafalaya Basin south of Lottie in Pointe Coupee Ph., LA (Plate 11); B: Rim swamp in the Spanish Lake area south of Baton Rouge at the junction of Ascension, East Baton Rouge and Iberville Parishes, LA (Plate 11) (cultivated areas lie on the Prairie Complex and the Mississippi River meander belt lies to the west beyond the limits of the photo)

Some backswamp areas, such as have been mapped in the Arkansas Lowland and the upper end of the Boeuf Basin (Plates 7 and 8), perhaps would be more appropriately designated as flood basin areas in consideration of the lithologic connotation of the term backswamp. Although these areas meet the definition of backswamp, the sediments are noticeably coarser than those elsewhere in the alluvial valley area as a reflection of the character of the Arkansas River sediment load. In fact, the percentage of silt and sand is sufficiently high that in the original mapping of depositional environments (Saucier 1967), many of the areas of backswamp were designated as valley train deposits (braided-stream deposits). That interpretation was recently changed, not on the basis of lithology, but rather because of new interpretations of regional stratigraphy and valley history.

Everywhere in the Lower Mississippi Valley area, backswamp deposits directly overlie and typically are abruptly separated from the coarse-grained substratum (glacial outwash deposits). The thickness of the backswamp deposits slowly but progressively increases downvalley from an average of about 40 ft at the latitude of Memphis to about 60 ft at the latitude of Natchez. Thereafter, the average thickness rapidly increases to well over 100 ft in the southern part of the Atchafalaya Basin.

As shown in Plates 13 and 14, the backswamp environment has been mapped as extending well downstream into the deltaic plain of Louisiana. The southern limit is shown as extending southward from the Mississippi River through Thibodaux to the vicinity of Houma, and thence westward (and conveniently) along the Teche meander belt. This line is largely arbitrary since the backswamp environment actually grades laterally into the inland swamp and interdistributary environments at the surface and interfingers at depth with the intradelta environment. There are no soils or vegetation differences that allow easy differentiation. Inland swamps are described in more detail in the discussion of deltaic depositional environments.

Lacustrine Environments

It has already been pointed out that lakes can form in the alluvial valley and deltaic plain in a variety of environments and for a variety of reasons. Some, like Catahoula Lake in central Louisiana (Plates 1 and 10), form in basins that have a very small sediment input because of the peculiarities of their geographic location. Others, like Reelfoot Lake in Tennessee (Plates 1 and 5) and Lake St. Francis and Big Lake in Arkansas (Plates 1 and 6), are caused or strongly influenced by tectonics. Still others like Lake Monroe that formed in the Ouachita River valley because of the daming effect of glacial outwash, as well as lakes that formed in valley train channels are due to a special set of circumstances. The vast majority of lakes, however, occur as ephemeral features in backswamp and interdistributary environments and are due to a prevalence of erosional processes (e.g., sediment flushing, subsidence, bank erosion) over inorganic sedimentation. This section pertains just to the latter and focuses on lacustrine areas in the Holocene backswamp and inland swamp environments: it does not include discussion of the larger lakes of the intratidal part of the deltaic plain which are discussed under the bay-sound environment. It also does not include the sunk lands of the New Madrid Seismic Zone in Arkansas because although they are backswamp environments (Guccione, Prior, and Rutledge 1990), their physical characteristics and bottom sediments were strongly influenced by earthquake-induced liquefaction.

Lacustrine deposits

Lake filling as a depositional environment is of areal consequence only in the southern part of the alluvial valley generally south of the latitude of Natchez. Lacustrine deposits have been mapped and studied (both at the surface and in the subsurface) essentially only in the Atchafalaya Basin area (Krinitzsky and Smith 1969, Smith and Russ 1974, May et al. 1984).

Lacustrine sedimentation is characterized primarily by fine-grained deposition in shallow, freshwater lakes (Smith, Dunbar, and Britsch 1986). Deposits are laid down under low-energy conditions mostly in quiescent water during periods of basin-wide, overbank flooding and slowly accrete vertically. The deposits periodically are disturbed by wind-generated waves and currents, and some winnowing and flushing and sediment removal may occur.

The horizontal limits of lakes and lacustrine deposits in a basically backswamp setting are highly variable. In large lakes, shoreline erosion often results in well-defined bank lines with low scarps and, in exceptional cases, small beaches (composed of silt and/or shell) may be present. In smaller lakes or where filling has been appreciable, there may be a progression from open water through standing, live cypress stands in perennial, shallow water to a deep swamp environment. In such cases, no true bank lines exist and the limits of the environments may be extremely hard to define. Moreover, in time frames measured in terms of decades or centuries, swamp forest vegetation may encroach into lakes that are filling or may be dying out and left as "ghost forests" in shallow water where erosion or lake deepening is occurring.

Lacustrine deltas

Lacustrine deltas, although considerably more limited in extent, have been the focus of appreciable recent interest and study because of their sedimentologic significance in fluviodeltaic sequences (Smith, Dunbar, and Britsch 1986; Breland 1988; Tye and Coleman 1989a, 1989b). The only contemporary examples occur in Lake Fausse Point, Grand Lake, and Six Mile Lake in the southwestern part of the Atchafalaya Basin (Plates 11 and 13). There is evidence of their presence in floodbasins elsewhere in the alluvial valley area, but optimal conditions for their formation and preservation in the sedimentary sequence occur only in the upper part of the deltaic plain.

Lacustrine deltaic deposits are relatively much coarser than those of purely lacustrine origin (but still only silts and sands) and occur as coarsening upward sequences. They form when a fluvial system (e.g., a distributary or newly enlargening course) progrades into shallow, open water. Within a given flood basin or backswamp area, lacustrine delta formation typically is a cyclic and relatively rapid process measurable in terms of decades.

Based on studies of historic-period lacustrine delta formation in the Atchafalaya Basin (Breland 1988, Tye and Coleman 1989a), the surface of a typical delta consists of a complex network of small, ephemeral, branching, lacustrine distributaries. A subsurface model of the deltaic sediments includes a characteristic sequence of deposits laid down in a series of distinctive subenvironments. From bottom to top, this includes prodelta, delta front, distributary mouth bar, channel, interdistributary trough, and natural levee. Once a delta has formed and the locus of active deposition shifts elsewhere, regional and local subsidence become predominant processes and the sedimentary cycle concludes with reversion to backswamp or lacustrine conditions.

Eolian Environments

Loess

No single geologic unit or depositional environment in the alluvial valley area has received as much attention or has been the subject of such varied investigations as loess. Significant literature is available dealing with loess origin, geomorphology, chronology, stratigraphy, sedimentology, mineralogy, pedology, and paleontology from the perspectives of geologists, soil scientists, engineers, and others. For basic references that address multiple aspects of loess and that contain current and generally accepted data and concepts, the reader is referred to Krinitzsky and Turnbull (1967), Snowden and Priddy (1968), Pye and Johnson (1988), and McCraw and Autin (1989).

Loess is a relatively homogeneous, seemingly nonstratified, unconsolidated deposit consisting primarily of well-sorted silt (Fairbridge and Bourgeois 1978). It occurs as a blanket, composed of several discrete loess sheets, that drapes upland formations of Quaternary and Tertiary age. It is conspicuous because of its unusually massive nature, typical uniformly tan to brown color, and its extraordinary ability to form and maintain vertical slopes or cliffs. Loess of the Lower Mississippi Valley area is contiguous with an extensive blanket in the central United States and together form one of the largest blankets in the world (Snowden and Priddy 1968). Deposits are present east of the alluvial valley as an essentially continuous 30- to 60-mi-wide veneer from the Cairo area south to below Baton Rouge. It is in this area that it reaches its maximum thickness (more than 75 ft) and has the most prominent outcrops. Loess is also present as a much thinner (10 ft or less) veneer in a narrow band along the edge of the Ozark Escarpment, on the higher valley train levels west of Crowley's Ridge, on the Grand Prairie and Macon Ridge, and as a band about 20 mi wide on Quaternary terraces west of the alluvial valley from southeast Arkansas to the Gulf (Autin et al. 1991). In the latter area, the loess is weathered and leached and requires careful analysis to recognize and differentiate from underlying silty deposits of fluvial origin.

Lyell (1847) is generally credited with being the first to describe the unusual nature of loess in the alluvial valley area, and Chamberlin (1897) the first to propose that it formed chiefly through the eolian redeposition of detritus from till sheets and glacial outwash in the floodplains of major alluvial rivers (Krinitzsky and Turnbull 1967). This concept became widely accepted and remained unquestioned until Russell (1944) argued that loess was of colluvial *in situ* origin, originating from the downslope reworking of backswamp deposits of Pleistocene age. He described the inferred process and called it loessification. This hypothesis was a direct outgrowth of the work of Fisk in recognizing Quaternary terraces as being uplifted alluvial sequences and mapping their distribution in the uplands adjacent to the alluvial valley. In the face of immediate and staunch criticism (Wascher, Humbert, and Cady 1948), Fisk (1951) strongly endorsed the concept of Russell and remained his only outspoken supporter. During the following decade, however, evidence in favor of an eolian origin of loess eventually became overwhelming, in part due to the debate and controversy stirred by Russell and Fisk.

One of the reasons given by Russell and Fisk for rejecting an eolian origin of loess was an inability to hypothesize a direction for winds that would explain the distribution of loess. That issue has been resolved by the work of Rutledge, West, and Omakupt (1985) and others who have correlated loess deposits of Crowley's Ridge and in the uplands east of the alluvial valley to valley trains of the Western Lowlands and the St. Francis Basin. This work substantiates that during times of near-maximum to early-waning glaciation, seasonally prevailing, strong, north and northwest winds deflated enormous quantities of silt from recently deposited, unvegetated masses of glacial outwash and transported the material for tens to hundreds of miles to the east and south. In the uplands, the deflated silt was incrementally deposited in the uplands as a drape over a dissected, hilly landscape, with the greatest amount of material and relatively coarsest material deposited closest to the source areas.

On a regional scale, the thickest loess deposits occur on Crowley's Ridge and in the Vicksburg-Natchez area. The former was immediately adjacent to and the latter was directly downwind from extensive valley trains that periodically characterized the central alluvial valley area. Location in relation to source areas, however, does not adequately explain the rapid thinning of loess sheets south of Natchez. This situation may have been influenced more by climatic conditions such as a southward decrease in the frequency or intensity of northerly winds. On the other hand, according to a loess process model proposed by Miller and Alford (1985), loess deflation episodes may have been attributable to the devegetation of valley train surfaces by rapid sedimentation during outburst floods. That model would also satisfactorily explain the southward decrease in loess deposition in the valley area.

The chronology of outwash deposition and associated loess deposition indicates that tens of feet of loess were deposited in some parts of the uplands during intervals as short as 10,000 years or less. Pye and Johnson (1988) have recently estimated that depositional rates exceeded 2 mm/year between about 15,500 and 17,000 years BP for the most recent loess sheet in the Vicksburg area. Because it is now realized that the times of maximum loess deposition were coincident with cooler and wetter climatic conditions, it is necessary to assume that the deposited silt could have accumulated to such thicknesses only under the condition of a dense forest cover. It is further necessary to assume that deflation took place during relatively dry late summer and autumn conditions and that heavy precipitation was restricted to the warm seasons.

Although massive and apparently unstratified in fresh exposures, weathering reveals that loess deposits within a given sheet typically are subtly banded or bedded with individual layers conforming to the configuration of the underlying hilly landscape. The bands reflect minor variations in lithology (clay content), calcareous content, organic matter, and density, and are believed to represent variations in the rate of silt deposition. In essence, the bands are weak soil horizons. With the assumption that they reflect regional rather than local conditions, and with improved knowledge of valley chronology, someday they may be correlated to specific outwash events or episodes.

Individual loess sheets are separated by much more highly developed weathered horizons (paleosols) that are easily recognizable on the basis of lithology, texture, and color. At least five loess sheets and four well-expressed paleosols have been identified in the Lower Mississippi Valley area (Follmer et al. 1989). Efforts to identify, date, and correlate the loess sheets have been almost the exclusive goal of loess investigations of the last decade. Major progress has been made despite complications imposed by limitations of dating techniques, discontinuous distributions, structural control, and original spatial and temporal variations.

The five loess sheets that have been identified in the alluvial valley area are, from youngest to oldest, the Peoria, Roxana, Sicily Island, Crowley's Ridge, and Marianna (Autin et al. 1991). Other names, such as Vicksburg, Pre-Vicksburg, Farmdale, and Loveland, have been used locally or in attempts at extrapolation from the midwestern United States, but currently are not in favor. The Peoria loess is the only sheet that has been recognized in all parts of the area: all five sheets have been identified only on Crowley's Ridge and only three have been identified in the Vicksburg area. The Peoria and Sicily Island sheets are the only ones that have been identified in Louisiana.

Multiple lines of evidence have been used to date the Peoria loess, including numerous radiocarbon and thermoluminescence dates, and this loess is unquestionably of Late Wisconsin age. The majority of the radiometric dates fall between 22,000 and 12,500 years BP, which is completely compatible with dates on the valley trains originating from the Late Wisconsin glaciation (Johnson, Pye, and Stipp 1984; Snowden and Priddy 1968). There is considerable disagreement over the ages (and even correlations) of the Roxana and

Sicily Island loesses (especially the latter), with some workers favoring an Early Wisconsin age (see McCraw and Autin 1989) but others favoring a pre-Wisconsin age. Most workers agree that the two oldest loess sheets date to the Illinoian Stage or earlier.

A puzzling aspect of loess chronology is the apparent complete absence from the Lower Mississippi Valley area of evidence for loess sheets older than the Illinoian (Middle Middle Pleistocene shown in Figure 4). Currently active efforts by the U.S. Geological Survey at dating loess are indicating that no loess deposits are older than about 300,000 years.[1] If this is the case, there are no loess deposits representing the glacial cycles that occurred during more than 80 percent of the Pleistocene. It is possible but unlikely that evidence of earlier loess sheets has been *completely* removed from the uplands by erosion, and there is no reason to believe that climatic conditions during earlier glacial stages were significantly different than during the Wisconsin stage. An explanation favored by this writer takes into consideration the gross size and geometry of the alluvial valley. As has been pointed out, the alluvial valley has progressively widened during the Quaternary by planation by both braided and meandering streams. It is possible that the alluvial valley during the first glacial cycles was so narrow that glacial outwash was funnelled directly to the Gulf in a flume-like situation rather than being deposited on a broad, alluvial plain where it would have been subjected to widespread deflation.

Sand dunes

Surprisingly, the presence of large and conspicuous fields of sand dunes in the northern portion of the alluvial valley went completely unnoticed for many years. Fisk failed to recognize them and even cited their absence (Fisk 1951) as evidence for a non-eolian origin of loess. This writer (Saucier 1964, Smith and Saucier 1971) has been the only one to map their distribution and publish a paper discussing their origin and age (Saucier 1978).

Dunes composed of massive, well-sorted, fine sands occasionally occur as discrete features (megadunes) up to as much as 30 ft high and 25 acres in extent, but they more typically occur as 10- to 15-ft-high features within approximately 75 dune fields varying from about 5 to 25 sq mi in extent each. Each field contains tens to hundreds of individual dunes that occur as coalescing, circular to ellipsoidal mounds and irregular, lobate ridges (Figure 27A). These fields constitute undulating sand sheets that typically include numerous, small, enclosed depressions or deflation basins and active blowouts where soil conservation measures are not practiced (Figure 27B). The fields have a distinctive topography and are easily recognizable on aerial photographs (Saucier 1978).

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[1] Personal Communication, 1993, H. Markewich, U.S. Geological Survey, Atlanta, GA.

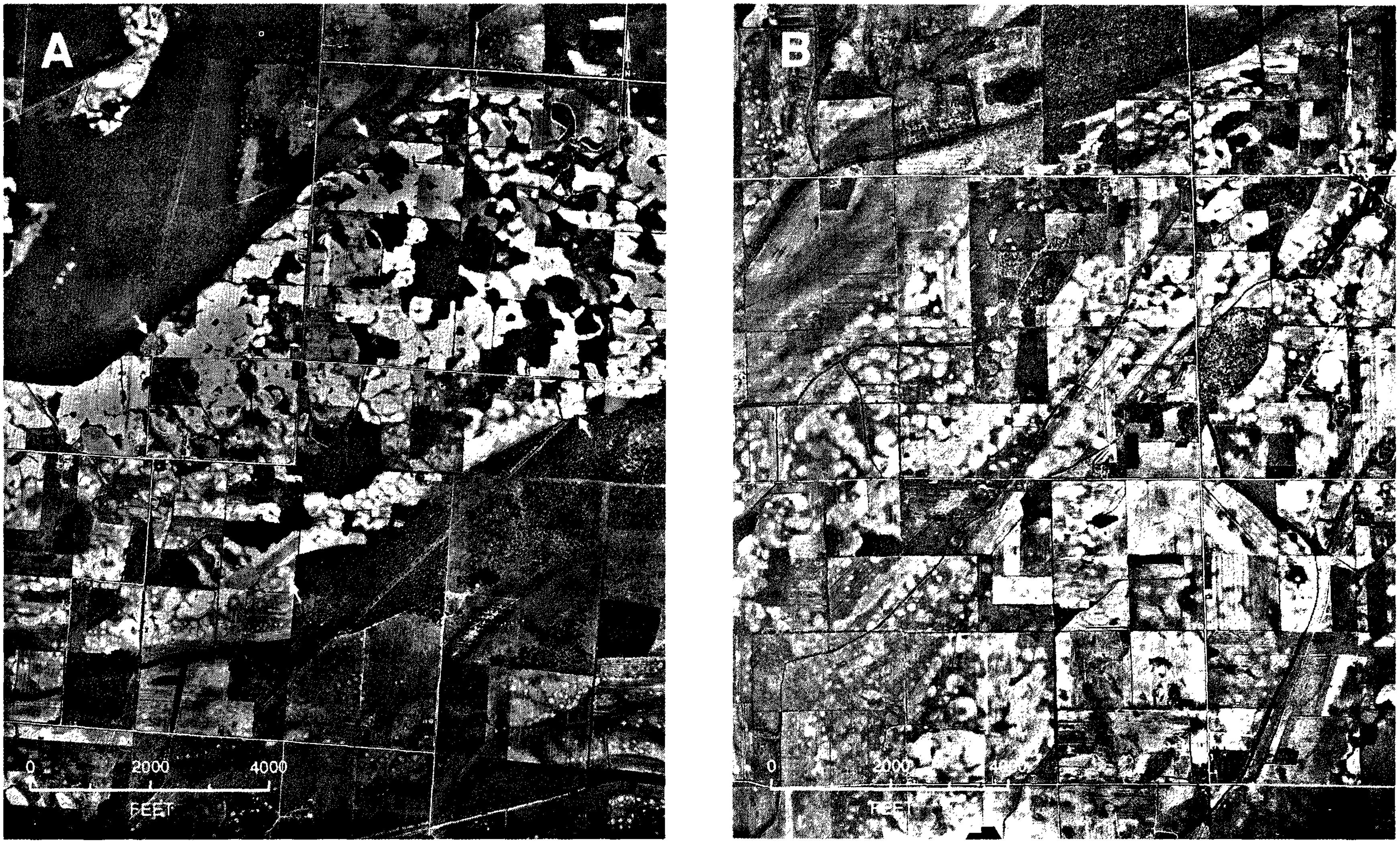


Figure 27. Sand dunes. A: Well-defined tract of sand dunes (between arrows) between braided channels south of Naylor in Ripley Co., MO (Plate 5) (some large depressions between dunes still forested); B: More intensely culitvated and subdued but still visible dunes (lighter soil tones) and depressions (darker soil tones) northwest of Walnut Ridge in Lawrence Co., AR (Plate 5)

Dune fields occur only along two distinct trends in the Western Lowlands and one in the St. Francis Basin. In the former area, the trend with the largest number of dune fields lies closest to the Ozark Escarpment and immediately east of the meander belts of the Current, Spring, and White rivers (Plates 4 through 7). The secondary trend flanks both sides of the Cache River (Plates 5 and 6), and includes a few isolated dune fields on higher valley train levels to the east. The trend in the St. Francis Basin occurs immediately east of and parallel to Sikeston's Ridge (Plates 4 and 5).

All dunes are located on valley trains of either Late or Early Wisconsin age and are situated within or immediately east of large, relict braided channels. They are believed to have formed on large midchannel braid bars or islands because of the deflation of sand from bare deposits of glacial outwash in the channels (between periods of seasonal flooding). As the strong, seasonal winds deflated the deposits, the sand was deposited close by, whereas the silt was transported much farther and deposited as loess. With multiple lines of evidence, including archeological site locations, the age of the dunes has been estimated to be between 12,000 and 30,000 years BP (Saucier 1978). Hence, the dunes and loess are manifestations of the same geomorphic process and are temporally equivalent.

Deposits intermediate between true dunes and loess occur as narrow, linear ridges a mile or so wide and up to several feet high along the western edges of the several valley train levels of the Western Lowlands. These deposits are coarser grained than typical loess yet too fine-grained to be regarded as dunes. They are interpreted as eolian features attributable to a particular range of wind conditions or a particular source of outwash sediments that existed at the edges of the braided channels at the bases of the low scarps.

Deltaic Environments

A delta is a body of sediment laid down by dynamic sedimentary processes in a zone of interaction where a river (fluvial system) enters a deeper and less turbulent body of water. When the fluvial system is large, like the Mississippi River, and the body of water is a marine system (the Gulf of Mexico), the resulting sedimentary body is a complex sequence involving multiple environments ranging from freshwater to saline. The precise nature, horizontal and vertical distribution, and relative importance of the environments are determined by the characteristics of the river regime, the coastal processes at work, the structural geologic setting, and climatic factors (Fairbridge and Bourgeois 1978). In the case of the Mississippi delta, the river regime involves a large load of fine-grained sediment; an appreciable range between high and low river stages; low wave energy, littoral current, and tidal ranges in the Gulf of Mexico; active regional and local subsidence; and a warm, moist climate that allows a dense vegetative cover. These variables have interacted within the context of a cyclic and progressively slowing sea level rise to produce the

Holocene deltaic plain. Both depositional and erosional processes have been involved.

Seven discrete environments of deposition are recognized as being present in a typical Mississippi delta lobe and are described below, relying heavily on the well-known studies of Scruton (1960), Coleman and Gagliano (1964), and Kolb and VanLopik (1958). A delta lobe herein is defined as that portion of a delta complex that formed during a relatively short period of time and that can be attributed to a single or discrete set of distributaries. Cumulatively, the environments encompass a cyclical, constructional, or progradational phase of deltaic plain formation in which fluvial processes dominate (Figure 28A and B). Environments involved in the destructional or transgressive phase of a lobe involve primarily deltaic-marine processes (Figure 28C) and are discussed in the next section. As pointed out by Coleman (1988), the constructional phase predominantly involves relatively coarse clastics (inorganic sediments) and rapid accumulation rates, whereas the destructional phase involves sediments rich in organic matter and slower accumulation rates.

It has been estimated (Coleman 1988) that the sedimentary sequence resulting from each delta cycle (a set of temporally associated lobes) covers an area of about 11,500 sq mi and has an average thickness of about 115 ft; however, this generalization may be misleading because of appreciable variation from one lobe to another. For example, the St. Bernard and Lafourche complexes (Figure 29) are areally quite extensive because they prograded into relatively shallow water (less than 50 ft deep). As a consequence, they are characterized by a lobate shape with the distributary ridges forming what has been referred to as a "horsetail" pattern (Fisk 1952). In contrast, there is the modern, or Balize, complex which has formed in waters hundreds of feet deep at the edge of the continental shelf. The distributaries of this lobe form what is referred to as a "birdfoot" pattern (Russell et al. 1936). Both complexes involve essentially the same depositional environments, but their distribution and characteristics vary considerably as discussed below.

Thanks to the intense interest of coastal sedimentologists in deltaic deposits and their facies, a large number of excellent illustrations of the surface and subsurface distribution of environments of deposition are already available. Figures 30 and 31 were selected to help the reader visualize the formation of a delta lobe and how multiple lobes interrelate to form complexes and, cumulatively, the deltaic plain.

Several attempts have been made to provide basic interpretations and to illustrate the distribution and sequence of delta complexes, and these in turn have been frequently modified. One of the most widely referenced basic models is that of Kolb and VanLopik 1958 (Figure 29) who initiated the much-copied "lollipop" diagram approach. Their model, recognizing seven complexes (termed subdeltas), has been superseded by the Frazier (1967) model that only recognizes five complexes (Figure 29). However, these are subdivided into 16 delta lobes as will be discussed in Chapter 7.

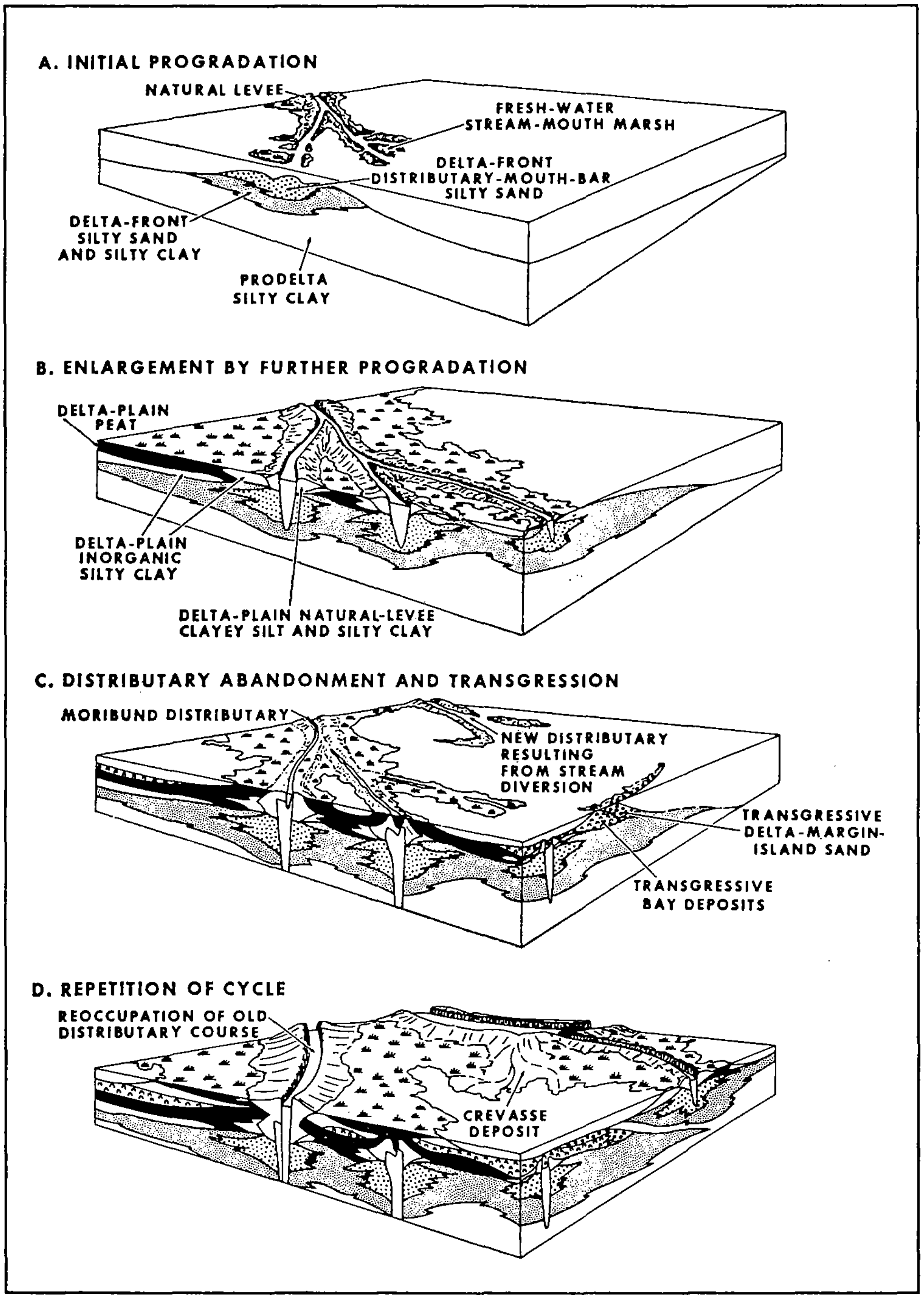


Figure 28. Idealized surface and subsurface distribution of environments of deposition at several stages in a typical delta cycle (from Frazier and Osanik 1965)

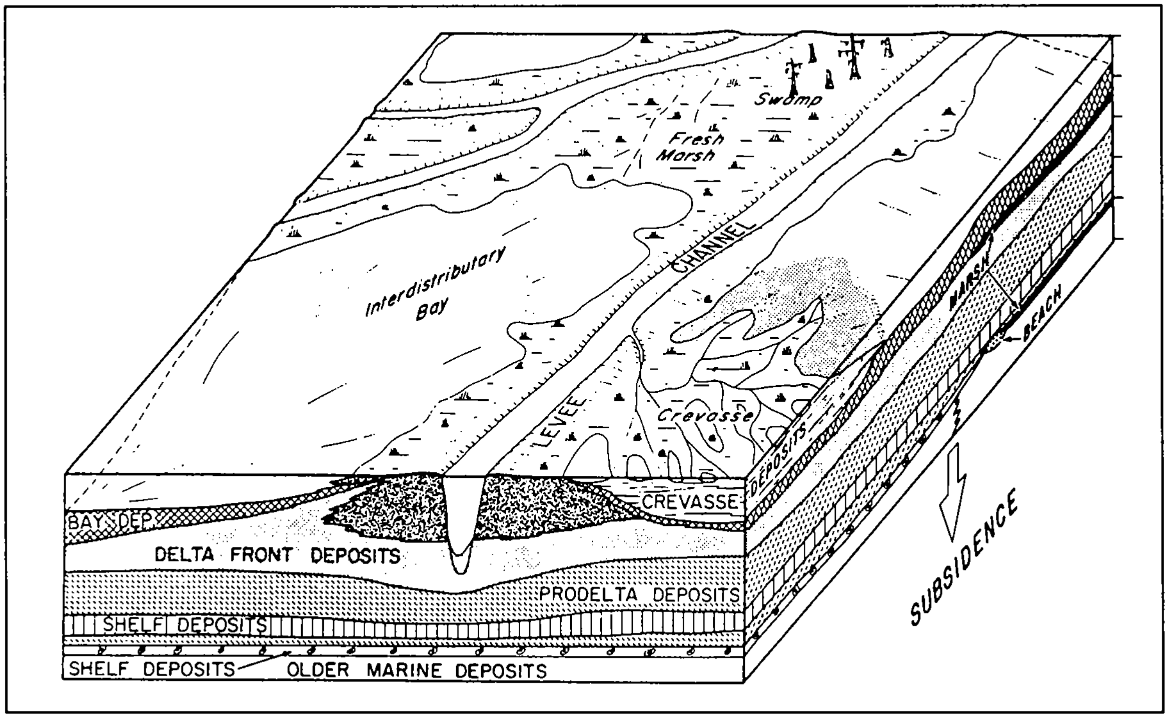


Figure 29. Block diagram showing relationships of subaerial and subaqueous deltaic environments of deposition in a single delta lobe (from Coleman and Roberts 1991)

Distributaries

Deltaic distributaries are the most conspicuous of the subaerial environments, being evident because of the natural levee ridges that flank the stream channels. The pattern of distributaries forms the skeletal framework of a delta lobe (Figure 30). As long as a distributary channel actively receives sediment, the river mouth progrades seaward at a rate directly related to the amount of discharge and sediment load, as well as the depth of the receiving water body. Distributary natural levee formation involves essentially the same fluvial processes that are involved in those in the alluvial valley area (i.e., deposition of sediments originating from overbank flow), but a few differences in morphology and lithology do occur. Because far less meandering takes place in the distributaries, the levees are more uniform in height and width. Rather than being laterally gradational with backswamp, natural levees of the lower deltaic plain grade into and/or are interfingered with interdistributary deposits and bay-sound deposits. Although distributaries frequently branch or bifurcate at relatively shallow angles, crevasse channels trending at high angles from the distributaries are uncommon. Crevasse splays are abundant only in the "bird-foot" delta, or Modern delta complex (also called Balize Complex), and are discussed under the interdistributary environment.

There are hundreds of deltaic plain distributaries that can be recognized on the ground or on aerial photos (May et al. 1984), but only the larger (wider

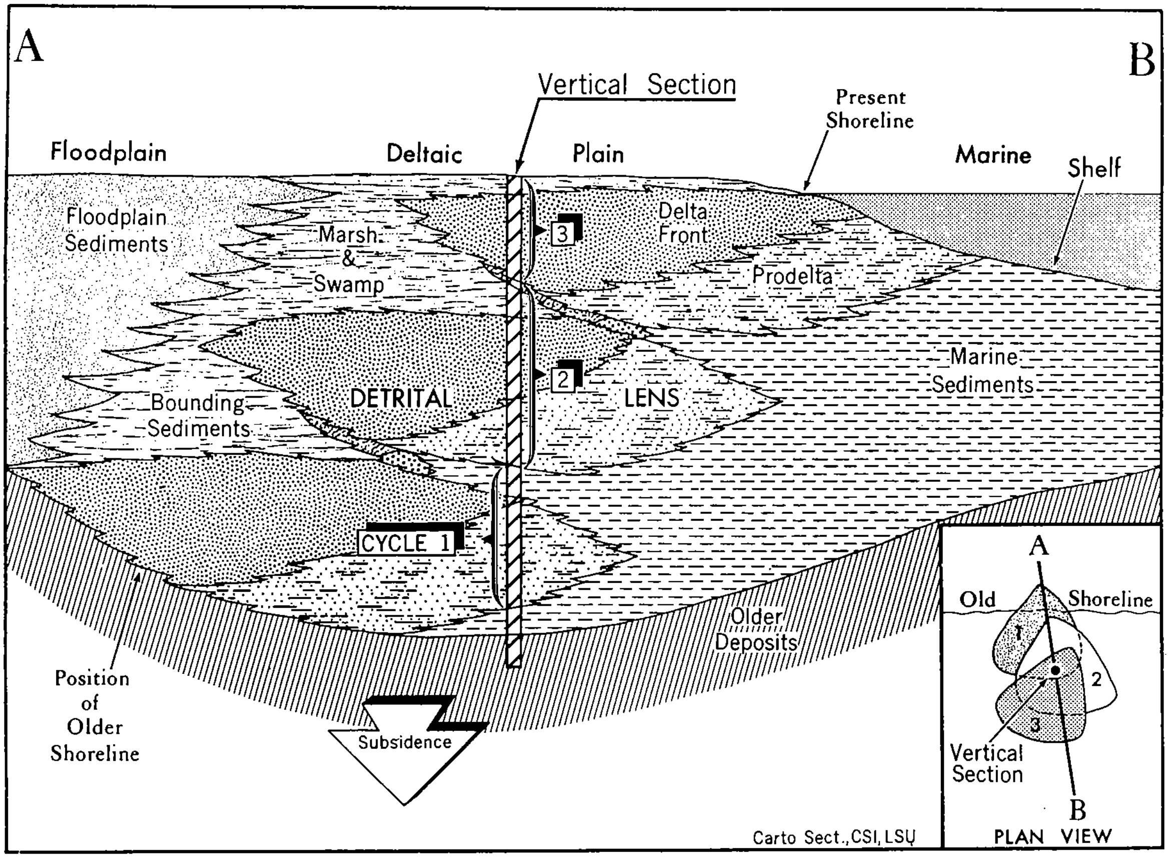
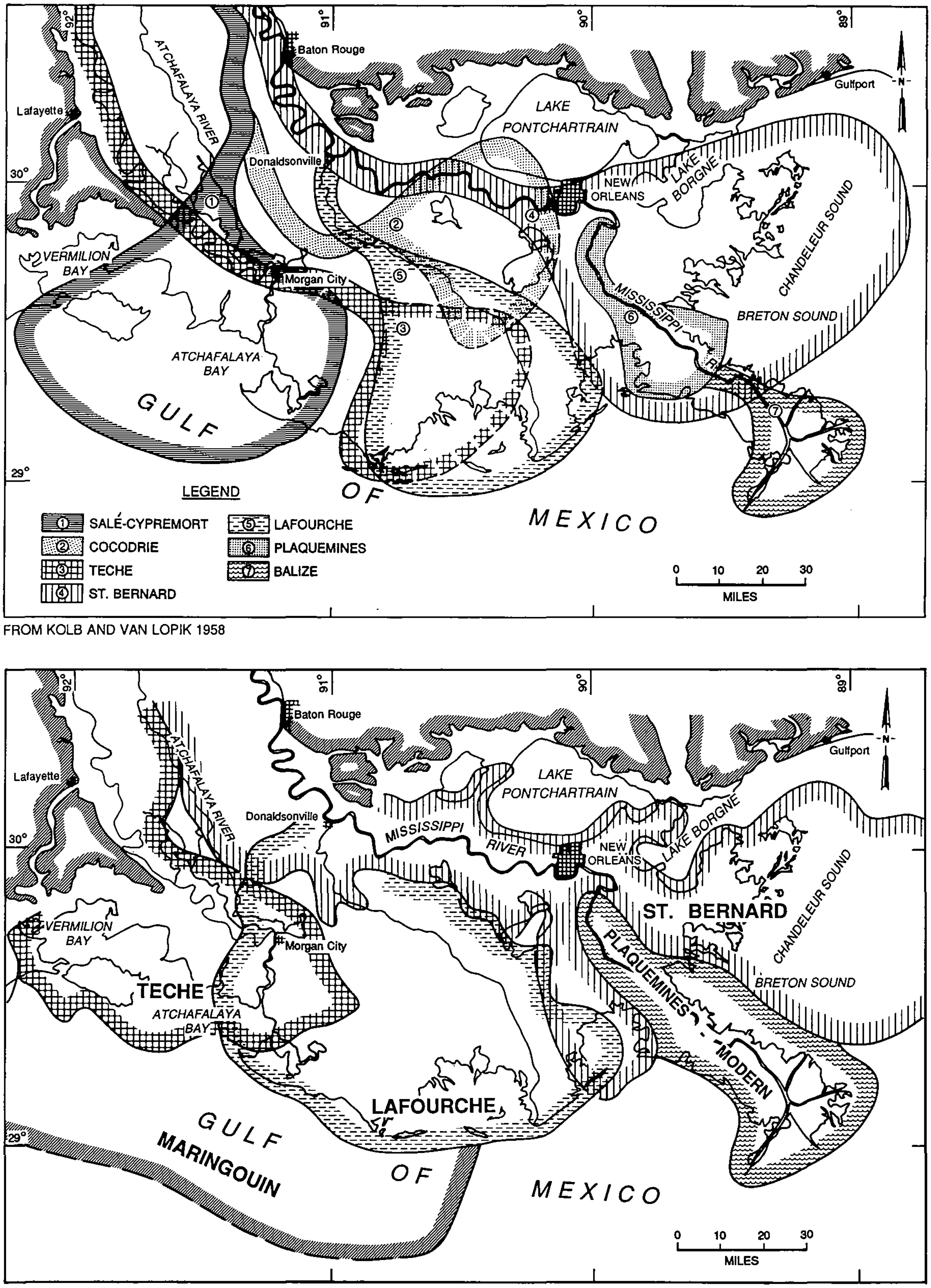


Figure 30. Hypothetical sedimentary sequence resulting from several overlapping deltaic cycles showing major environments of deposition (from Coleman and Gagliano 1964)

and longer) ones are shown in Plates 12 through 14. Those with natural levees narrower than about 1,000 ft, although numerous and often tens of miles long, are not delineated because of scale limitations. Natural levee ridges along the larger distributaries may be several miles wide and attain elevations of 10 to 15 ft above sea level (Figure 32A). In all cases, the natural levees steadily narrow and slope seaward at an average rate of about 0.25 ft/mi: the distal ends of most eventually disappear beneath intratidal marsh deposits (Figure 32B). Even where natural levees may be completely subsided without even a vegetation change to mark their presence, anomalies in local marsh drainage may still suggest their trends.

Based on the most definitive effort to date to establish a chronology of abandoned deltaic plain distributaries, with data from over 500 borings and over 150 radiocarbon dates (Frazier 1967), it is apparent that multiple distributaries were active at any given time in a prograding delta lobe. A typical major distributary (excluding those of the Modern Complex) probably never carried more than about 20 percent of the total discharge of the river. Borings



FROM KOLB AND VAN LOPIK 1958

FROM FRAZIER 1967

Figure 31. Highly popular, but outdated, interpretation of Holocene delta complexes (A), and currently accepted interpretation (B)

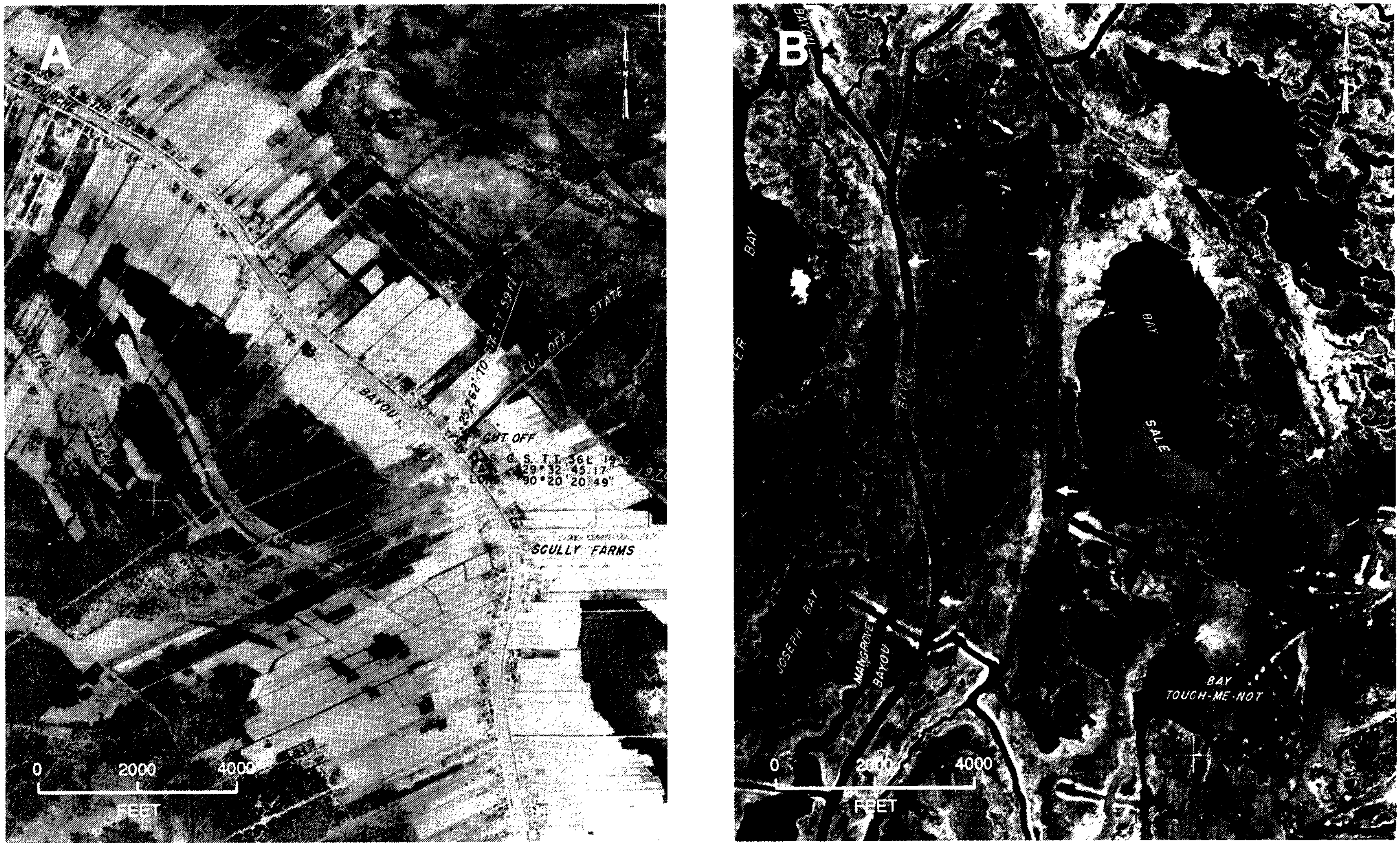


Figure 32. Distributaries. A: Bayou Lafourche, a major Mississippi River distributary in Terrebonne Ph., LA (Plate 14) marked by well-developed natural levees cleared for agriculture; B: Small distributaries in southern Terrebonne Ph., LA (arrows) indicated by conspicuous relict channels but whose subsided natural levees are barely discernible by slight vegetation differences (shrubs) from adjacent marsh

reveal that most distributary channels had a maximum width of less than 1,000 ft and a maximum depth of about 85 ft (Coleman and Roberts 1991). Some distributaries may have been active for more than a millennium, but most had an active life of only a few hundred years.

Inland swamps

As a depositional environment, inland swamps are highly similar to (and arbitrarily delineated from) backswamp areas, forming because of poor drainage in a basin or depression area characterized by very low sedimentation rates. Inland swamp tracts are extremely flat and seasonally flooded and, like backswamp areas, are typically bounded by natural levees or low terraces.

Inland swamps occur in freshwater areas at the upper ends of interdistributary basins (Figure 33A) and are differentiated on the basis of vegetation rather than soils, morphology, or causal processes (Figure 30). They involve a swamp forest community dominated by bald cypress (*Taxodium distichum*) and tupelo gum (*Nyssa aquatica*) that is essentially intolerant of brackish water except for very short periods of time (e.g., during a hurricane). Typical elevations are less than 5 ft above sea level. Where flooding is more prolonged (Figure 33B), a freshwater marsh community composed of grasses, sedges, and rushes exists rather than a swamp; where the prevailing surface water salt content exceeds even a few parts per thousand, a brackish marsh community occurs.

Inland swamps attain their greatest areal extent during the latter stages of a delta progradation cycle when the freshwater input to interdistributary basins is at a maximum. When delta lobe (or complex) deterioration begins, brackish water conditions move progressively inland, causing the swamp forest to die out and to be replaced with marsh vegetation. In the present deltaic plain, there has been no significant formation of new inland swamp during historic times. Tracts of dead timber standing as "ghost forests" are evidence that saltwater intrusion is taking place since all but the Modern Complex and the newly-forming Atchafalaya Complex are undergoing deterioration. During this century, however, this process has been greatly amplified by flood control, channelization, and drainage to the extent that inland swamps are rapidly disappearing from the deltaic plain landscape. Although the extent of inland swamp shown in Plates 12 through 14 appears significant, from a biological point of view, the plant communities are not in a healthy, vigorously reproducing state.

Deposits of the inland swamp environment consist of organic clays and woody peats. Because of ubiquitous subsidence, the thickest deposits occur in the older delta complexes and attain maximum thicknesses of about 20 ft (Frazier and Osanik 1965). Inland swamp deposits originating from earlier progradational cycles and now buried beneath sediments of other origin (e.g., natural levee or interdistributary marsh) have been detected in various parts of the deltaic plain but consist of thin layers not more than a few feet thick.

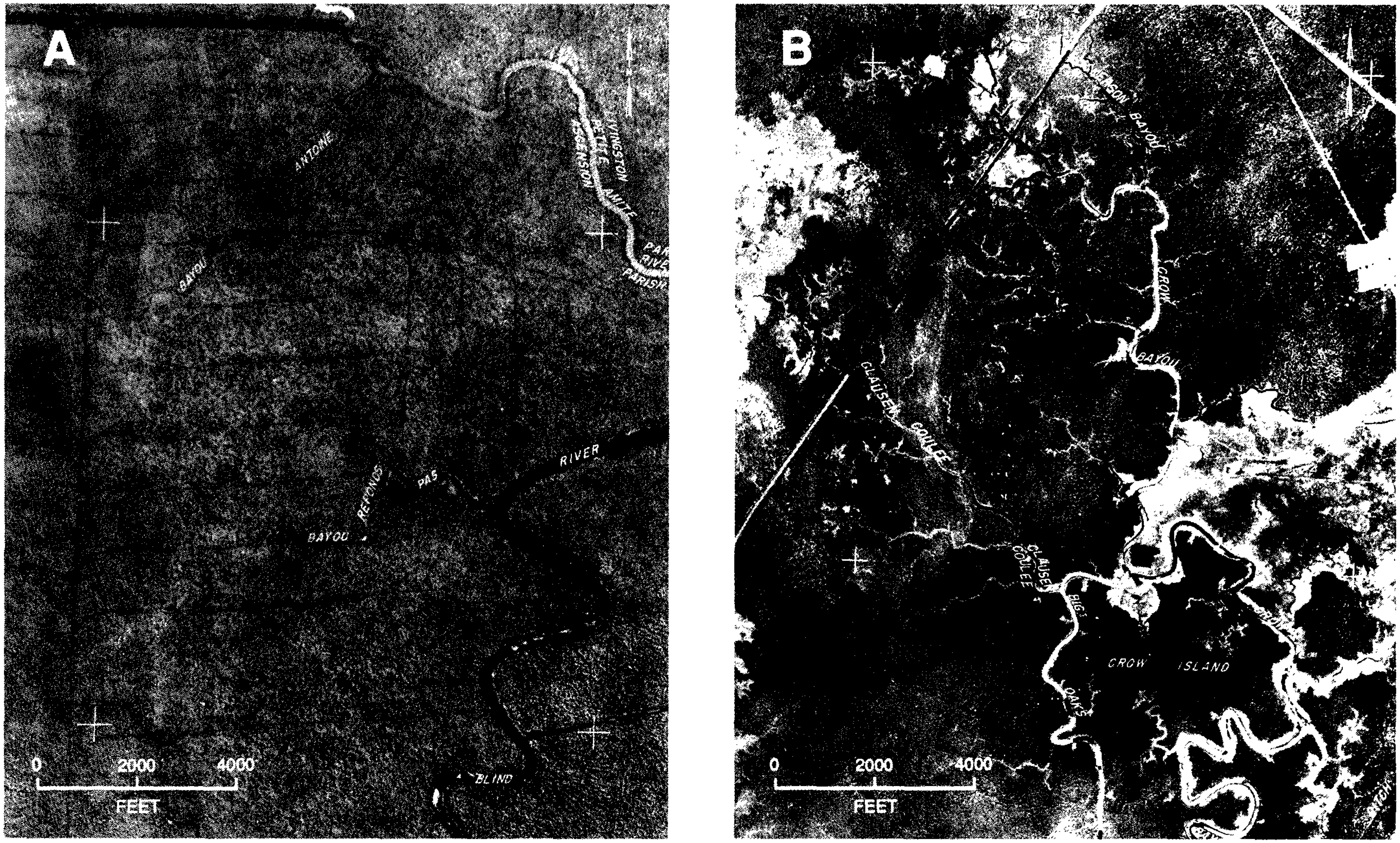


Figure 33. Inland swamp. A: Nearly featureless inland swamp southwest of Lake Maurepas in Ascension Ph., LA (Plate 12) (linear features are abandoned railroad spurs used in cypress logging operations in the early 20th century); B: Inland swamp in a more-coastal setting, interrupted by areas of freshwater marsh (lighter tones), near Calumet in St. Mary Parish, LA (Plate 13)

Most subsurface deposits are coincident with the present areas of inland swamp, indicating that the swamp environment in these areas has persisted for at least a few thousand years.

Interdistributary deposits/intratidal marsh

This environment, like the inland swamp environment, is defined primarily on the basis of the vegetative communities that it supports (Mitsch and Gosselink 1993). More than 50 percent of the present deltaic plain is characterized by interdistributary marsh (Kolb and VanLopik 1958), and more than 90 percent of the chenier plain is characterized by intratidal marsh. Technically, the two are different origins but otherwise are essentially identical. Distributaries form the bounding limits of the environment in the deltaic plain, whereas these features are absent in the chenier plain. Although relatively few deposits remain in the sedimentary record, there can be no doubt that interdistributary deposits (and marshes) were the characteristic landscape of former delta lobes that have been destroyed during transgressive phases.

With interest primarily on the characteristics of the resulting deposits, most earth scientists accept a simple classification of interdistributary marsh based on the salinity of the environment in which the marsh vegetation grows. This classification includes freshwater marsh, floating marsh (also called flotant), brackish-freshwater marsh, and saline-brackish water marsh (Kolb and Van-Lopik 1958). Minor variations in this nomenclature can be found from one publication to another based on authors' preferences. On the other hand, biologists and ecologists typically classify the marsh according to the predominance of certain key marsh species (of which about 30 are present) and may recognize as many as nine types (O'Neil 1949) (Figures 34A and 34B).

Because of variations in interpretations and the rapidly changing state of salinities in coastal Louisiana, interdistributary marshes are not classified in Plates 12 through 14. Freshwater marshes occur only in the inlandmost parts of interdistributary depressions: in contrast, saline marshes occur only within about 10 mi of the shores of the coastal bays and sounds.

Interdistributary marshes form and persist only because of an interdependent relationship and favorable balance between inorganic sedimentation and vegetative growth (and organic sedimentation). During delta lobe progradation, enough inorganic sediments must enter interdistributary bays from overbank flooding to form intratidal mudflats (Figure 28). These are rapidly colonized by marsh vegetation (usually beginning with freshwater species) which then helps trap sediment from future flood events. Because of the intratidal setting, inorganic sedimentation rates seldom are high enough to convert the marsh to more terrestrial communities (e.g., such as on natural levees). Conversely, inorganic sedimentation rates often decline to the point where the buildup of organic detritus is insufficient to keep pace with subsidence, and marshes are replaced with shallow ponds or lakes or encroaching bays (Coleman and Gagliano 1964). Prior to this happening, the marsh vegetation at a

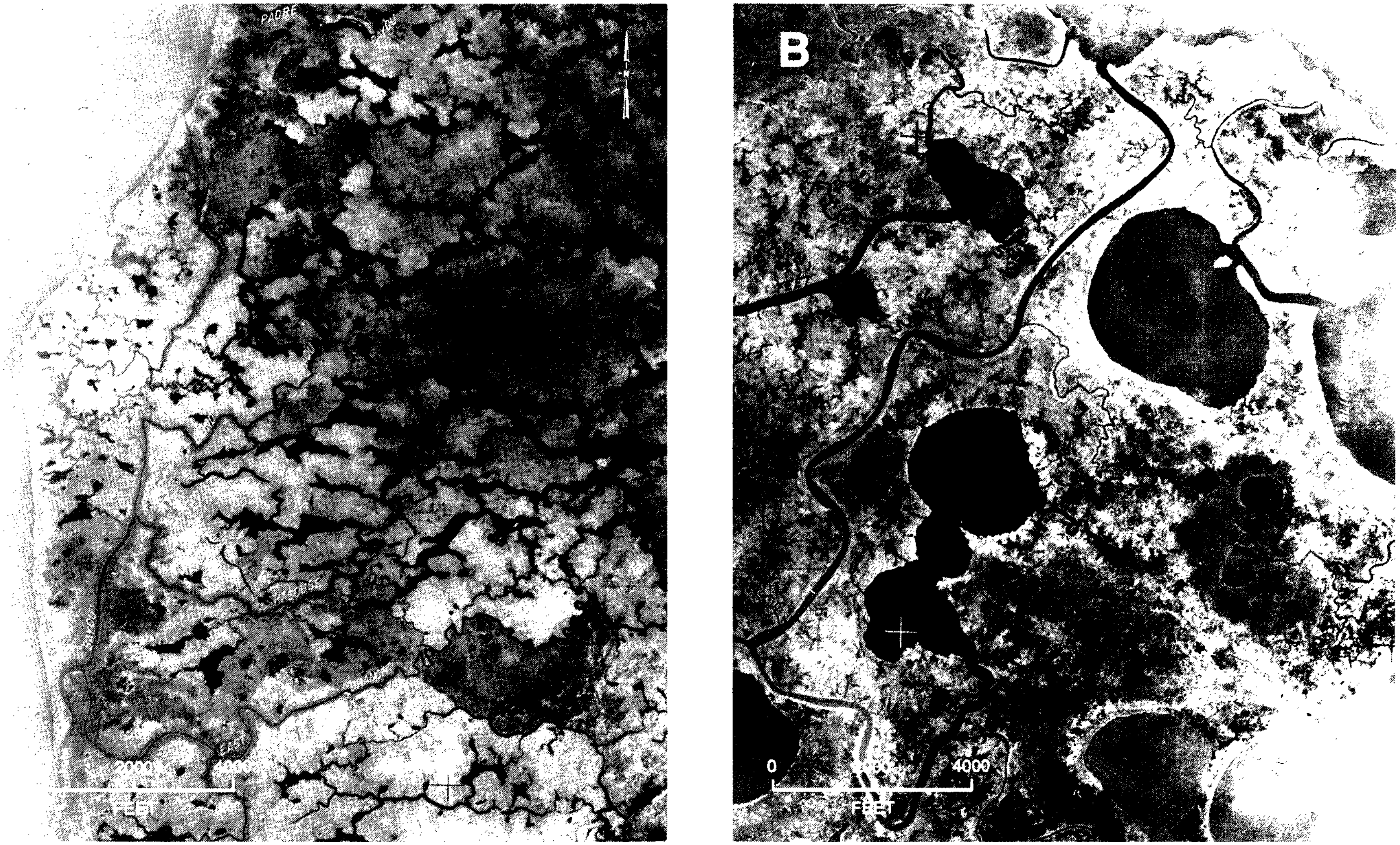


Figure 34. Intratidal marsh. A: Brackish marsh east of Lake Borgne in St. Bernard Ph., LA (Plate 14) characterized predominantly by three-cornered grass (*Scirpus olneyi*); B: Well-drained salt marsh along Mississippi Sound in St. Bernard Ph., LA (Plate 12) composed mainly of black rush (*Juncus roemerianus*) and wiregrass (*Spartina patens*)

given point typically will undergo a change from fresh to brackish and from brackish to saline plant communities. This situation prevails during the destructional or deterioration phase of a delta lobe.

In the chenier plain area, inorganic sediments have not been derived directly from overbank flooding from the Mississippi River but rather from turbid water carried westward along the Gulf Coast by longshore currents from the mouth(s) of the Mississippi River. This has led to the periodic development of mudflats and their subsequent colonization by marsh vegetation, a process that has been active during historic times (Morgan, VanLopik, and Nichols 1953). Otherwise, some sediments available to maintain the intratidal elevation of marshes are derived from materials carried inland during coastal, tropical storms. It should be remembered that the chenier plain experiences a much lower subsidence rate than the deltaic plain; therefore much less inorganic sediment is needed to maintain a stable environment. Although large lakes occur in the chenier plain (Plate 13), they are considerably older and much more stable than ones of comparable size in the deltaic plain.

In the older delta complexes, sedimentation in interdistributary bays appears to have been derived primarily from extensive sheet flooding since well-defined crevasse splays are not numerous. This mode of sedimentation is to be expected since the heights of floods decrease progressively downstream and reach zero at the river mouth. However, crevasses do occur during major floods and have been the principal factor in creating intratidal marshes in the Modern delta complex. Essentially all of the marsh between the several major present passes has been created as the direct result of six major crevasses, four of which are known to have occurred and to have been active between 1838 and the present (Coleman and Gagliano 1964). Well documented by historic maps, each of the splays has a radiating, bifurcating pattern of small distributaries (Plate 14) and has filled tens of square miles of shallow interdistributary bays (Figure 35A). Each undergoes a cycle of progradation and deterioration that is analogous to that of a delta lobe but of much shorter duration, i.e., 100 to 150 years (Coleman 1988).

Under the constant presence of subsidence, the bay-fill cycles caused by the crevasses eventually result in a characteristic vertically stacked but laterally offset sedimentary sequence in a delta lobe (Figure 32). Each cycle includes a lower, coarsening-upward sequence during the progradational phase and an upper, fining-upward sequence during the deterioration phase that terminates in an intratidal marsh environment. The delta front (intradelta) and prodelta environments that characterize the progradational portion of each cycle, and which are also present at the delta-complex scale, are discussed below.

Floating marsh, also referred to as "flotant," is a distinctive type of interdistributary marsh that occurs in areas that are almost exclusively fresh water (Figure 35B). It occurs mainly just seaward of the inland swamp environment in Terrebonne Parish and the upper part of the Barataria Basin (Kolb and Van-Lopik 1958). This marsh type consists of a 6- to 12-in.-thick floating mat of living and dead organic material held together by intertwined plant roots. The

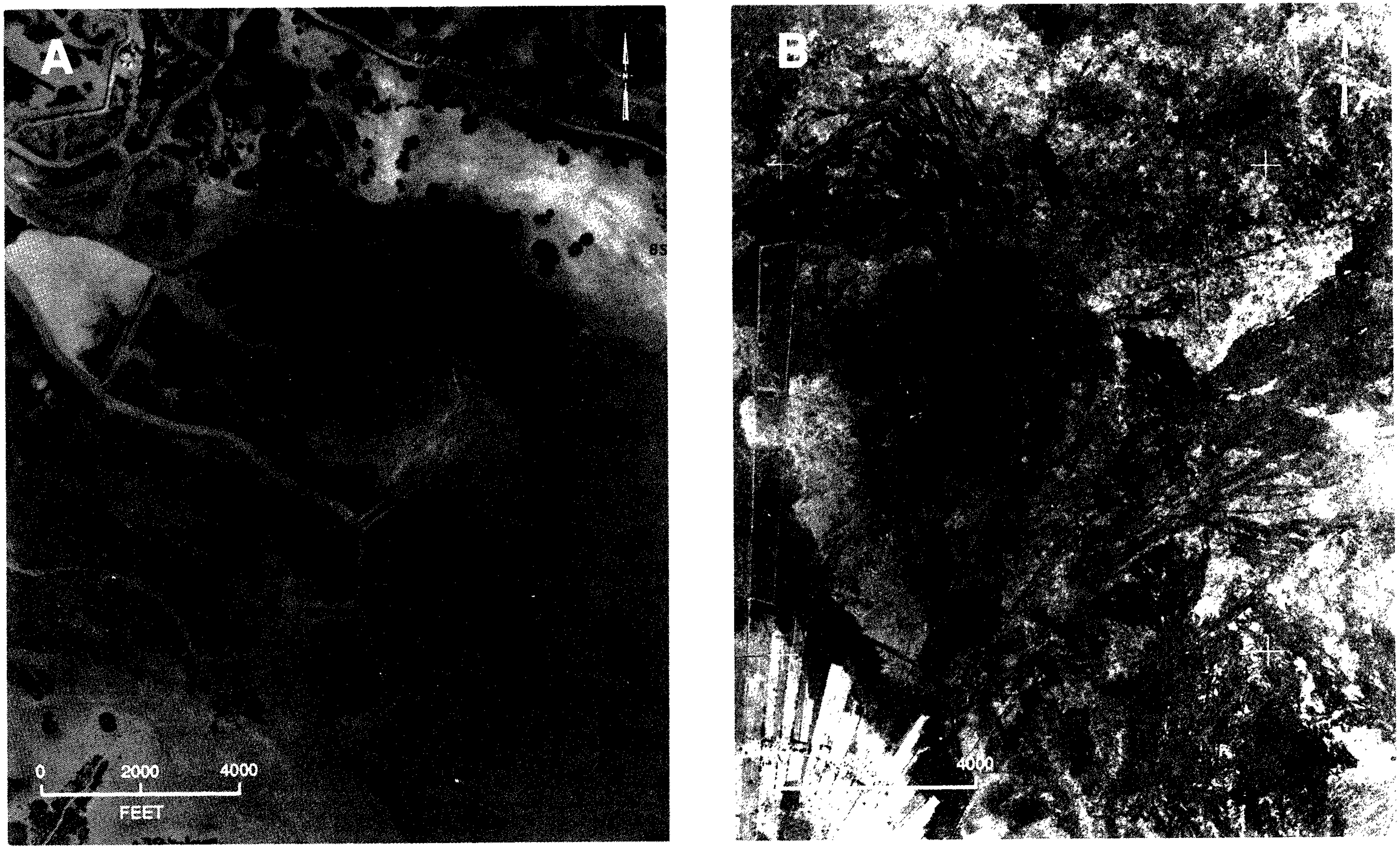


Figure 35. Intratidal marsh. A: Freshwater marsh, dominated by cattail (*Typha spp.*) and roseau cane (*Phragmites communis*), created after 1891 by a crevasse splay in Garden Island Bay in Plaquemines Ph., LA (Plate 14); B: Nearly uninterrupted tract of floating fresh marsh west of Lake Salvador in Lafourche Ph., LA (Plate 14) (tonal patterns on marsh caused by episodes of deliberate burning to enhance marsh productivity)

water on which the mat floats is generally several feet deep and beneath this is several feet of organic ooze or muck that eventually grades downward into a soft to medium gray clay. In some areas, the surface organic mat rises and falls with tidal and other water level changes: in other areas, it is periodically submerged. Factors that lead to the formation of floating marsh are poorly understood, and it is not known if it represents an early stage in the formation of an attached marsh or a late stage in which the attached marsh is deteriorating because of subsidence or some other process.

Delta front (intradelta) deposits

As a newly formed deltaic distributary advances seaward into a body of water, the relatively coarse-grained sediments carried by the stream as bed load are deposited at its mouth because of an abrupt reduction in current velocity (Fisk et al. 1954). Deposition results in shoaling and the formation of a distributary mouth bar (Figure 28A). In a relatively deepwater situation such as the Modern Complex, the bar that advances ahead of and is cut through and overridden by the distributary eventually forms a narrow, linear, lens-shaped (in cross section) coarse-grained unit known as a distributary mouth bar or a bar-finger sand (Fisk 1961) (Figure 28B). Occasionally, a distributary will split around a bar rather than cut through it: when this happens, the channel bifurcates and a new distributary branch will form. Both will then be underlain by a bar-finger sand as they develop seaward. In a relatively shallow-water situation such as the older delta lobes, waves and currents rapidly disperse the distributary mouth bar and distribute the coarse-grained material as a thin sand sheet rather than a narrow, linear mass (Figure 30).

Irrespective of the geometry of the sedimentary mass, the deposits constitute a recognizable lithologic unit of engineering significance. From this point of view, Kolb and VanLopik (1958) referred to the predominantly coarse-grained deposits as *intradelta*. However, with geomorphic processes more in mind, sedimentologists use a broader definition and use the term *delta front* environment (Coleman and Gagliano 1964). In turn, they divide it into four subenvironments, including distal bar, distributary mouth bar, channel, and subaqueous levee. The first three constitute the intradelta environment. However, by including the latter subenvironment, they include by definition some finer grained deposits that are gradational into the subaerial natural levee environment and which were excluded by Kolb and VanLopik (1958). Nomenclature or definition aside, the deposits represent subaqueous environments that have no surface expression in the deltaic plain.

Prodelta deposits

As shown in Figures 30 and 31, delta front deposits do not directly overlie marine deposits of the continental shelf environment. Rather, they are separated by a unit of fine-grained sediments (clays and silts) of appreciable thickness known as prodelta deposits. Prior to the progradation of a distributary to a given point and the formation of a distributary mouth bar, fine sediments flocculate from buoyant, expanding plumes of turbid water that are carried seaward from the river mouth and dispersed by waves and currents (Coleman 1988). Prodelta deposits consequently occur as a broad fan of fluvial sediments that accumulate slowly in a deepwater environment, giving rise to a sedimentary sequence that is probably the most homogeneous, widespread, and continuous of all deltaic environments (Kolb and VanLopik 1958).

It has been determined from oceanographic studies that the vast majority of prodelta deposits are laid down within 5 to 10 mi of the present mouth of the Mississippi River (Kolb and Kaufman 1967). Deposits were probably laid down to much greater distances ahead of other advancing delta lobes or distributaries when the river was discharging into much shallower water. This situation, plus the cumulative pattern of distributary development, has resulted in prodelta deposits being present at the base of the Holocene sequence throughout nearly all of the deltaic plain. Greatest thicknesses, varying from 50 to 100 ft or more, occur in a seaward-thickening wedge south and east of New Orleans and especially beneath the Modern Complex where they have an estimated volume of about 100 cu mi.

Mudlumps

Mudlumps are submerged mounds or low, irregular, ephemeral, emergent islands of plastic clay, varying from less than an acre to several tens of acres in size, that occur only at the mouths (passes) of the present Mississippi River. Although of extremely limited numbers (perhaps less than 200) and geographic distribution, and considered by some to be primarily a geological curiosity, they actually have had historic engineering significance in terms of navigation at the river mouth.

As a consequence of intensive investigations at the mouth of South Pass where mudlumps are best developed, they have been found to be spine-like, near-surface expressions of diapiric (intrusive) folds of prodelta clay (Morgan 1961; Morgan, Coleman, and Gagliano 1963). They are the direct result of the deformation and uplift of masses of prodelta deposits because of the deposition of large amounts of overlying delta front deposits (mostly distributary mouth bar deposits). Because of overburden pressure, clay is squeezed upward within a few thousand feet of the distributary mouth with spines remaining as islands for several decades before being destroyed by wave action or buried by delta front deposits. Uplift is episodic, with most movement occurring during flood stages and active bar growth, when as much as to to 15 ft of uplift has been observed (Coleman 1988).

Adding to the curiosity of mudlumps is the presence of occasional mud vents and cones. These are pseudo-volcanic features up to several feet high, resembling shield volcanoes, that are caused by the buildup of fluid mud that is extruded to the surface. They occur either on mudlumps or closeby and attest to the considerabe hydrostatic and gas pressures that exist within the upper several hundred feet of the deltaic sequence.

Mudlumps at the river mouth are not shown on Plate 14 because of their small size. Because their formation requires thick prodelta deposits overlain by thick, rapidly accumulating delta front deposits, they probably are a phenomenon related only to the lower part of the birdfoot delta (the Modern Complex).

Deltaic-Marine Environments

Depositional environments associated with a deteriorating delta lobe or complex have not received nearly as much attention as those associated with a prograding system, perhaps because of their much more restricted areal distribution (of land, but not of water). They were first investigated as analogues to petroleum reservoirs and to provide geotechnical data for offshore platform design and construction. Recent interest in the impacts of a potential acceleration in the rate of sea level rise on southern Louisiana has drawn some increased attention to these environments and the development of a widely accepted model (Penland and Boyd 1981; Penland, Boyd, and Suter 1988) that serves to complement the following discussions (see Figure 36).

Bay-sound deposits

Shallow lakes, bays, and sounds are present in all stages of a deltaic cycle, but they become large and numerous in the transgressive (deteriorating) phase when interdistributary marsh gives way to unvegetated shallow water at an accelerating pace. They first become most abundant at the seaward margin of a delta lobe and then increase inland and coalesce: as the deterioration process continues, the water bodies become deeper and more saline. At least during early stages of delta lobe deterioration (Figure 36, Stages 1 and 2), barrier islands are present to partially shelter the more seaward bays and sounds from open marine conditions in the Gulf of Mexico.

Bottom sediments of bays and sounds are perhaps the most variable and unpredictable of any deltaic environment. This is because the sediments include not only materials that eroded and winnowed from destroyed interdistributary marsh (and occasionally even distributary natural levees and channels) but also preexisting deposits that simply have been planed off and exposed at the bottom of the water bodies with no reworked materials present (an erosional surface or ravinement in stratigraphic nomenclature). Deposits resulting from winnowing consist mostly of silts and sands, but *in situ* deposits of organic clays of the interdistributary environment may characterize large areas. Coarsest deposits originating from winnowing and wave and current transport generally occur farthest seaward and may reach a thickness of 10 to 15 ft in relatively deepwater areas.

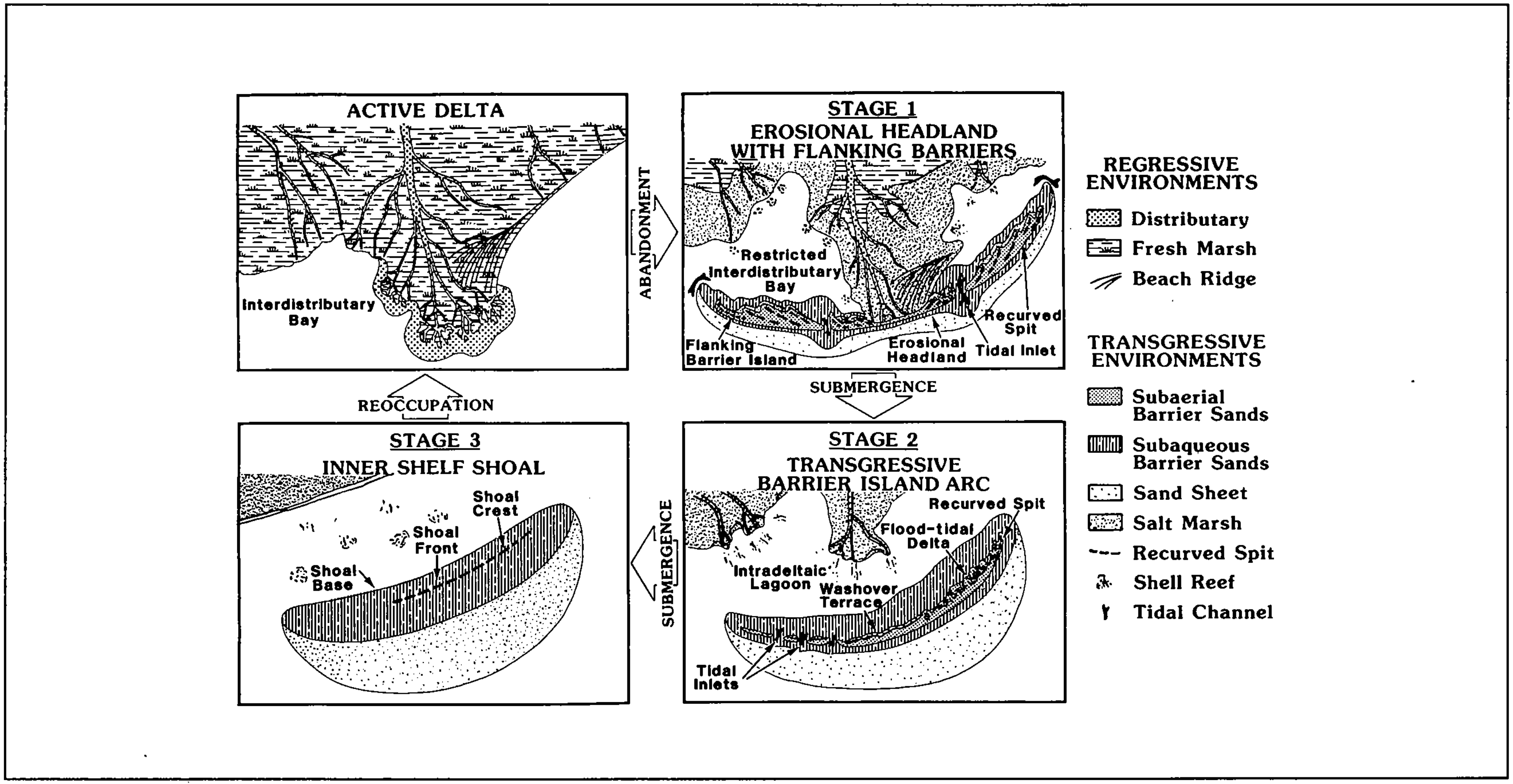


Figure 36. Model showing the distribution of environments at several transgressive stages in a deteriorating delta lobe (from Penland and Boyd 1981)

The bottoms of bays and sounds typically are extremely flat at an average depth of 8 to 12 ft (Kolb and VanLopik 1958). Shoal areas do occur, however, and mostly reflect the presence of the submerged remnants of a distributary natural levee ridge and associated channel that are being planed off by wave erosion. Reefs and shell deposits are also characteristic of the bay-sound environment and are discussed below.

Beaches/barriers

Fine-grained sediments (clays and silts) dominate the upper and outer parts of a deteriorating delta lobe or complex. Nevertheless, there is enough coarser grained materials (sands) present so that after winnowing, waves and currents can concentrate them into beaches and barrier islands along the Gulf shoreline. The sediments are derived from the direct erosion of headlands but also from materials flushed out of bays and sounds by tidal currents and distributed by longshore currents. Some sediments are also added from the erosion of offshore areas during coastal storms. The inorganic sediments are augmented by large quantities of shells and shell fragments from a large variety of estuarine and marine organisms.

The two stages of barrier formation, illustrated by Stages 1 and 2 in Figure 36, reflect differences in the degree of delta deterioration that can be observed in the present Mississippi River deltaic plain. The first stage, referred to as an erosional headland (Figure 37A) with flanking barriers (Figure 37B), is characteristic of a rather young system and is exemplified by the outer margin of the Lafourche Complex in southern Terrebonne and Jefferson parishes (Plate 14). The second stage, referred to as a transgressive barrier island arc, is characteristic of an older system and illustrates conditions at the edge of the St. Bernard Complex in St. Bernard Parish (Plate 14).

In Stage 1, sediments reworked and transported by marine processes accumulate along and override a headland composed of deteriorating interdistributary marsh. The beach at the headland migrates slowly inland while the flanking barrier islands and spits both migrate inland and laterally under the influence of longshore drift. Accretion ridges on the islands reflect the direction and rate of migration. Tidal inlets, many originating from breaches formed during storms, increase in number with time and eventually transform the arc into a series of discrete islands. Under the continuing presence of subsidence and erosion and under a progressively declining sediment supply, the islands rapidly decrease in size. In the Timbalier Island and Isles Dernieres areas of the Lafourche Complex, the barrier islands have dramatically declined in area during the historic period (as documented by maps) and have migrated inland at rates varying from about 25 to more than 50 ft/year (Williams, Penland, and Sallenger 1992).

In Stage 2, the barrier island arc becomes completely detached from the mainland, and areas of interdistributary marsh are no longer present as "core" areas in any of the islands. Otherwise, inlets open and close in response to

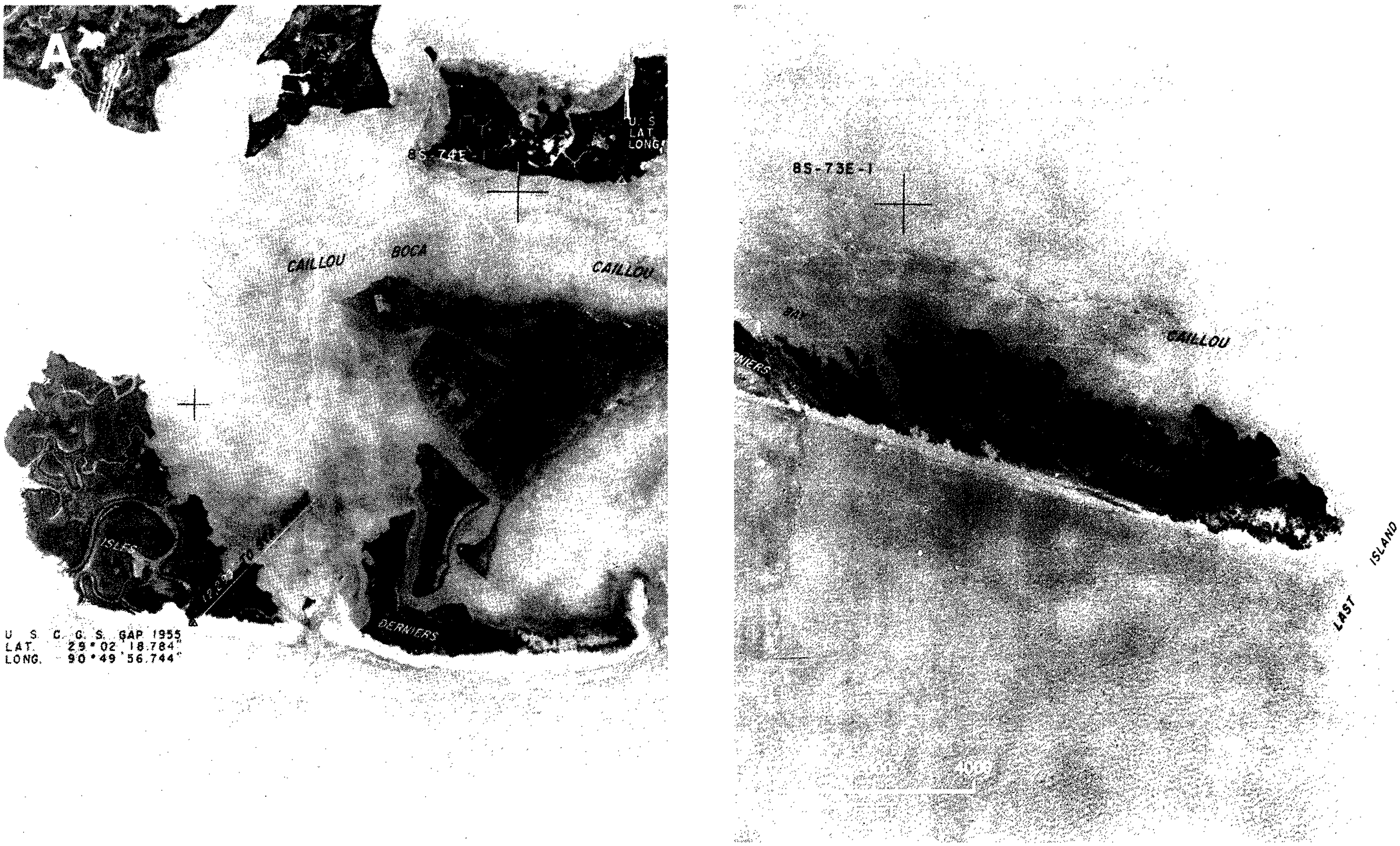


Figure 37. Beaches and barriers. A: Erosional headland in the Isles Derniers area of southern Terrebonne Ph., LA (Plate 14) (intratidal marshes represent surviving portions of the Lafourche delta complex); B: Portion of a flanking barrier island of the Isles Derniers chain

coastal storms, and some sediments move laterally in response to longshore drift. As in Stage 1, the island system migrates landward and diminishes in size and extent. The Chandeleur Island arc (Plate 14) exemplifies this stage and is known to be retreating at a rate of 13 to 18 ft/year (Williams, Penland, and Sallenger 1992).

In the most advanced stage of delta deterioration (Figure 36, Stage 3), barrier islands are completely destroyed by erosion, leaving only a submerged shoal. Because of the presence of large quantities of sand in the shoal, they persist as a prominent offshore bathymetric feature and sand deposit long after the subaerial parts of the delta lobe or complex are completely destroyed. Ship Shoal and Trinity Shoal offshore from central Louisiana (Autin et al. 1991) remain as evidence of the former outer margin of the Maringouin Complex (Figure 31).

One of the more important contributions to the knowledge of deltaic plain stratigraphy of the last several decades has been the recognition that the final stage of a delta cycle results in a broad, regional, marine erosional surface (ravinement surface) formed on older deltaic deposits (Penland 1990, Penland et al. 1991b). At that stage, Gulf waters have transgressed well inland, probably in direct response to a period of rather rapid sea level rise. Penland (1990) has postulated that as a result of the deterioration of the Teche Complex, a Teche ravinement surface formed and extended inland to a hypothetical Teche shoreline. He designated the deltaic deposits that underlie the ravinement surface as the late Holocene delta plain and those that overlie it (formed by the St. Bernard and Lafourche complexes) as the modern delta plain (not to be confused with the Modern Complex). While there is evidence from both seismic surveys and borings for the ravinement surface, there is no unequivocal evidence for the existence of a Teche shoreline. However, if Penland's model is correct, no barrier islands would be present to mark the shoreline since they would have been destroyed by erosion. Consequently, all deltaic plain barrier islands known to exist are associated with the present shoreline: there are none buried in the Holocene sedimentary sequence.

Two major buried barrier island trends do occur beneath the northern part of the deltaic plain in the Lake Pontchartrain Basin area. However, these formed prior to the development of the first Holocene delta lobe in the area and are analogous in origin, morphology, and age to those along the present Gulf Coast of Mississippi and Alabama. These trends, referred to as the Pine Island and Milton's Island trends (Saucier 1963, 1977), are discussed in detail in Chapter 7.

Regardless of the stage of the barrier islands, all that are attributable to the deterioration of a delta complex are low, narrow features generally less than a mile wide and typically not over 10 ft high. A low dune ridge is usually present behind the active beach, and often there is a small area of saline marsh and/or washover fans on the sound side. Although barrier islands are shown in Plate 14, no attempt has been made to delineate beaches elsewhere along the Gulf shoreline where they lie adjacent to interdistributary or intratidal marshes.

The latter generally are too narrow and too often changing because of storms to warrant delineation. It should be noted, however, that the beaches normally are bordered by a "sea rim" or "high marsh" a few feet above sea level that gently slopes inland over a distance of a mile or so. This feature is caused by the nourishment of the marsh by sediments washed inland during storms.

There are no buried, relict deltaic plain barrier islands, but there are several known beach trends at shallow depths beneath the deltaic plain. The more significant of these are well known and occur northwest of Lake Penchant in Terrebonne Parish, northeast of the Grand Terre Islands in Plaquemines Parish, and south of and parallel to the south shore of Lake Pontchartrain in Orleans Parish (Plate 14) (May et al. 1984, Kolb and VanLopik 1958, Saucier 1963). In each case, the trends can be followed for at least 5 mi. They have been located and identified largely on the basis of distinctive silt and sand deposits 10 ft or more thick mixed with unusually large amounts of oyster and clam shells and shell fragments. In fact, in the case of the trend in Terrebonne Parish, the shell content is so high that it is possible that a significant portion of the beach trend may actually consist of archeological sites (shell middens), especially in view of the presence of several known sites and artifacts (Weinstein and Kelley 1992). It is possible that the Terrebonne Parish trend may represent the Teche shoreline of Penland; however, this writer believes the high content of clam shells (mostly the brackish-water species *Rangia cuneata*) argues that the beach formed along a large lake or sound rather than the Gulf shoreline. This same situation is interpreted to be the case for the beaches in Orleans Parish. It is hypothesized that a rather sudden influx of fresh water (the development of a new distributary?) or some other ecological change caused the abrupt death of vast numbers of clams in a shallow deltaic plain lake or bay. Subsequently, wave action concentrated the shells along the shoreline into a beach of extraordinary size. As an analogy, a mass mortality of *Rangia* and the formation of a large shell beach is known to have occurred in the southwestern corner of Lake Pontchartrain several years after the 1952 opening of the Bonnet Carre Spillway.

Reefs

Thousands of years ago, prehistoric inhabitants of the deltaic plain recognized that the lakes, bays, and sounds of the interdistributary lowlands were prolific producers of clams, especially *Rangia*. Based on the hundreds of shell middens that are present, the inhabitants obviously exploited them in large quantities. More recently, these shallow water bodies have sustained a significant shell dredging industry for several decades. Irrespective of their great abundance, however, *Rangia* clams do not produce reefs *per se*. Rather, the organisms live dispersed in the soft clays and silts of the water bodies and become part of bay-sound deposits.

True reefs have been constructed in south Louisiana only by the oyster (*Crassostrea virginica*). This organism grows best on firm bottoms in shallow, brackish to saline water where new generations often attach themselves to older living oysters or to dead shells in the intratidal or subtidal zone. Where favorable conditions persist, and in the presence of subsidence, oysters can eventually create reefs many feet in thickness (Figure 38). The linear nature and orientation of many reefs suggest that submerged distributary natural levees have been a favored substrate for oyster colonization (Coleman and Gagliano 1964).

Borings have encountered cemented masses of oyster shells (relict reefs) at numerous locations in the deltaic plain sedimentary sequence west of the Mississippi River and especially in southern and western Terrebonne Parish (Kolb and VanLopik 1958). Reefs in the sedimentary record indicate the locations of former bays and sounds. The general absence of reef deposits east of the Mississippi River in the St. Bernard and Modern complexes is not readily explainable but must be related to an absence of favorable ecological conditions. Small, live oyster reefs occur widely in the more saline bays and sounds of the present deltaic plain, but none approach the magnitude of the former Point au Fer reef (Plate 13), a several-thousand-foot-wide feature that, until recently, extended about 20 mi across the mouth of Atchafalaya Bay. In the last few decades, shell dredging has essentially destroyed this 5- to 10-ft-thick reef.

Cheniers

As elements of both the cultural and natural landscapes, cheniers are to the chenier plain essentially what distributary natural levee ridges are to the deltaic plain. In both cases, they constitute the only relatively high ground in the midst of large expanses of marsh and are conspicuous because of their forest vegetation (Figure 39). Since prehistoric times, they have served as the principal sites for permanent habitation and transportation.

Howe et al. (1935) first described and mapped cheniers in southwestern Louisiana and recognized their mode of origin, but it was not until more than 20 years later that they were the subject of comprehensive geologic studies and a detailed chronostratigraphic model was developed (Byrne, LeRoy, and Riley 1959; Gould and McFarlan 1959). Penland and Suter (1989) have recently suggested a slightly modified model based on a different interpretation of the effects of sea level rise, but the basic concept of formation set forth more than 50 years ago remains unchallenged.

Cheniers are associated with many of the world's major deltas (Augustinus 1989) and are intimately related to periodic shifts in river mouths (delta lobe or complex formation). In the case of the Mississippi delta, extensive mudflat formation and shoreline progradation occurred in the chenier plain area when the river mouth was relatively close by (e.g., the Maringouin and Teche complexes), and large quantities of fine-grained sediment were carried westward by longshore currents. That was a time of extensive intratidal marsh development. When the river mouth shifted eastward (e.g., the St. Bernard and Lafourche complexes), the supply of sediment to the chenier plain area

Figure 38. Reefs. Extensive oyster reefs extending southward into the Gulf of Mexico from Marsh Island in Iberia Ph., LA (Plate 13)

Figure 39. Cheniers. Truncated series of relict beaches, indicating former Gulf shorelines, forming Pecan Island, a large chenier in Vermilion Ph., LA (Plate 13)

diminished to the extent that marine, erosional processes became dominant and the shoreline retreated inland. Wave-winnowing of the intratidal marsh deposits and the subsequent concentration of the coarser sediments and faunal remains formed mainland beaches and spits, closing off the mouths of bays and overriding the marsh at the sea rim. When a new cycle of mudflat formation began, the beaches and spits were left stranded as relict features, becoming cheniers.

Cheniers associated with the Mississippi delta extend from just west of Vermilion Bay westward into extreme southeastern Texas: those in the eastern half of this zone are shown in Plate 13. Although there have been changes in shoreline orientation along the coast and the truncation of older cheniers by younger ones, such as on Pecan Island, there are four basic trends that have been recognized (Gould and McFarlan 1959) and are most evident southwest of Grand Lake. These are designated, from north to south, the Little Chenier, High Island, Creole Ridge, and Oak Grove Ridge trends. A detailed model illustrating events in the formation of the trends and the distribution of subsurface sedimentary facies is shown in Figure 40.

Each chenier ridge consists of a long, ribbon-like mass of silt, sand, and shell that is biconvex in cross section, 5 to 10 ft thick, several hundred feet wide, and up to several tens of miles long. The ridges are smooth on their seaward side, irregular on their landward side, and rise from a few inches to more than 10 ft above the surrounding marsh (Byrne, LeRoy, and Riley 1959). They directly overlie a thin prism of gulf-bottom deposits and are flanked by intratidal marsh deposits (Figure 40). The oldest (farthest inland) cheniers appear to be slightly less than 3,000 years old (based on radiocarbon dates and archeological evidence), and all have formed only since sea level has been at its approximate present level. Cheniers that may have formed earlier at times of lower sea level would have been located offshore and have been destroyed by erosion during the last stages of postglacial sea level rise.

Nearshore Gulf deposits

Apparently a term coined by Kolb and VanLopik (1958), deposits of this depositional environment generally are equivalent to materials variously referred to as gulf-bottom, strand plain, or shelf deposits. They have been recognized and mapped beneath the St. Bernard Complex (e.g., Kolb and VanLopik 1958, Kolb 1962, May et al. 1984), directly overlying the buried, subaerially eroded surface of the Prairie Complex and underlying prodelta, intradelta, and/or interdistributary deposits. Beneath the Plaquemines-Modern Complex, they overlie a prism of early Holocene delta front sediments that, in turn, overlies the Prairie Complex (Coleman and Gagliano 1964, Dunbar et al. 1994).

Nearshore Gulf deposits are directly associated with the postglacial rise in sea level (the Holocene transgression) and, in the area mentioned, date from about 12,000 to 5,000 years BP. They include not only a heterogeneous

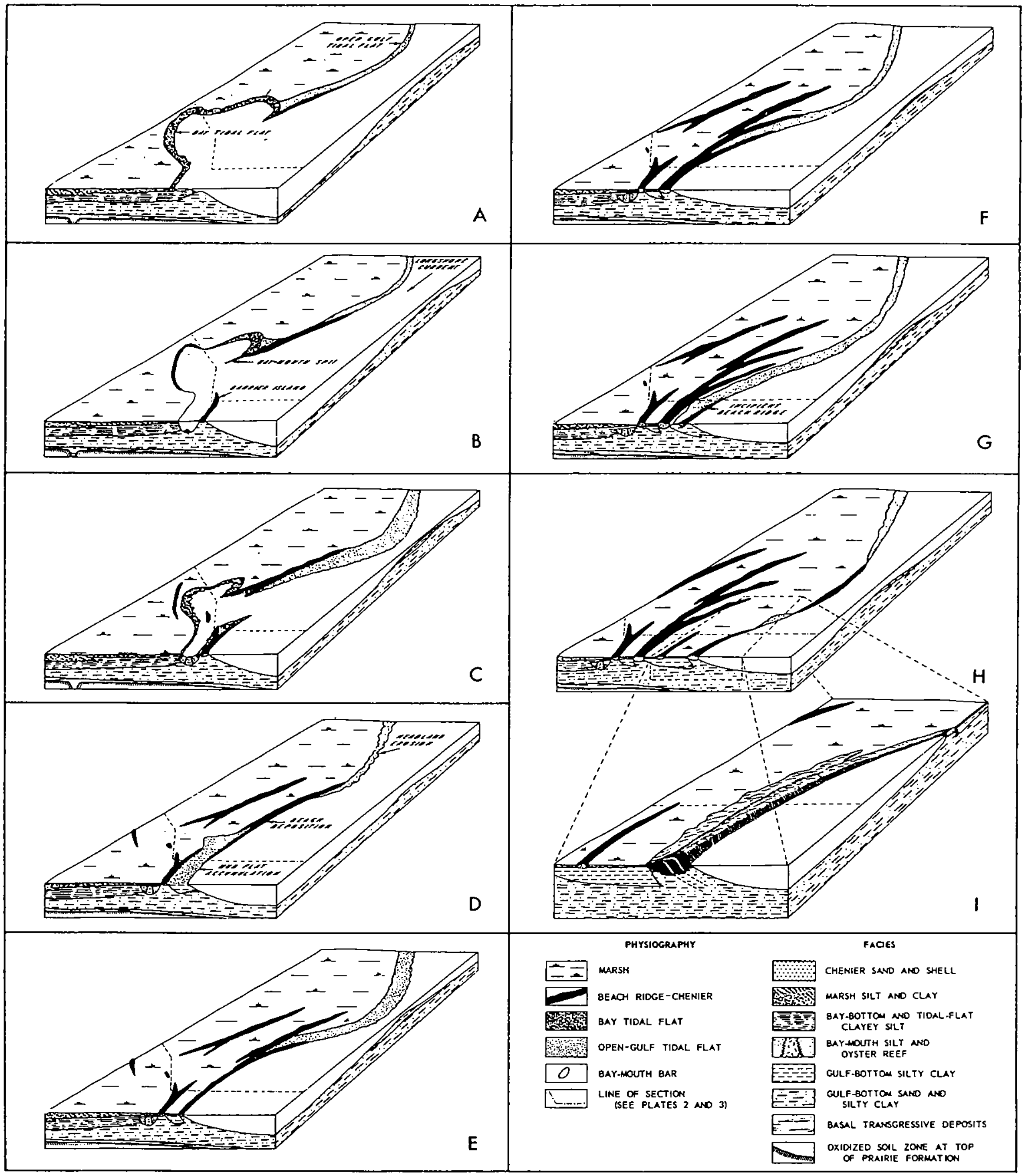


Figure 40. Model of chenier development showing surface physiography and subsurface sedimentary facies (from Gould and McFarlan 1959)

mixture of materials primarily derived from erosion and winnowing of the underlying Prairie Complex surface, but also sediments carried by longshore currents from sources east of the Mississippi River, and perhaps a small amount of Mississippi River sediments. By definition, nearshore Gulf deposits include those materials deposited in relatively shallow water (50 to 100 ft) seaward from the Gulf shoreline before the first Mississippi River delta complex was manifest.

At any given location, nearshore Gulf deposits involve a sedimentary sequence that reflects a series of events directly related to the transgression of the sea. The deepest deposits, directly overlying the eroded Pleistocene Prairie Complex surface, typically include a sand-shell hash-organic matter mixture that represents deposition immediately behind, beneath, or just seaward of a transgressing shoreline/beach complex. This is the true stand plain portion of the sedimentary sequence. As the transgressing shoreline moves inland from its previous location, water depths increase, wave- and current-energy levels decline, and finer grained sediments (i.e. clays and silts) became more prevalent. Typically, the deposits include a rich, diagnostic faunal assemblage that can be very helpful in identifying this environment (Parker 1956). Sedimentation rates in the open Gulf waters remained rather low until the first Mississippi River delta lobes advanced into the area and the nearshore Gulf environment changed to the prodelta environment.

In cross section, the nearshore Gulf sedimentary sequence is a seaward-thickening, irregular, and discontinuous layer that is thickest over channels and valleys on the underlying erosional surface and thinnest over topographic highs. Thicknesses vary from a few feet to a maximum of about 40 ft: in some locations, only a thin, sandy shell hash layer is present immediately above the Pleistocene surface. This layer may incorporate peat, wood, or other organic debris from the organic-rich surface soil horizon (representing a marsh, swamp, or forest environment) across which the shoreline transgressed.

From a foundation engineering viewpoint, there are few considerations in the deltaic plain area more important than determining the depth to the top of the weathered and eroded Prairie Complex (commonly just referred to as the "top of the Pleistocene"). This is often, but not always, possible because of appreciable differences in sediment color, lithology, internal structures, and geotechnical properties between the Pleistocene sediments and the overlying Holocene sediments. This writer has found that where the differences are absent, subtle, or ambiguous, the top of the Pleistocene often can be more easily determined indirectly by noting the characteristically well-defined base of the nearshore Gulf deposits. Quite often this marks the deepest occurrence of shell-rich deposits in the shallow sedimentary sequence.

Unusual Features and Deposits

Pimple mounds

Should one ask this writer if there is a truly perplexing, enigmatic aspect of Lower Mississippi Valley Quaternary geomorphology that has defied concerted efforts at explanation and for which there is no consensus, he would immediately respond that the origin of pimple mounds clearly meet these criteria. Pimple mounds are insignificant features from the standpoints of engineering geology and regional chronostratigraphy, but they deserve brief discussion herein for at least two reasons. First, discussions of their origin and characteristics have generated a voluminous literature dating back more than 100 years. Second, the mounds are a phenomenon of such curiosity and notoriety that no treatise would be complete without their mention.

Pimple mounds, also referred to as prairie mounds and prairie blisters, and sometimes incorrectly as mima mounds, are a common occurrence in several parts of the western and central United States. One contiguous area occurs in eastern Texas and Oklahoma, southern Missouri, most of Arkansas, and western and southern Louisiana (Knechtel 1952). In the latter state, they are most abundant in the Coastal Plain province and especially in the Great Southwest Prairie area. Totally, there are probably tens of millions of individual mounds in the area. A typical pimple mound is a low, rounded, circular- or elliptical-shaped hillock about 2 to 3 ft high and 50 ft in diameter. They typically occur as discrete features either uniformly scattered across the landscape or in clusters: where well developed, there will be from 100 to 300 per square mile. They occur on shallow slopes and flat areas but normally are not present in creek bottoms or poorly drained areas. Because they are slightly better drained and often slightly coarser textured than the surrounding deposits, pimple mounds are clearly evident on aerial photos as light-toned "dots" against a darker background (Figures 41A and 41B).

The published literature contains mention of at least 27 different theories of origin that have been proposed and defended (Holland, Hough, and Murray 1952). No one concept, however, has been able to satisfactorily explain all observable characteristics of their composition, distribution, and morphology. Some of the more popular concepts include ant hills, burrowing animals, differential erosion, gas pressure, Indian mounds, sand dunes, and uprooted trees. Assuming as virtually all geomorphologists do that all pimple mounds are of common origin, an eolian origin, for example, cannot be accepted when many mounds are composed of plastic clay rather than silt or sand. Similarly, tree uprooting cannot explain mounds in prairie areas, and differential erosion is unlikely to have occurred in completely flat areas. Despite the wide appeal (because of modern analogs) of an origin by burrowing rodents, there have never been skeletal remains or traces of rodent burrows actually found in the features.

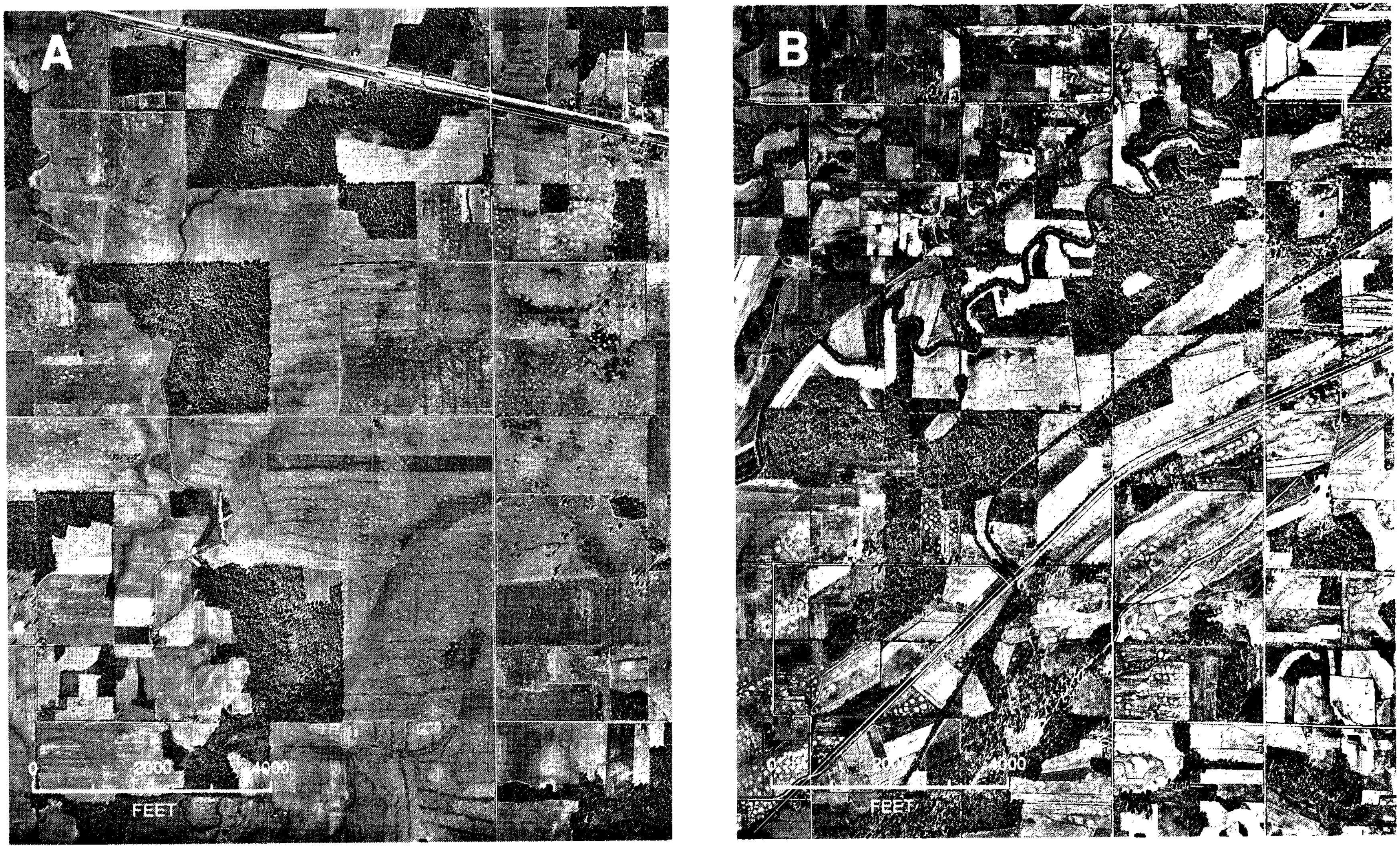


Figure 41. Pimple mounds. A: Widely scattered, small pimple mounds on silty clay soils (Early Wisconsin valley train deposits) east of Bald Knob in White Co., AR (Plate 6); B: Pimple mounds of moderate size and density on sandy loam soils (Early Wisconsin valley train deposits) southwest of Naylor in Ripley Co., MO (Plate 5)

Part of the problem of developing an acceptable hypothesis is the tremendous range of conditions under which pimple mounds exist. In the central United States, they occur on upland formations ranging in age from at least Cretaceous to the Quaternary as well as on Quaternary terraces of fluvial and marine origin. They occur on soils and deposits that vary from hard, plastic clays to loose, dense sands. They occur not only in areas long characterized by prairie vegetation but also in densely forested tracts. They occur on deposits representing glacial outwash (valley train), coastal sand flats, and meander belt environments of deposition (Krinitzsky 1949b).

Excluding the Prairie and Intermediate complexes of southwestern Louisiana where they are dramatically evident (Holland, Hough, and Murray 1952; Varvaro 1957), pimple mounds are present but not abundant in the Lower Mississippi Valley area. As a result, they have not been as widely investigated as they have been elsewhere. Much of the more recent literature has focused on their relationship with archeological sites (Aten and Bollich 1981; Jones and Shuman 1988; O'Brien, Lyman, and Holland 1989). Despite the limited occurrence, however, this writer believes the distribution of pimple mounds in he alluvial valley area might hold some vital clues as to their origin. In another way, there are critical aspects of their distribution in the alluvial valley area that have not been, but must be, satisfactorily explained if an hypothesis is to be viable.

In Arkansas and Missouri, pimple mounds are relatively more abundant on the Ozark Plateau and steadily decline in numbers and density eastward to the Mississippi River. As has been pointed out (Saucier 1974, 1978), perhaps the most interesting aspect of pimple mound distribution is the acknowledged fact that irrespective of landform age, morphology, or lithology, not a single mound exists anywhere *east* of the Mississippi River. Certainly some previously proposed causal processes such as wind or surface erosion did not stop at the banks of the river. This writer has also observed that almost without exception, pimple mounds occur as single, discrete mounds that do not touch each other. Further, an evaluation of stratigraphic and archeological evidence shows that no pimple mound formation (by whatever cause) has occurred more recently than about 5,000 years BP (Saucier 1978; O'Brien, Lyman, and Holland 1989).

This writer has become progressively more convinced that all observable evidence favors pimple mounds being the result of either ant or termite colonies. This is by no means an original idea, having been first proposed by Hilgard (1873) and more recently entertained as a possible explanation for mima-like mounds in South Africa (Cox, Lovegrove, and Siegfried 1987). The ubiquitous physical separation of the mounds in the Lower Mississippi Valley area suggests territoriality by an organism; presence on a variety of soil types and terrain conditions suggests a biological origin; absence in poorly drained areas suggests a burrowing organism; and absence of skeletal remains suggests an invertebrate.

It is speculated that because of the slightly warmer and drier climate of the Altithermal, ant or termite populations expanded in the eastern portion of the Great Plains area and into Missouri, Arkansas, and Louisiana. The organism then spread rapidly eastward out of the uplands into the alluvial valley. Several thousand years later, because of a return to cooler and moister conditions (or because of some biological factor such as disease or predation), populations rapidly declined just before the eastern limit of the organism's range reached the Mississippi River.

Sand blows

In the St. Francis Basin of Arkansas and Missouri, there are countless numbers of low, roughly circular, sandy mounds of about the same size as pimple mounds that extend almost uninterruptedly over an area of about 4,000 sq mi (Fuller 1912, Saucier 1977a, Obermeier et al. 1990). They occur on both valley train and meander belt deposits and sometimes have been confused with pimple mounds.

Extensive investigations have proven beyond question that they are sand blows caused by the liquefaction of shallow subsurface sands during the New Madrid earthquake series of 1811-1812 and possibly a few preceding seismic events. In contrast to pimple mounds, the sand blows are more irregular in shape and spacing, typically coalesce and resemble beads on a string rather than discrete, single mounds (Figures 42A and 42B). Readers are cautioned that the distributions of pimple mounds and sand blows (and indeed also sand dunes) do overlap in the upper part of the St. Francis Basin, and the features can be hard to differentiate using only aerial photo interpretation. Sand blows and related liquefaction features are discussed at length in Chapter 8 of this synthesis.

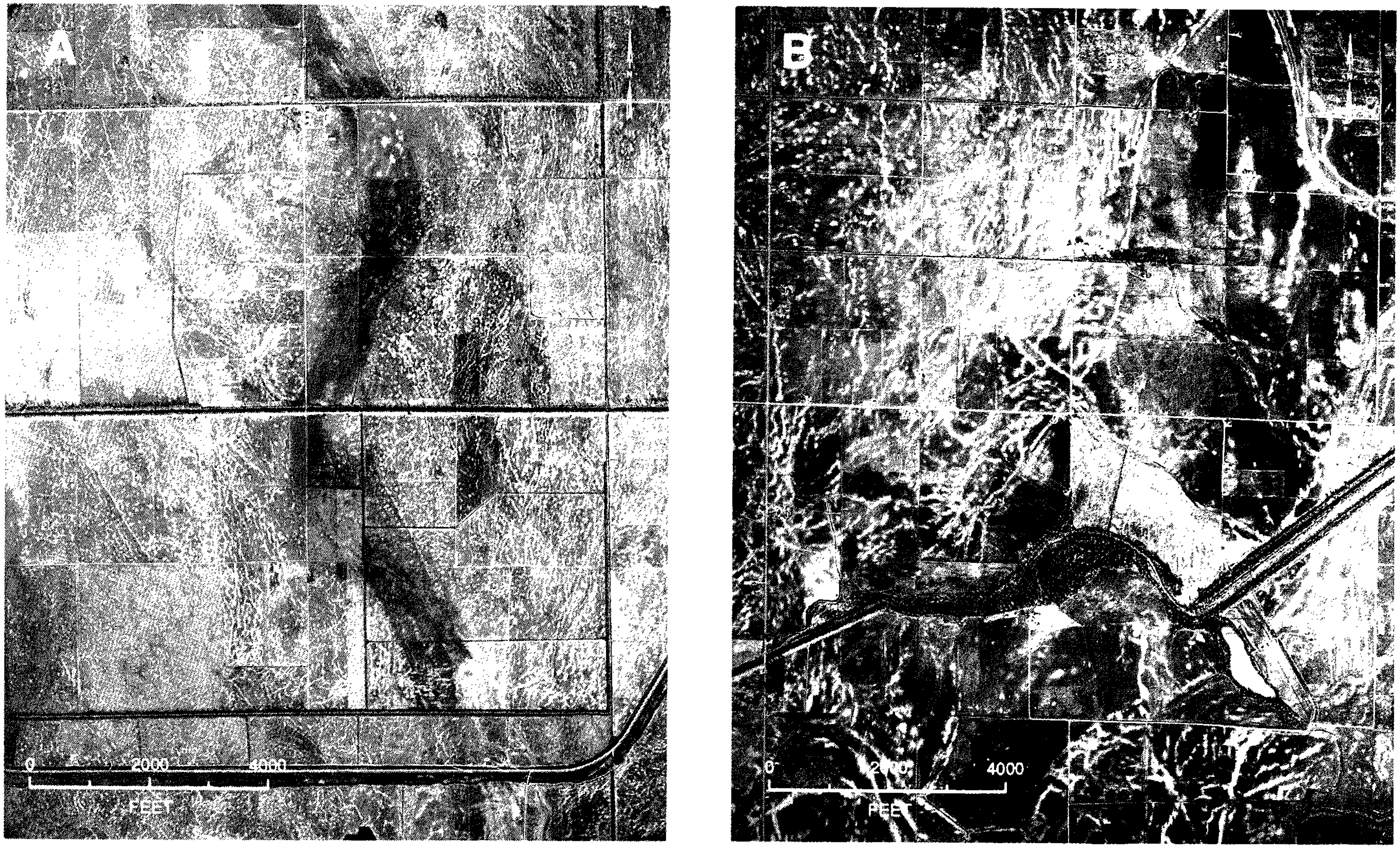


Figure 42. Sand blows and fissures. A: Dense scatter of relatively undisturbed sand blows on Late Wisconsin valley train deposits near Rivervale in Poinsett Co., AR (Plate 6); B: Sand blows and fissures well diffused by cultivation on Late Wisconsin valley train deposits southeast of Lepanto in Poinsett Co., AR (Plate 6)

6 Lithology, Soils, and Geotechnical Properties

Discussions in this chapter are intended to provide a general description of the lithology and physical and sedimentary characteristics of the major landforms and depositional environments for those unfamiliar with the area. Since many of the units have highly similar and nondiagnostic lithologies and overlapping ranges of properties, emphasis is on characteristics such as color, bedding, internal structures, inclusions and other more significant recognition criteria. The discussions include an indication of the range of characteristics that occur throughout the area of occurrence as well as at any given location.

Information on geotechnical properties are restricted because of data availability to the shallow Pleistocene (Prairie Complex) deposits beneath the deltaic plain, the Wisconsin-age valley train deposits, and the Holocene alluvial and deltaic environments. Furthermore, since specific test results are often nondiagnostic or nondefinitive in themselves, the focus of the discussions is on parameters and ranges of values that can be used either to help determine the origin and depositional environment of a particular tested sample or to eliminate certain ones from further consideration. On the other hand, a knowledge of the typical range of geotechnical properties of a given environment, coupled with insight into its likely vertical and horizontal extent and continuity based on the geomorphic processes involved, can assist in designing exploration and testing programs.

To avoid a significant interruption in the continuity of the text, the results of the statistical analysis of the large database of geotechnical characteristics are presented in entirety in Appendix A. The figures presented therein graphically portray the ranges of values of specific geotechnical parameters (e.g., water content or dry density) for the major depositional environments for which information is available. They also portray, for comparative purposes, the ranges of values for a given property as they vary from one environment to another.

Because the ranges of variation among most of the parameters are similar and overlap to a significant extent, the discussions in the text focus on the two that have proven to be most diagnostic as far as differences in environments are concerned, i.e., water content and cohesive strength. These are also perhaps the most important parameters from an engineering point of view.

To the fullest extent possible, descriptions of the lithology of deposits make reference to both the widely accepted texture classes of the U.S. Department of Agriculture (1938) soil classification system and the soil groups of the Unified Soil Classification System (USCS), which is in universal use within the Department of the Army (1960). The former is most widely used by soil scientists and geologists because of its more descriptive nature while the latter is the standard in the engineering profession because of its basis in the geotechnical properties of soils. Following convention, USCS group symbols such as (CH) and (SM) are indicated in parentheses where applicable. The USCS is used exclusively in the discussion and illustration of data on soil geotechnical characteristics.

Tertiary and Older Uplands

It is beyond the scope of this synthesis to present detailed descriptions of formations or geological units older than the Quaternary. General characteristics are presented in Table 2: for more detailed information, readers are referred to summaries such as those of Boswell, Moore, and MacCary (1965), Cushing, Boswell, and Hosman (1964), Hosman, Long, and Lambert (1968), and Murray (1961) and the extensive bibliographies contained therein.

In overview, several generalizations are possible. As has been mentioned (Chapter 4), truly lithified deposits occur in certain Tertiary formations and characterize all of the Paleozoic formations, whereas they are essentially absent in those of Quaternary age. Where rock or cemented sediments *per se* are absent in Tertiary and older formations, the deposits typically are much more indurated than they are in any comparable depositional environment of Quaternary age. Cohesive sediments have much higher average strengths and noncohesive sediments have much higher average densities, reflecting the effects of much longer periods of weathering and consolidation.

As another generalization, Tertiary formations are devoid of gravel-sized materials since sources of such coarse clastics were not present in the region. Moreover, in the deltaic, estuarine, and shallow marine environments in which the sediments were deposited, energy levels were insufficient to transport sediments coarser than sand. Therefore, geologists in the Lower Mississippi Valley area are always alert to determining the base of a graveliferous sedimentary sequence, either in the uplands or beneath the alluvial plain, as this often marks the unconformity separating Quaternary from underlying Tertiary units.

Pleistocene Terrace Complexes

Upland complex

Deposits typical of the Lafayette and Citronelle gravels, herein designated the Upland Complex, include sandy gravels (GW and GP), muddy (clayey) sandy gravels (GC), silty sands (SM), and muddy gravelly sand (Self 1993). In sequences up to 100 ft or more thick, individual beds are a few feet thick and are highly lenticular and discontinuous. Beds of clay (CH or CL) or silt (ML) are uncommon, but fine-grained soil zones up to a few feet thick are present in Louisiana, and paleosols have been identified where the graveliferous deposits are overlain by loess (Guccione, Prior, and Rutledge 1990).

The gravel fraction of the deposits consists of well-rounded particles ranging from granules to cobbles in size, but with most being of pebble size. Pebbles typically constitute at least 50 percent (by weight) of the deposits and average 1 to 3 in. in size in the northern part of the area and seldom over 1 in. in the southern part (Campbell 1971). Chert and secondarily quartz are by far the predominant materials involved. Bedding of the gravel is indistinct, but the sand layers vary from massive to cross-bedded with distinct planar or broad trough cross-stratification. In the northern part of the area, the sands are coarse grained and poorly sorted (SW) but tend to become moderately sorted and of medium grain size (SP) toward the south (Self 1993). Cut and fill structures and scoured channels are abundant in all areas (Potter 1955b).

Besides the presence of gravels, the Upland Complex is conspicuous in outcrops or cores by its highly oxidized nature. Reflecting extensive iron oxide staining and cementation, its color typically varies from a uniform dark red, orange-red, or pink to a bright yellow brown. Thin, irregular and wavy layers (less than 1 in. thick) of hematite and/or limonite are rather common. Ripup clasts and balls of purplish-red and whitish mud are present in some exposures.

As discussed in Chapter 5, the presence of igneous and metamorphic pebbles (glacial debris from the Canadian Shield area) and apparently ice-rafted erratics in some graveliferous deposits argues that they may be of glacial origin rather than an alluvial fan derived from the east. In gravel pits in western Mississippi and on Crowley's Ridge, this writer has observed thousands of cobbles of 6 in. or larger size and boulders of chert and quartzite up to 2,000 lb in weight scattered throughout the deposits. Similar boulders have been observed in central Louisiana (Woodward and Gueno 1941). It is postulated that masses of coarse-grained glacial outwash occur as inliers within the broad alluvial fan that constitutes the bulk of the Upland Complex, but there are no discernible sedimentologic differences that would allow differentiation of the deposits according to the two different sources. Careful mapping of the distribution of the erratics could be fruitful in this regard.

Outcrops of the Upland Complex constitute the only source of "clay gravel" (GC), an unusual association of grain sizes that makes an excellent road subbase and that is in widespread use (and demand) throughout the Lower Mississippi Valley area. The origin of this particular type of deposit has not been investigated but probably represents the product of weathering and breakdown of chert to form the fine-grained matrix rather than primary deposition. The high degree of weathering of chert gravels in the Upland Complex of southeastern Louisiana has been noted (Campbell 1971, Cullinan 1969) and interpreted as an indicator of the considerable antiquity (i.e., at least Early Quaternary age) of the deposits.

Information on selected characteristics of soils developed on deposits of the Upland Complex has been compiled only for Louisiana (Autin et al. 1991). About 15 different soil series have been mapped in *in situ* surface deposits, and data are presented in Table 3 for five of them that represent a lithosequence with decreasing content of silica sand in the parent material. Each is a Ultisol with a distinct A and E horizon overlying a reddish Bt horizon. The sequence demonstrates a progressive decrease in total A + E horizon thickness and an increase in the clay content in the B horizon.

Intermediate complex

The principal published descriptions of the Intermediate Complex are of outcrops of the Montgomery terrace along the lower Red River (Fisk 1940) and are not representative of conditions flanking the Mississippi alluvial valley. Otherwise, very little is known, both by way of outcrops and borings, of the lithology of these deposits, and it is not even possible to identify the actual environments of deposition that were present. It appears safe only to conclude that much of the complex consists of alluvium deposited as a broad alluvial apron by small streams draining the adjacent higher terraces and uplands. The nebulous term "undifferentiated Coastal Plain deposits" has been occasionally used to describe the deposits and regrettably cannot be improved upon.

In the very few exposures and shallow borings that are available in southwestern and southeastern Louisiana, the upper several tens of feet of the deposits consist of stiff, well-oxidized, red, brown or buff, massive silty clays (CL) or loams (SC) (Mossa and Autin 1989, Varvaro 1957). Calcareous, phosphatic, and limonitic nodules of pea-gravel size are typically abundant. Water well data suggest that below depths of about 50 ft, the upper fine-grained deposits grade downward into sands (SP) and eventually sands and gravels (GP) (e.g., see Turkey Creek quadrangle in Smith and Russ 1974). Hence, the complex grossly resembles a typical coarsening-downward alluvial sequence.

Based on data from southeastern and southwestern Louisiana, soils of the Intermediate Complex are Ultisols and highly weathered Alfisols (Table 3). Decreases in elevation, sand content, and depth and seasonal duration of the water table all cause an increase in clay content (Autin et al. 1991). These

TABLE 3.
SELECTED CHARACTERISTICS OF SOILS DEVELOPED IN DEPOSITS OF VARIOUS GEOLOGIC UNITS IN LOUISIANA (FROM AUTIN ET AL. 1991)

A. IN-SITU DEPOSITS OF THE UPLAND COMPLEX

Soil Series	Alaga	Eustis	Lucy	McLaurin	Ruston
Classification	Typic Quartzipsamment	Psammentic Paleudult	Arenic Paleudult	Typic Paleudult	Typic Paleudult
Setting	Gently to moderately sloping convex interfluves	Gently to moderately sloping convex interfluves	Gently to moderately sloping convex interfluves	Gently to moderately sloping convex interfluves	Gently to moderately sloping convex interfluves
Solum thickness range (cm)	10 to 25	>150	>150	>150	>150
Typical horizon sequence	A-C *or* A-B-C	A-E-Bt-BC-C	A-E-Bt-BC-C	A-E-Bw-Bt-B/E-Bt-BC-C	A-E-Bt-B/E-Bt-BC-C
Clay Content[1] (%)	<18	<18	20 to 30	10 to 18	18 to 30

[1]Range for average of upper 50 cm of argillic horizon in Ultisols; for 25 to 100 cm zone in Entisol.

B. COAST-PARALLEL INTERMEDIATE COMPLEX

Soil Series	Ruston	Malbis	Beauregard	Caddo
Classification	Typic Paleudult	Plinthic Paleudult	Plinthaquic Paleudult	Typic Glossaqualf
Setting	Gently sloping convex interfluves	Gently sloping convex side-slopes	Gently sloping sideslopes	Nearly level to level surfaces
Bt horizon thickness (cm) (Modal Pedon)	205	165	135	75
Typical Horizon Sequence	A-E-Bt-B/E-Bt-BC-C	A-E-Bt-Btv-BC-C	A-E-Bw-Bt-Btv-Btg-Cg	A-E-Btg-Cg
Clay Content[1]	18 to 35	18 to 35	18 to 35	18 to 35

[1]Range for average of upper 50 cm of argillic horizon.

C. COAST-PARALLEL PRAIRIE COMPLEX IN SOUTHEASTERN LOUISIANA

Soil Series	Prentiss	Stough	Myatt	Abita	Brimstone	Guyton
Classification	Glossic Fragiudult	Fragiaquic Paleudult	Typic Ochraquult	Glossaquic Paleudalf	Glossic Natraqualf	Typic Glossaqualf
Setting	Convex ridges with <5% slopes	Convex ridges with <5% slopes	Nearly level broad stream terraces	Broad flat stream terraces	Broad flat stream terraces	Broad flat stream terraces
Solum thickness range (cm)	>150	>150	100 to 150	150 to 200+	100 to 200+	125 to 200+
Typical horizon sequence	A-E-Bw-Bx-C	A-E-E/B-Bt-BC-C	A-Eg-Btg-Cg	A-E-B/E-Bt-Btg-BCg-Cg	A-E-E/B-Btg-Btng-BCg-Cg	A-Eg-B/E-Btg-BCg-Cg
Clay content*(%)	12 to 18	<18	18 to 35	18 to 35	18 to 35	18 to 35

*Range for average of upper 50 cm of argillic horizon.

(Continued)

TABLE 3. (Concluded)

D. DEWEYVILLE COMPLEX							
	Calcasieu River				Ouachita River		
Soil Series	**Bienville**	**Cahaba**	**Wrightsville**	**Guyton**	**Haggerty**	**Mollicy**	**Groom**
Classification	Psammentic Paleudalf	Typic Hapludult	Typic Glossaqualf	Typic Glossaqualf	Aeric Ochraquult	Aquic Hapludult	Aeric Ochraqualf
Setting	Gently sloping	Gently sloping	Nearly level to slight depressions	Nearly level to slight depressions	Nearly level	Gently sloping	Nearly level
Solum thickness range (cm)	150 to 200	90 to 150	100 to 183	125 to 203	64 to 140	<140	152 to 254
Typical horizon sequence	A-E-B/E-Bt-BC-C	A-E-E/B-Bt-BC-C	A-Eg-Btg/Eg-Btg-BC-C	A-Eg-B/E-Btg-BCg-Cg	A-Bt-BC-C	A-Bt-2Bt	A-Bg-Btg-BCg-Cg
Clay Content[1] (%)	<18	18 to 35	18 to 35	18 to 35	<18	18 to 35	18 to 35

[1]Range for average of upper 50 cm of argillic horizon.

E. SICILY ISLAND AND PEORIA LOESSES				
	Soils in thick loess deposits		Soils in thin loess deposits	
Soil Parent material	**Sicily Island Loess**	**Peoria Loess**	**Sicily Island Loess**	**Peoria Loess**
Soil Series	Evangeline	Olivier	Toula	Gigger
Classification	Glossic Paleudalf	Aquic Fragiudalf	Typic Fragiudult	Typic Fragiudalf
Solum thickness (cm)	295	157	165	335
Horizon sequence	A-E-Bw-Bt-B/E-Bt-C	A-E-Bt-Bx-Cc-C	A-E-Bt-2Bx-2Bt	A-E-Bt-2Bx-2Bt
Maximum clay content in Bt horizon (%)	38	24	33	28

F. MISSISSIPPI RIVER MEANDER BELT 1 DEPOSITS									
Soil Series	**Crevasse**	**Convent**	**Commerce**	**Mhoon**	**Kobel**	**Sharkey**	**Fausse**	**Barbary**	**Maurepas**
Classification	Typic Udipsamment	Aeric Fluvaquent	Aeric Fluvaquent	Typic Fluvaquent	Vertic Haplaquept	Vertic Haplaquept	Typic Fluvaquent	Typic Hydraquent	Typic Medisaprist
Setting	Inbank, levee crest	Levee crest	Levee crest and backslope	Levee backslope	Levee backslope, backswamp	Levee backslope, backswamp	Backswamp	Backswamp	Backswamp
Solum thickness range (cm)	10 to 25	10 to 25	50 to 100	50 to 125	75 to 150	90 to 150	60 to 125	0 to 25	125+
Typical Horizon Sequence	A-C	A-C	A-B-C	A-Bg-Cg	A-Bg-Cg	A-Bg-Cg	O-A-Bg-Cg	A-Cg	Oa1-Oa2
Clay content[1] (%)	0 to 18	0 to 18	18 to 35	18 to 35	35 to 60	>60	>60	>60	<30

[1]Range for average of 25 to 100 cm zone.

relationships account for the differences among the chief soil series shown in Table 3.

Prairie complex

As discussed in Chapter 5, some outcrop areas of the Prairie Complex (formerly designated the Prairie terrace) consist of as many as three stratigraphic units. The units range in age from pre-Wisconsin to Late Wisconsin and are separated by erosional unconformities. Moreover, thanks to the well-preserved surface morphology and numerous borings, it is possible to identify and delineate seven basic environments of deposition. Because of this complex origin, the lithology and sedimentary characteristics vary widely and must be discussed separately. The overall distribution of the Prairie Complex is shown in Plate 2, and the locations of six major depositional environments are shown in Figure 43. The seventh recognized environment, tributary valley fill, is not shown since it is beyond the areal scope of this synthesis.

In general terms, the gross lithology of each of the six depositional environments is the same as in their Holocene counterparts. Differences occur mainly in such parameters as consistency, color, and inclusions (e.g., nodules) because of the age of the Prairie Complex deposits and their considerably greater exposure to weathering, desiccation, and consolidation.

Valley train deposits. In a small (3- by 12-mi) area adjacent to the western side of the southernmost segment of Crowley's Ridge near Helena these deposits occur (Figure 43, Plate 7). No detailed borings are available in the area, but exposures in gullies indicate that beneath several feet of loess, the deposits consist of well-oxidized, buff to yellowish brown, stiff, massive silty and sandy clays (CL and SC). Nodules of various composition are abundant. At a depth of 10 to 15 ft beneath the surface, the deposits become considerably sandier and, while not exposed, probably grade into dense, massive, silty sands (SM), and sands (SP) within another 10 ft or so. Based on the stratigraphy of the Wisconsin-aged valley trains farther to the west, it is inferred that coarse sands and gravels (GP and GW) will occur about 50 ft beneath the surface and continue to the base of the unit at a depth of about 100 to 120 ft.

Within a mile of the western edge of Crowley's Ridge, the valley train surface is heavily veneered with younger alluvial fan/apron deposits consisting mostly of silts (CL and ML) eroded from loess deposits on the ridge. Immediately adjacent to the ridge, these deposits attain a thickness of about 50 ft but thin rapidly away from the ridge. In this narrow zone, it can be assumed that the underlying valley train deposits are interfingered with alluvial fan deposits since upland erosion must have also occurred during the time of outwash deposition.

Backswamp deposits. The Prairie Complex in three areas is characterized by these deposits: in the Grand Prairie region, along the western alluvial valley margin from the vicinity of Little Rock to south of Monroe, and in small

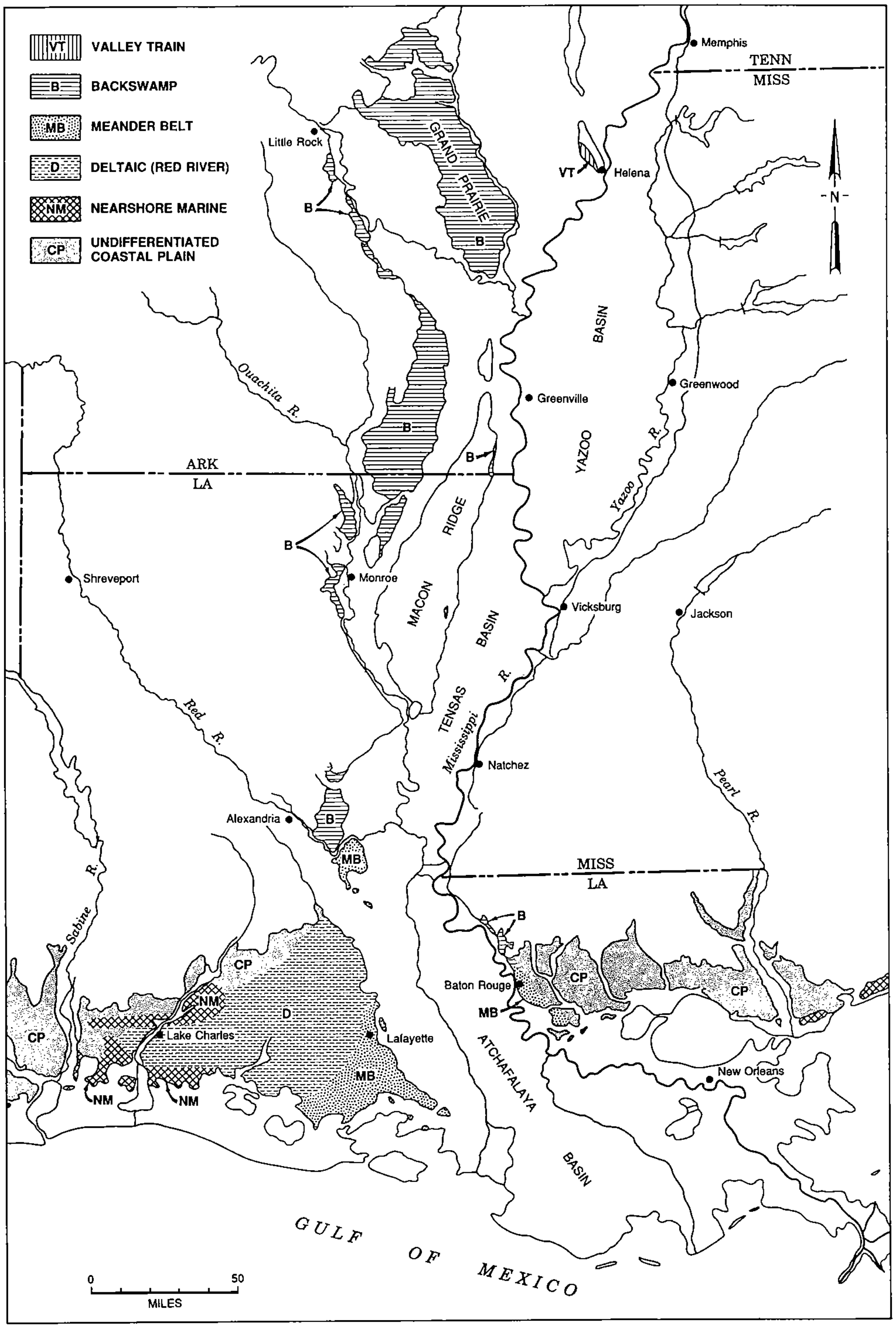


Figure 43. Major depositional environments that characterize the near-surface portions of the Prairie Complex

areas along the eastern valley margin north of Baton Rouge (Figure 43). In each area, there is a several-foot-thick veneer of loess and colluvial deposits from the adjacent uplands. The Grand Prairie region has an additional extensive, surficial layer of appreciable thickness that obscures the underlying backswamp deposits. This layer, averaging 15 to 20 ft thick, is a broad alluvial fan of the Arkansas River that exhibits hundreds of miles of abandoned meandering channels, natural levees, and point bar areas (Plate 7) (Saucier 1967, Smith and Saucier 1971). It unconformably overlies the backswamp deposits as a silty and sandy sheet, but the abandoned channel are incised into deposits and are filled with sand.

Turning to the backswamp deposits themselves, they consist of 60 to 80 ft of stiff to hard, well-oxidized, mottled dark gray to brown, massive clays (CH) and silty clays (CL). The clays incorporate scattered organic fragments and woody debris, show evidence of extensive biogenic reworking (root fills and worm burrows), often exhibit slickensided surfaces caused by desiccation and consolidation, and contain abundant calcareous nodules. In the area north of Baton Rouge where the deposits have been extensively cored and tested at various industrial sites (ENCOR, Inc. 1992), the backswamp sequence is interrupted by one or more erosional unconformities and paleosols. These reflect interruptions in the backswamp environment and changes in local base levels caused by glacial-cycle degradation and aggradation in the alluvial valley.

The sediments that constitute the backswamp deposits originated primarily from Mississippi River backwater flooding but, near the upland margin, are mixed with siltier and sandier deposits eroded from the uplands and carried into the flood basins by local creeks. Occasionally, the creeks extended their courses from the upland into and across the backswamp areas, leaving narrow entrenched channels that are evident in the backswamp sedimentary sequences by interrupting, narrow bands of muddy sands up to several tens of feet thick. It is also apparent that during the several interruptions in backswamp deposition when the alluvial valley area experienced degradation, dendritic gully systems analogous to those of the present (e.g., see Plate 8) extended themselves inland from the edge of the valley and are now evidenced by sandy valley fill sequences.

All areas of Prairie Complex backswamp deposits are immediately underlain by tens of feet of massive sands (SP) and sands and gravels (GW and GP) that extend downward to the Tertiary-age formations that constitute the suballuvial surface. No specific evidence is available to indicate the precise origin or depositional environment of the coarse-grained substratum; however, the bulk of the material must represent glacial outwash laid down by braided streams before the Mississippi River changed to a meandering regime during an earlier glacial cycle.

Meander belt deposits. The Prairie Complex in the Marksville Hills and the area to the northwest, in the immediate Baton Rouge area, and south of Lafayette in southern Louisiana is composed of these deposits (Figure 43). In each case, the sedimentary sequence and surface morphology of the terrace clearly indicates the presence of natural levees, point bar accretion topography, crevasses, and abandoned channels of Mississippi River size (Goodwin et al. 1991) (Plates 10, 11, and 13). The origin of the fluvial features is so obvious that since their first delineation using aerial photos, no one has questioned that they are remnants of Pleistocene-aged Mississippi River meander belts (Howe and Moresi 1933; Fisk 1940, 1948), and the area south of Lafayette was formally designated the Lafayette Meander Belt. The precise age and glacial-cycle relationships of the meander belt segments are problematic, however, and are discussed in Chapter 7.

Because so many fluvial environments are present in the meander belt segments, it is impossible to generalize regarding the lithology of the deposits. There are no borings of engineering quality that are available for evaluation, but it can be safely assumed from Holocene analogs that the areas of point bar accretion involve a thin silty and sandy clay (CL) topstratum underlain by 100 ft or more of sands (SM and SP) grading downward into sands and gravels (GW). These deposits should be stiff to hard in consistency and well-oxidized with buff, yellow, and brown coloration. Abandoned channels are known from surface soils to contain relatively soft, gray, organic clays (CH and OH), and these should be at least several tens of feet thick. Both point bar and abandoned channel deposits, as well as associated natural levee deposits, are overlain by a several-foot-thick veneer of loess.

It should be noted that some abandoned channels filled with fine-grained deposits have been detected on the surface of the Prairie Complex where it lies buried at a shallow depth beneath the intratidal deposits of the Chenier Plain east of White Lake (Plate 13) (VanLopik 1955). This is the seaward extension of the Lafayette Meander Belt where it trends from the outcrop area southward onto the continental shelf for some unknown distance.

Autin et al. (1988) recently have carefully examined, sampled, described, and reinterpreted the famous 86-ft-high exposure of the Prairie Complex along the Mississippi River bluff at Mt. Pleasant, LA., about 16 mi north of Baton Rouge (Plate 11). They concluded that the fluvial sequence (beneath 12 ft of loess) includes a natural levee deposit underlain by five thin (10- to 15-ft-thick) channel fill sequences, each consisting of silty clays (CL) that grade downward into sandy loams (SM and SC).

Based on extensive subsurface studies at a nearby industrial complex (ENCOR, Inc. 1992), this writer believes their interpretation of the deposits as Mississippi River meander belt facies needs to be reexamined. An interpretation of the sequence as crevasse splays with associated natural levees, rather than being a stacked sequence of channel fills (an entertained but dismissed possibility), would be more compatible with the sedimentary sequence at the industrial site. More than 200 closely-spaced deep borings at the site substantiate that the fluvial sequence below the natural levee deposits represents materials deposited in a rim swamp environment and interrupted only occasionally by the entrenchment and filling of small gullies. Nevertheless, Autin et al. (1988) do describe what appears to be a classical outcrop (Prairie Complex) of

Mississippi River point bar deposits about 1,300 ft south of the Mt. Pleasant Bluff locality. They describe the sedimentary sequence as consisting of 50 ft of fine sand to granule gravel (SW?) with large- and small-scale trough cross-stratifications, horizontal stratifications, scour surfaces, and clay balls. These deposits are overlain by 15 ft of fine-grained sediments representing swale fills and overbank deposition (natural levee?).

Deltaic deposits. The upper portion of the Prairie Complex over an extensive area between Lafayette and Lake Charles in southwestern Louisiana is characterized by these deposits of Red River origin (Figure 43, Plates 11 and 13). The terrace surface in this area is characterized by numerous, northeast-southwest trending streams (local drainage) that occupy relict, meandering, Red River deltaic distributaries that formed at a time when the Red River discharged directly into the Gulf of Mexico without becoming tributary to the Mississippi River. The distributaries, sometimes including cutoff meander loops, have low, broad natural levee ridges that are separated by interchannel lowlands (Goodwin et al. 1991). Erroneously referred to as backswamp areas (Fisk 1948), the latter are actually interdistributary lowlands that probably were once characterized by swamps or marshes.

The natural levee and interdistributary deposits have lost many of their original sedimentary characteristics and now consist of brown to reddish, firm to stiff, clays (CH) and silty clays (CL). Abandoned distributaries are filled with clayey silts (ML) and fine sands (SP) of comparable color that probably attain a maximum thickness of about 50 ft. Average overall thickness of the mass of deltaic deposits is not known but probably is about 20 to 30 ft. At least in the southern part of the area, the fluvial deposits overlie fossil-rich ones that were deposited in shallow marine, lagoonal, and estuarine environments. The latter are indicative of the coastal setting into which the Red River discharged and prograded during a time of relatively high sea level (see discussions in Chapter 7). Based on correlations with marine units to the east of the Lower Mississippi Valley, these deposits have been designated as part of the Biloxi formation (DuBar et al. 1991).

Nearshore marine deposits. These deposits are found in extreme southwestern Louisiana and along the Mississippi Gulf Coast and eastward into Florida (Figure 43). Although they occur as outcrops only beyond the area included in this synthesis, they are worthy of mention because portions of the trend lie buried beneath a portion of the deltaic plain in southeastern Louisiana.

In the outcrop areas, the deposits primarily include the Houston Ridge and other segments of the Ingleside Barrier Trend which is the several-mile-wide, Sangamon-age barrier/beach complex that extends along much of the Gulf Coast east and west of the Mississippi deltaic plain (Price 1933, Bernard and LeBlanc 1965). In the central Gulf Coast area, Otvos (1991) has used the term Gulfport formation to include these marine features.

Deposits of the barrier trend include oxidized, poorly to moderately sorted, fossiliferous shoreface sands overlain by well-sorted, cross-bedded foreshore sands and dunes. At the surface, the barrier ridges display coast-parallel accretion ridges with shallow, intervening swales.

Also included as nearshore marine deposits (Figure 43) are a series of poorly defined, subdued, east-west trending, small beach ridges that occur near the edge of the Prairie Complex surface just north of the intratidal marshes of the chenier plain (Saucier 1977b, Autin et al. 1991). These features lie on what was the shallow shelf seaward of the Ingleside Barrier Trend and consist of well-oxidized sandy loam (SM and SC) at least 12 to 15 ft thick. The deposits are only slightly sandier than the surrounding nearshore Gulf deposits that consist of fossiliferous silty and sandy clays. Although younger than the Ingleside trend, the age of these small marine features remains speculative and is discussed in Chapter 7.

West of the Pearl River in the Lake Pontchartrain Basin area, a westward continuation of the Ingleside trend occurs at a depth of 60 to 120 ft below sea level as a consequence of subsidence and faulting along the Baton Rouge Fault Zone. Numerous deep borings in the greater New Orleans area have penetrated into a several-mile-wide mass of marine sand at least 60 ft thick along a trend that roughly parallels the south shore of Lake Pontchartrain from the Rigolets to near the town of LaPlace, LA (Plates 12 and 14) (Kolb, Smith, and Silva 1975). The beach deposit, the deepest of three encountered in the New Orleans area, lies below the second of two strongly weathered horizons (paleosols) within the Prairie Complex and is interpreted as Sangamon in age. In turn, the shallower nearshore Gulf and beach deposits, which follow a similar trend and that underlie the first weathered horizon, are considered to be of Middle Wisconsin age but also a part of the Prairie Complex (Autin et al. 1991). In both cases, the beaches are massive units of fossiliferous, fine to medium sand (SP and SM), and the surrounding deposits are well-oxidized, stiff to hard, fossiliferous, irregularly interbedded clays, silts, and sands. Events leading to the formation of the two nearshore marine units beneath the deltaic plain are discussed in detail in Chapter 7.

Undifferentiated Coastal Plain deposits. The remainder of the Prairie Complex is made up of these deposits which are exposed over large areas in southwestern and southeastern Louisiana (Figure 43). In the Florida Parishes area, these deposits occur as two reasonably distinct sedimentary sequences. Below a depth of 20 to 30 ft, the lower unit consists of mixtures of fossiliferous silts and clays (CL, ML, and SC) deposited in a brackish-water environment. It is interpreted that they were laid down in a large sound or lagoon that existed between the Ingleside Barrier Trend and the mainland shoreline. These deposits probably correlate with Otvos's Biloxi formation (DuBar et al. 1991).

The upper 20 to 30 ft of the Prairie Complex in this area, as well as that of the area in southwestern Louisiana, consists of complexly interfingered, lenticular masses of fluvial clays, silts, and silty sands (CL, ML, and SM). They are interpreted as representing alluvial and colluvial deposits laid down by small streams and as slope wash from the Intermediate Complex and older formations to the north. They also include and merge with true fluvial terraces that extend inland along the larger of the local Coastal Plain streams such as the Amite and Tangipahoa rivers (Plate 12). These materials were initially deposited in and eventually filled the shallow sound or lagoon and transformed the area into a broad, gently sloping, terrestrial alluvial plain. The bulk of this sedimentary sequence is believed to be of Wisconsin age (Autin et al. 1991).

According to Autin et al. (1991), two distinct suites of surface soils characterize the undifferentiated Coastal Plain deposits of the Prairie Complex in southeastern Louisiana (Table 3). Alfisols occupy the entire southern part of the complex in this area and decrease in abundance northward. Ultisols predominante to the north and are associated with Alfisols developed in local topographic lows.

An interesting unresolved aspect of the geomorphology of the Prairie Complex of the Florida Parishes, analogous in some respects to the problem of pimple mounds, involves the origin of several generally north-south trends of sand hills that parallel some of the larger streams such as the Tickfaw and Tangipahoa rivers (Plate 12). The hills are conspicuous, isolated, rounded and elongate masses of silty or loamy sand (SM and ML) up to several hundred feet long that attain a height of as much as 20 to 30 ft above the surrounding surface. Otvos (1971), Gagliano (1963), and others have favored the hills being true eolian dunes, which they certainly resemble in terms of morphology, but Mossa and Autin (1989) present geomorphic and stratigraphic evidence that favors a fluvial origin. They believe the sand hills, which are capped with Peoria loess and a geosol, are inliers of older Pleistocene fluvial deposits that were dissected by stream action and largely buried by deposition of the upper Prairie Complex sediments.

Buried Pleistocene deposits. Located beneath the Holocene deltaic deposits these materials are lithologically best known in southeastern Louisiana, especially the New Orleans area, where they have been encountered in tens of thousands of borings and have been the subject of special investigations, notably those by Kolb and Van Lopik (1958) and Kolb, Smith, and Silva (1975). Although the composition of those Prairie Complex deposits is extremely well known and has been portrayed in numerous lithologic cross sections to depths of several tens of feet, no attempt has been made to differentiate them regionally according to environment of deposition. Other than for features like the buried segments of the Ingleside Barrier Trend, they are known only as undifferentiated deposits of nearshore marine or Coastal Plain origin.

Deposits of marine sands occasionally characterize the shallowest Prairie Complex deposits (directly below the first weathered horizon), but mostly they consist of stiff to hard, well-oxidized clays (CH) or silty clays (CL). Typical geotechnical characteristics of the deposits are summarized in Figure A1. Because of the effects of prolonged desiccation and preconsolidation, the deposits generally have average water contents that are considerably lower and cohesive strengths that are considerably higher than those of any of the overlying Holocene deposits (see Figures A13 and A19). Unlike in any of the Holocene environments, strengths of over 1.0 ton/sq ft are quite common. However, because a thin, organically enriched soil horizon or weathered zone is sometimes preserved as a true paleosol, lower strengths can be encountered immediately below the erosional surface and are reflected in the wide range of values shown in Figure A1.

Although strength and water content are the most diagnostic *physical* parameters in terms of recognizing buried Prairie Complex deposits, color often is of equal or greater value as a recognition criterion. Almost always the Holocene deposits overlying the surface are uniformly steel gray in color, whereas those of the Prairie Complex are either mottled gray, red, brown, and yellow or mottled green and light gray. The latter also show evidence of bioturbation, often display slickensided surfaces, and typically contain abundant ferruginous and/or calcareous nodules.

Deweyville complex

Considerable attention has been focused on the Deweyville Complex because of its unique surface morphology ("giant" meander scars considerably larger than those of the present rivers) (Gagliano and Thom 1967). However, relatively little is known about the lithology of the complex, especially in the Lower Mississippi Valley area. In its only principal outcrop area, which is along the western side of the Grand Prairie (Plate 7), it has been penetrated by a few borings (Saucier 1967), but there has been no sedimentological description of its deposits. What little that can be said comes by way of extrapolation from areas like the Ouachita River valley (Fleetwood 1969) (Plate 9) and the Sabine and Pearl river valleys.

Although of Arkansas River origin rather than the Mississippi River, the Deweyville Complex in the Grand Prairie area is similar to that of the Prairie Complex in that it includes multiple fluvial environments such as point bar, backswamp, and abandoned channel. Point bar deposits consist of the characteristic sedimentary sequence: a relatively thin surficial layer of mostly overbank deposits of well-oxidized silts and clays (ML and CL) that grades downward through cross-bedded silty sands (SM) and sands (SP) into thick, massive, graveliferous sands (SP and GP). The total sedimentary sequence averages about 100 ft thick and, considered as a whole, is relatively coarser grained than those of similar origin in either the Prairie Complex or Holocene meander belts of the same river. The lithology is no doubt a reflection of appreciably higher stream discharges and energy levels that were present when the sediments were deposited (see also Chapters 3 and 7).

Several large abandoned channels (neck cutoffs) occur on the Deweyville Complex just northwest of Stuttgart, Arkansas (Plate 7). Borings suggest they have an average depth of about 60 ft, but subsurface data are not adequate for a meaningful characterization of the deposits. By inference, it is reasonable to conclude that the upper 10 to 20 ft of the channel filling consists primarily of relative soft clays (CL) and silts (ML), and the deeper deposits consist primarily of sands and sands and gravels (SP and GP).

Without benefit of borings for verification, portions of the complex are interpreted (by process of elimination) as backswamp areas because of a lack of surficial ridges and swales indicating the presence of point bar accretion. If the interpretation is correct, the deposits in these areas should consist of 40 to 60 ft of moderately stiff, oxidized, mostly massive clays and silts (CL and ML).

Surface soils of the Deweyville Complex differ from those on adjacent units and from each other in various parts of the region (Autin et al. 1991). Except for the Arkansas River, most streams with a mapped Deweyville Complex drain old landscapes comprising mostly Ultisols and highly weathered Alfisols. Erosion of these soils yields sediments low in bases and weathered minerals that readily develop into mature soils. Selected characteristics of soils series along the Calcasieu and Ouachita rivers are shown in Table 3. Those along the latter stream have distinct depositional strata and are more acid and less red than those along the former stream.

Pleistocene Valley Trains

Interfluve and channel deposits

The lithology and sedimentology of the glacial deposits laid down by braided streams need to be considered in terms of both the characteristics of the thin surficial and nearsurface deposits (a fine-grained topstratum) and those of the much thicker and coarser grained underlying mass (the substratum).

With few exceptions, the valley train topstratum consists of a blanket of stiff, well-oxidized, tan or brown, very silty and sandy clays (CL and SC), silts (ML), and silty sands (SM) that is less than 10 ft thick and often not over 5 ft thick. The topstratum blanket typically is persistent across both gathering channels and interfluve areas (braid bars). In channels that are apparent from surface evidence, the fine-grained deposits are mostly slightly organic, horizontally bedded, slack-water accumulations of clays, silts, and sands. These sediments directly overlie massive, clean sands (SP) that characterized the channels when they were active. Based on radiocarbon dates and several discoveries of the remains of extinct Pleistocene megafauna that became entrapped in the sediments (Graham 1990, Morse and Morse 1983), channel filling by fine-grained sediments apparently began immediately after meltwater flow ceased.

The typical geotechnical properties (six parameters) of braid bar deposits of valley train interfluve areas are shown in Figure A2. Reflecting their considerable age (weathering history) and well-drained topographic position, the water contents (of both interfluve and channel areas) are comparable to those of

Pleistocene deposits (Figure A13). However, the cohesive strengths, while somewhat higher than in many Holocene deposits, are moderately low (Figure A19). In terms of the grain size distribution of the shallow noncohesive deposits of the topstratum, there is no statistical difference between those of interfluve areas and gathering channels (Figure A21). In both cases, the D10 grain size of the deposits is dominantly in the fine sand (0.1- to 0.3-mm) size range but also sometimes in the very fine sand and silt size ranges.

In interfluve areas, the sediments can be of various sources. On all valley train surfaces, some of the topstratum may be loess, and on the topographically lower levels of the St. Francis Basin, it may include materials deposited in Holocene Mississippi River backswamp and related overbank environments. In all areas, the lower portions of the topstratum may represent sediments laid down during the very last stages of outwash deposition when sediment loads and outwash volumes in the braided channels were markedly declining.

It should be realized that the uppermost several feet of sediments in the relict braided channels in many areas represents "postsettlement alluvium" originating from the erosion and runoff from nearby agricultural fields. In at least one instance, the material has been definitely established as postdating the 1811-1812 New Madrid earthquake series (Saucier 1989).

Horizontal limits of channel-fill deposits usually correlate with bank lines that are discernible by surficial geomorphic or topographic evidence when these features are present. However, at depths as shallow as only a few feet, surface evidence or expression may not be present. Based on correlations using lines of closely spaced borings (Whitworth 1988), it is evident that because of the instability of the braided channels, the channel-fill deposits are quite thin (generally less than 10 ft) and highly lenticular and interfingering, both laterally and vertically. It is not uncommon for channel-fill deposits to be encountered at a shallow depth beneath what was an interfluve during the final phase of outwash deposition that is evident by the surface morphology.

There is lithologic evidence from exposures in drainage channels on Macon Ridge and from shallow excavations in the Western Lowlands area that some of the channel-fill deposits actually represent a lacustrine environment. It is easy to envision how long, narrow, shallow lakes could have existed in recently abandoned braided channels on the valley trains. These apparently were the sites of the deposition of massive, gleyed silts that included the shells of freshwater mollusks, including gastropods and pelecypods (Saucier 1968).

Extensive investigations, including trenching, in the New Madrid Seismic Zone as part of studies of the liquefaction of shallow sands during earthquakes (see Chapter 8), have provided considerable information on the nature of the upper part of the valley train substratum. Immediately below the topstratum, the upper several tens of feet of the deposits consist of fairly well-sorted, fine- to medium-grained sands (SP). These are typically tan to brown in color and contain abundant, widely disseminated, small fragments of lignite and coal. At a local scale, the sands are clean and massive; however, at a regional scale, they include numerous lenticular masses of finer grained deposits. As shown in Figure 44, which is a conceptual facies diagram of a typical valley train sequence, the substratum includes scattered masses of channel-fill and lacustrine deposits that were laid down during an earlier phase of the aggradational cycle. These inclusions of finer grained sediments are continuous over horizontal distances measurable in hundreds to thousands of feet but rarely more than 20 ft thick.

In gross terms, the deeper part of the valley train substratum, often 75 to 100 ft or more thick, is a massive unit of sands (SP) and sands and gravels (GP and GW) that fines upward and toward the south. However, this generalization is misleading. Data from several deep borings and from electric logs of water wells indicate the deeper substratum actually consists of a series of fining-upward sequences from 5 to a maximum of 50 ft thick (Whitworth 1988). Some coarsening-upward sequences of similar dimension are also occasionally present, and in all cases the coarsest deposits (i.e., a lag concentrate) are not present at the immediate base of the valley train unit as might be expected. Each sequence probably represents a lobe-like pulse of outwash deposition that lasted anywhere from a few years to a few centuries. Few undisturbed samples of the sands and gravels are available and have been described in the literature. Based largely on braided stream analogs in other river systems, the deposits should exhibit extensive cross-bedding with planar, trough, and convolute stratification, thin silt layers, and small pockets of lag concentrate. Occasionally, graded beds (coarse to fine) are encountered but result from small migrating ripple trains (Coleman 1966a).

Theoretically, it should be possible to trace the lateral extent of the lobate sequences over distances of miles or tens of miles, but no sets of borings are spaced closely enough or logged or tested in sufficient detail to allow this tracing.

Sand dunes

Sand dunes have only recently been recognized in the Lower Mississippi Valley (Saucier 1978). The only available literature relates to their geomorphology, there is practically no discussion of the sedimentology, and there is nothing pertaining to the geotechnical characteristics of dune deposits. No detailed sample descriptions from borings are known to occur, and the only information on lithology comes from observations and samples by this writer from sand pits and dune blowouts.

All dunes appear to consist of homogeneous deposits of massive, well-sorted, fine sands (SP) or slightly silty sands (SM) that vary in color from tan and buff to grayish brown and reddish brown. There is a conspicuous absence of bedding or cross-stratification of the type that is typical of eolian dunes in coastal areas, but a lack of sedimentary structures appears to be characteristic of at least the shallower parts of dunes that occur in river valleys in the southeastern United States. This is thought to be due to the extensive weathering

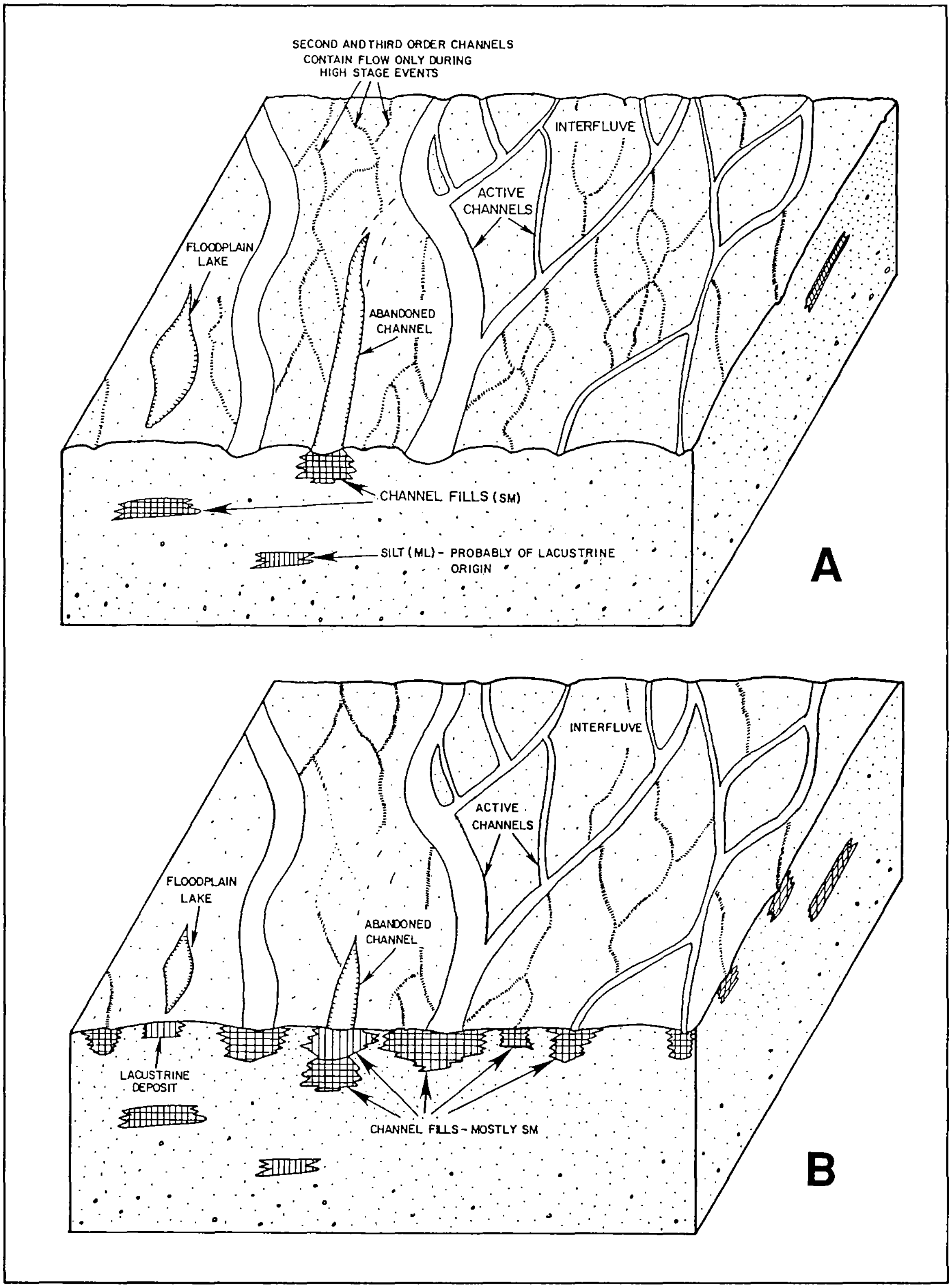


Figure 44. Model showing the distribution of subenvironments and deposits of a typical valley train during early (A) and late (B) phases of outwash deposition during waning glaciation (from Whitworth 1988)

and biogenic reworking that have affected the dunes. Most have been stable and subjected to soil forming processes between about 12,000 years ago and the late historic period.

Field observations suggest that most masses of dune sand unconformably overlie a thin, fine-grained, valley train topstratum a few feet thick. The presence of this layer of sediments of low permeability has served to allow a paludal environment to form and be preserved in a few of the deeper and more isolated depressions in the dune fields. These are characterized by swamp or bog vegetation or small, shallow, permanent ponds which have led to the accumulation of several feet of highly organic clays and silts (OH).

Soils of dunes are well oxidized, excessively drained and leached, and strongly acid, as would be expected considering their age, morphology, and lithology. These conditions have not facilitated archeological investigations which have been extensive in the dune areas because of the presence of large numbers of sites of considerable antiquity. For example, no pollen and only tiny fragments of bone have been preserved at the Sloan Site, a Dalton-period site on a sand dune in the Western Lowlands area of Greene County, Arkansas (Plate 5), that may be the earliest recognized cemetery in the New World (Morse and Morse 1983).

Loess

In contrast to sand dunes, there is voluminous information on virtually all aspects of loess--to the extent that it is hard to summarize. The most complete description of the textural, mineralogical, physical, and chemical aspects of loess in a single source is provided by Snowden and Priddy (1968) while Krinitzsky and Turnbull (1967) provides the most definitive information on parameters of engineering significance. Other descriptions mainly focus on pedogenic aspects of loess as they relate to the differentiation of the multiple loess sheets (layers) and their age determination.

In an unweathered state, loess generally consists of tan, brown, or reddish brown, fossiliferous, massive, uniform, calcareous silts (ML) and silt loams (CL). Weathered loess within several feet of the surface generally is slightly higher in clay content (CL), darker in color, and leached, and therefore noncalcareous. Thin leached zones at depth within a loess sequence denote the presence of paleosols which arise from the slowing or cessation of loess deposition, followed by a period of weathering, and then resumption of deposition. Unweathered loess has the ability to maintain vertical slopes if protected from surface runoff: this characteristic is attributable to its high vertical permeability (which reduces or eliminates water saturation), binding of silt- and larger-sized particles by thin clay and carbonate films, and an internal "skeleton" of hollow, vertically-oriented, calcareous root tubules. Loess also contains relatively large numbers of dispersed calcareous nodules and irregular, spheroidal, or elongated concretions in branching masses up to 6 in. or more in length. The latter are believed to form by the precipitation of the carbonates in groundwater in channels left by the decay of tree roots.

Fossils that are preserved in unleached loess consist primarily of the well-preserved, typically unbroken, delicate shells of pulmonate gastropods (snails). Smaller quantities of freshwater snails and mollusks have been encountered in loess, indicating the former presence of ponds or small lakes (Krinitzsky and Turnbull 1967). Occasional discoveries of scattered bones to near-whole skeletons of extinct Pleistocene vertebrate fauna, especially mastodon (*Mammut americanum*) and mammoth (*Mammuthus jeffersonii*), have been encountered buried in loess in all parts of the Lower Mississippi Valley from Crowley's Ridge and western Tennessee to southern Louisiana.

More than 50 different surface soil series have been distinguished on loess in Louisiana: selected characteristics of four soils are shown in Table 3. Alfisols, Mollisols, and Inceptisols developed in the relatively young Peoria loess (see Chapter 5), and highly weathered Alfisols and Ultisols formed in the Sicily Island loess (Autin et al. 1991). Basal mixed zones beneath the loesses generally thicken with distance from the source and intersect the present land surface where the loess is less than about 4 ft thick. Soils developed in thin loess contain admixtures of the underlying material throughout their sola, probably because of bioturbation.

Pleistocene Substratum

It is reasonably safe to assume that the coarse-grained deposits that immediately underlie the valley train surfaces (i.e., to a depth of a few tens of feet) represent the same glacial cycle as the surficial deposits *per se*. However, if concepts of multiple, large-scale cutting and filling cycles in the alluvial valley area are correct, a significant amount of the deeper substratum deposits at a given location may not be temporally related (of the same cycle) as the surficial deposits. It is conceivable that masses of outwash from one or more earlier Pleistocene outwash cycles underlie all or some of the outwash of Early or Late Wisconsin age that constitutes the valley train surfaces: the latter may be only a few tens of feet thick or, on the other hand, could extend completely to the base of the Quaternary alluvial sequence.

No undisturbed samples of the graveliferous substratum are known to exist, and the deposits have never been directly observed (or at least described) because of their depth. Consequently, present information and technology do not allow chronostratigraphic differentiation. No geophysical or other method is capable of distinguishing the complexly interfingered, fine- to coarse-grained, graveliferous sands (SP, GP, and GW) of one unit from those of another. Furthermore, on the basis of existing data, it is not possible to lithologically delineate the base of channel scouring that has taken place within the substratum beneath the Holocene meander belts. Sands and gravels of the deeper point bar sequences are almost never distinguishable from deeper deposits of other origin. Although areas of the point bar environment are explicitly delineated on the many cross sections that accompany the quadrangle-scale maps of the environments of deposition (e.g., Kolb et al. 1968, Saucier 1967), the vertical limits are only estimated from known depths of flood-stage scouring in thalwegs on the present river.

This writer is aware of occasional occurrences in borings of iron-cemented sands and gravels and organic-rich clays, both of which could be indicative of paleosols that formed on outwash of pre-Wisconsin age. However, neither the age nor the areal extent of such materials has been determined. Numerous radiocarbon assays have been conducted on pieces of wood from the deeper substratum in various parts of the alluvial valley area; however, invariably the materials have turned out to be "dead," i.e., beyond the dating limits of the technique (about 35,000 years). This is not unexpected considering that such materials no doubt have been eroded out and redeposited many times. Similarly, concentrations of bones of several species of extinct Pleistocene megafauna have been encountered at low river stages on point bars of the modern Mississippi River. These faunal remains have been scoured from the deeper substratum (glacial outwash?), but in no case has their precise provenance been established. Dates obtained on such materials would only provide minimum dates for the deposits and would help little in stratigraphic determinations.

Consequently, the substratum remains chronologically and stratigraphically undifferentiated and probably represents mostly glacial outwash that has been reworked repeatedly by fluvial action. As a result of the reworking and the deflation of large volumes of silt, the deeper substratum deposits have been washed and winnowed and therefore are relatively coarser on the average than the shallower outwash deposits.

The coarse-grained substratum of the alluvial valley has received extensive consideration by the U.S. Geological Survey and state agencies because it constitutes the immense "alluvial aquifer" that is many cubic miles in extent (e.g., Broom and Lyford 1982, Turcan and Meyer 1962, Whitfield 1975). Most consideration, however, has focused on the gross geometry of the aquifer and its hydraulic and water quality characteristics (e.g., conductivity, transmissivity, and dissolved solids). These studies have noted only that in terms of lithology, the deposits tend to get coarser with depth with the largest gravels (and sometimes cobbles and boulders) encountered in wells near the base of the sequence.

Holocene Fluvial Environments

Natural levees

Lithologically, natural levees are the locally, regionally, and temporally most variable of all fluvial environments. Perhaps because of this fact, they may be the least described especially in the alluvial valley area.

Depositional rates, as they are causally related to the nature of overbank flooding on the adjacent river channel, are the dominant factors in determining natural levee lithology at a local scale. Regionally, they reflect the progressive downstream decrease in the average grain size of the suspended load of the river. In all parts of the Lower Mississippi Valley area, age of the natural levee strongly influences the degree of weathering and secondary biogenic reworking of the sediments.

Between Cairo and the head of the Atchafalaya River (the lower limit of the alluvial valley), a "typical" Mississippi River natural levee consists of medium to stiff, mottled gray, tan, and brown, silty clay (CL), sandy clay (CL), or silty sand (SM). The sediments are highly oxidized with abundant iron and manganese nodules and are moderately to highly affected by bioturbation (Table 4). In the deltaic plain area, natural levees of the Mississippi River and its abandoned distributaries are finer grained on the average, and there is a discernible downstream decrease in the silt and sand content (Figure 45). In the modern delta, natural levees consist entirely of plastic clays (CH).

However, departures from these typical situations are frequent. For example, in the alluvial valley area, natural levees that formed on the outer (concave) banks of rapidly migrating bends where crevassing was frequent contain large amounts of silts (ML), silty sands (SM), and fine sands (SP) as thick lenses or layers. On the opposite sides of the river at the same locations, or laterally within distances of only a few thousand feet where velocities of overbank flows may have been considerably less, the levees may be composed primarily of clays (CH and CL) and silts (ML).

It is to be expected that natural levee deposits would be texturally laminated since they are formed by the incremental deposition of sediments from overbank floods, and since the severity of individual floods varies considerably from event to event. This is definitely the case with young deposits (less than a few hundred years old) which have been observed to consist of thin, parallel layers of clays, silts, and fine sands. However, in older natural levees, biogenic reworking by plant roots and burrowing organisms frequently has removed most traces of bedding, and the clay and silt layers have been reworked (homogenized) into a silty clay with irregular silt pockets.

In the alluvial valley area, most natural levees overlie point bar deposits and to a lesser extent backswamp deposits. In both cases, the natural levees are gradational with the underlying deposits rather than separated by an unconformity. Because of the lithologic variability of the levee deposits, it is often difficult in cross sections to indicate a line separating the two based on soil types alone, although this is commonly done. Sedimentary structures such as oxidation mottling, ferruginous nodules, and burrows often are a more reliable indicator of the presence of conditions that typify the natural levee environment.

Within each of the six Holocene Mississippi River meander belts, the deposits are sandier on the highest parts of natural levees and more clayey in

TABLE 4.
OCCURRENCE OF MINOR SEDIMENTARY STRUCTURES IN HOLOCENE FLUVIAL DEPOSITIONAL ENVIRONMENTS

	Depositional Environments					
Sedimentary Structures	**Natural levees**	**Point bar accretion**	**Backswamp**	**Abandoned channels***	**Abandoned courses**	**Crevasse splays**
Bedding						
thick >7 in.					D	
medium 2-7 in.	C			C	C	C
thin <2 in.	C-A		C	A		A
Morphology						
parallel laminations (texture)	C	A	C	C-D	A	C
parallel laminations (color)			A	C		
lenticular laminations			R			
wavy laminations		C	R	C		
Cross-laminations						
simple					C	
planar		C			A	
trough		A				
Ripple laminations						
current		D-A			C	A
ripple-drift		A			A	
wave		A				
Scour and fill		D			C	
Plant remains						
distinct particles	C	C	A	R	R	
finely divided		A	C	A	C	R
bedded			C	R		
Shell fragments			R			
Clay inclusions		C			A	
Load casts					C	
Distorted laminations	R		C	C		
Burrows	A		D	R		D
Nodules						
ferruginous	A		C			C
calcareous	C		A	R		
Desiccation Features	C		C	R		C
Oxidation mottling	D-A		C-R			A

D-dominant; A-abundant; C-common; R-rare. *Upper, fine-grained "clay plug" portion only.

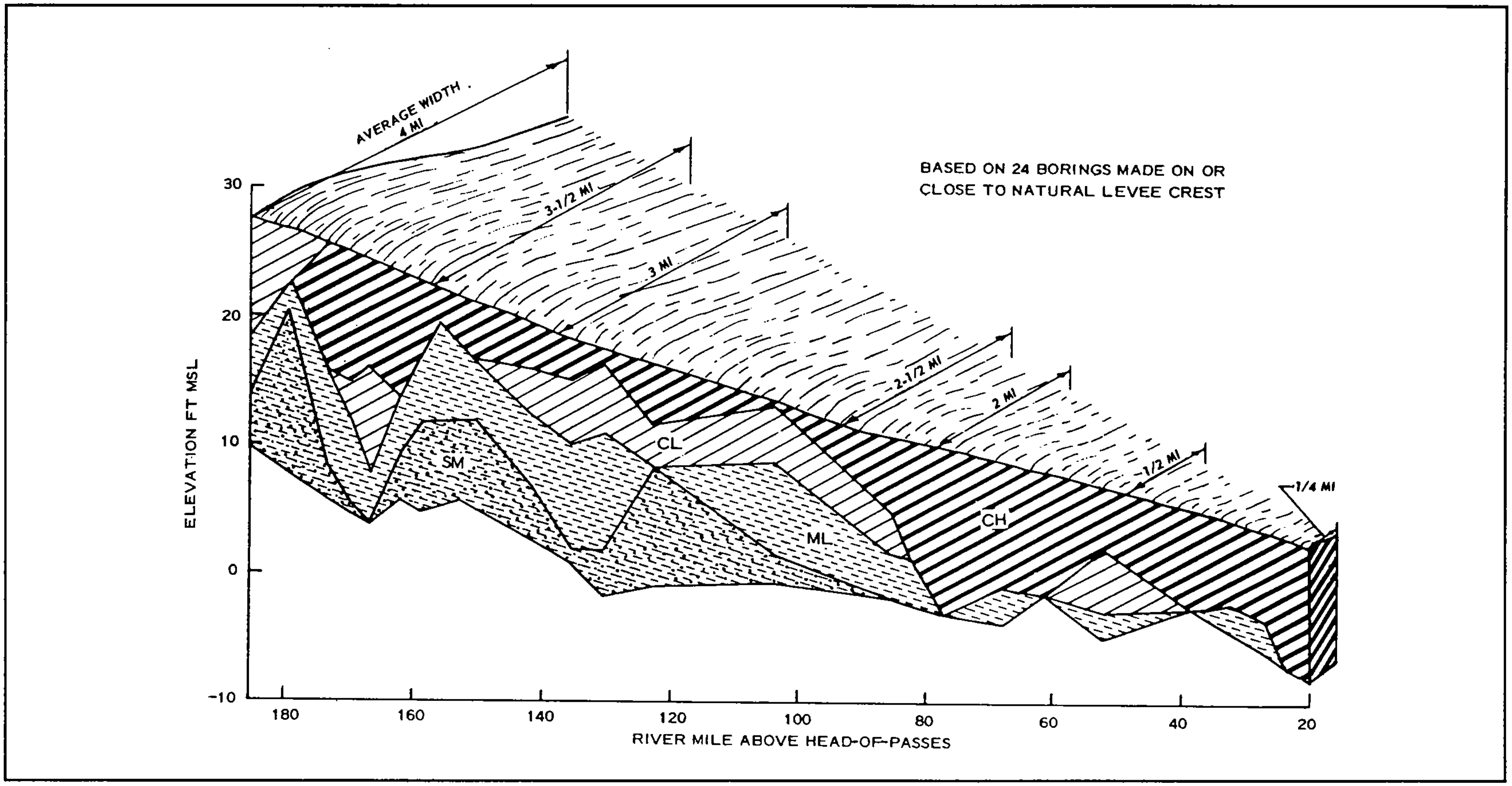


Figure 45. Longitudinal slope, width, thickness, and general composition (soil-type frequency diagram) of natural levee deposits from near Donaldsonville (Plate 11) to near Fort Jackson, LA (Plate 14) (from Kolb 1962)

backswamp locations (Autin et al. 1991). Comparisons among soils within a given meander belt illustrate the influence of clay content as well as depth and duration of water table on soil development. Soil characteristics from the highest part of a natural levee to the lowest part of a backswamp are summarized in Table 3 for the youngest of the meander belts.

Geotechnical properties of natural levee deposits are summarized in Figure A3. Unfortunately, the vast majority of the tested samples are from the lower portion of the alluvial valley and the upper part of the deltaic plain. Consequently, the properties of typical natural levee deposits for the entire alluvial valley area are not reflected.

Since natural levees are subject to desiccation and oxidation, for a given engineering soil type (i.e., CH or CL), the water contents are generally lower and the cohesive strengths are generally higher than in other deltaic environments where these processes are limited or absent. However, the differences among environments is not readily evident in Figures A13 and A19. This is due to the wide variation in strength and water content that exists both regionally and from one location on a natural levee to another, conditions that are both represented in the data set. In a situation where natural levee deposits overlie point bar deposits, for example, cohesive strengths of the former will be appreciably greater than those of the latter near levee crests along older distributaries but appreciably less in distal natural levee settings in the modern delta.

Point bar accretion

Beneath that portion of a point bar sequence that can be most appropriately regarded as natural levee, the topstratum of a point bar ridge consists of a few feet of gray or tan, oxidized, silty or sandy clay (CL) or silty sand (SM). Below the topstratum is a thick, coarsening downward sandy substratum that constitutes the "typical" point bar deposits. Although the deposits as a whole are coarser grained in the upper part of the alluvial valley area and become progressively finer grained downstream, all areas exhibit a typical vertical sequence that grades downward from well-sorted, fine and medium sands (SP) to medium and coarse sands with gravel (SP or GP) (Figure 46). On smaller rivers like the Arkansas and the White, the same coarsening downward sequence prevails, but silts (ML) and fine sands (SP) are more prevalent and there is significantly less gravel.

Most information regarding the nature of point bar deposits comes from borings (Sacre 1987) and therefore contains relatively little data on sedimentary structures. Nevertheless, there have been several opportunities to observe the deposits in large excavations. From these, it is apparent that the sands in all point bar sequences are highly cross-bedded with fine current ripples and laminations in the upper part and coarser textured trough stratification in the lower part. By far the best description and illustration of point bar sedimentary structures is that of Frazier and Osanik (1961) who documented the

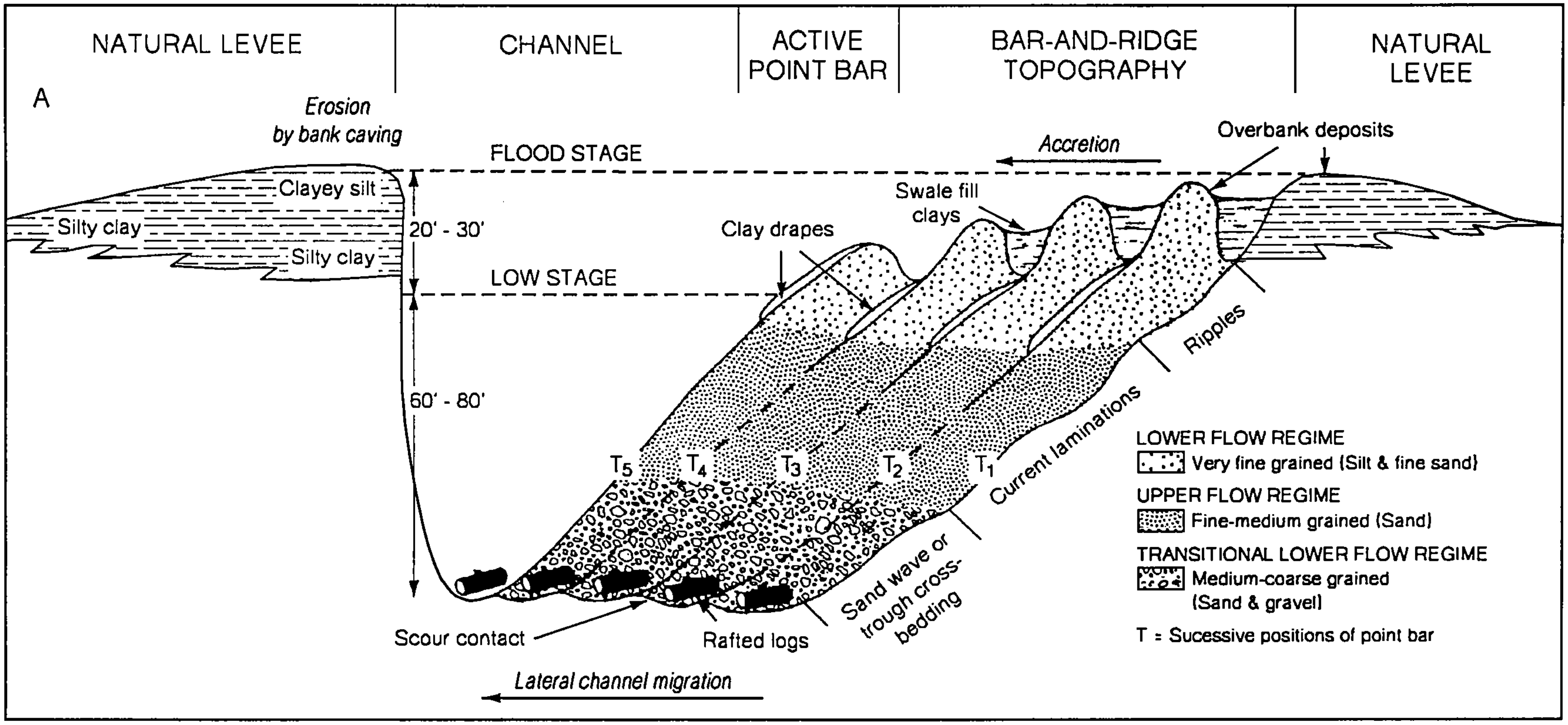


Figure 46. Diagrammatic cross section of a meander belt showing typical distribution of deposits and sedimentary structures in a point bar sequence (from Gagliano and van Beek 1970)

extraordinary 40-ft section exposed in the excavation for the Old River locksite in Louisiana (Plate 10). They provide photographs and interpretive drawings of fine- and medium-grained sands (SP) with magnificently developed festoon cross-bedding of scour and fill origin. These also reveal that the sands contain thin layers of pebble-sized gravels, frequent thin lenses of macerated plant material, layers of heavy mineral concentrations, and wood fragments of all sizes. Other materials observed included armored mudballs, reworked fresh-water shells, and even bone fragments.

In reference to Figure 46, the visual prominence and relative size of the rafted logs that were shown by the authors (Gagliano and van Beek 1970) at the base of the point bar sequence seems to overemphasize their relative importance. However, this writer has consistently been surprised by the amount of stream-transported organic matter (including logs) that is present in Mississippi River point bars. High organic contents are normally associated with the backswamp and abandoned channel environments and not in granular deposits representing higher energy levels; however, on a volume basis, many point bars possibly have an actually higher amount of total organics.

In point bar sequences, fine-grained, cohesive deposits occur mainly in the topstratum and the upper part of the substratum as either very thin clay drapes (generally less than an inch thick) or as swale filling. Small swales may contain only a few feet of silty or sandy clay (CL), clayey sand (SC), or silty sand (SM) unconformably overlying clean sands (SP), whereas the larger, deeper swales (a hundred or more feet wide and perhaps thousands of feet long) may contain several tens of feet of soft, gray fat clays (CH), organic clays (OH), or clayey silts (CL-ML).

The geotechnical characteristics of point bar deposits are summarized in Figure A4. It should be noted that most of the tested samples were obtained from fine-grained topstratum deposits in the upper part of the deltaic plain. For most parameters, there is essentially no difference between the point bar deposits and natural levee deposits (see Figure A3), reflecting the fact that the two environments are indeed gradational and the distinction often being a matter of definition rather than one based on physical properties.

The D10 grain size parameter (Figure A21) represents samples derived from the upper several tens of feet of the point bar substratum. That value indicates that the deposits are generally finer grained than valley train deposits (braided stream interfluves).

From an engineering viewpoint, the vast majority of coarse-grained point bar deposits are dense to very dense and therefore provide competent foundation conditions. However, as notably pointed out by Carey (1969), anomalously loose deposits of considerable extent can occur in point bar areas under certain unusual conditions. These conditions exist at four locations between Natchez and Baton Rouge (Butler Lake, Fort Adams, Port Hickey, and Devils Swamp areas shown in Plates 10 and 11) where the normal migration of a Mississippi River bend has been inhibited by impingement against the valley wall. When this occurs, the bend migrates downstream rather than laterally, cutting into its point bar on the downstream convex bank and depositing materials in a pressure eddy zone on the upstream concave bank. The resulting "eddy accretion" consists of extremely loose clayey silts (CL or ML) and fine sands (SP) that are striking different from point bar deposits either upstream or downstream. At the surface, a shallow lake or swampy depression occurs rather than typical point bar accretion topography.

Backswamp

Detailed information on backswamp sediments is available only from studies of the stratigraphy (Krinitzsky and Smith 1969) and sedimentology (Coleman 1966a) of deposits in the Atchafalaya Basin. In both efforts, the intent was to use X-radiography and traditional sedimentologic analyses of large-diameter continuous cores to establish paleoecological changes in the thick (greater than 100 ft) sedimentary sequence. Although the interpretations differ considerably, three subenvironments of deposition were recognized by both and were designated well-drained swamp, poorly-drained (deep) swamp, and lacustrine (including lacustrine delta). In the basin area, the three subenvironments exhibit a complexly interfingering relationship both horizontally and vertically.

Lacustrine deposits comprise a majority of the Atchafalaya Basin sequence (Krinitzsky and Smith 1969) and are discussed later in this part of the report. Poorly-drained swamp deposits, although well represented in the sequence, are not thought to extend much farther north than about the latitude of Natchez. Well-drained swamp deposits comprise a relatively small part of the Atchafalaya Basin sequence but essentially all of the backswamp sequences to the north of Natchez.

Swamp deposits of both types consist of firm to stiff, mostly gray to black clays (CH) and silty clays (CL) with thin silt laminations and frequent burrows (Table 4). Organic matter is abundant both as woody fragments and scattered small particles. Bedded organics in the form of peat layers or zones of compacted leaf litter are infrequent even in poorly-drained swamp deposits. Well-drained swamp deposits typically exhibit color mottling (grays, browns, buffs), abundant ferruginous and calcareous nodules and staining, and slickensides resulting from shrink and swell associated with periodic wetting and drying.

As discussed in Chapter 5, areas between Arkansas River meander belts in the Arkansas Lowland (Plates 7 and 8) qualify as backswamp under the definition of this environment of deposition. However, the deposits are considerably coarser grained then elsewhere in the alluvial valley area with silty clays (CL) predominating over fat clays (CH). Moreover, because of the proximity and numbers of meander belts present, backswamp areas within this lowland have experienced the frequent deposition of layers of silts (ML) and fine silty sands (SM). Perhaps because of somewhat higher sedimentation rates, the organic content of the deposits is also noticeably less.

When the Atchafalaya Basin is considered as a whole, there are two aspects of the backswamp sequence that are worthy of note. First, as determined from lines of closely spaced borings (Krinitzsky and Smith 1969), layers of silt (ML), silty sand (SM), and fine sand (SP) up to several feet thick are scattered throughout the area at various depths but are not regionally continuous. They occur in all subenvironments and seldom can be correlated from one boring to another over distances of hundreds of feet. These layers are probably attributable to major crevasses or minor distributaries on the Mississippi River meander belts that flank the basin. It is apparent that rather than forming a thin, widespread silt or sand sheet, the flow from these events must have been largely restricted to existing backswamp drainage features and the sediments were laid down as ribbon sands. Second, the backswamp deposits directly and abruptly overlie glacial outwash deposits with little or no apparent transition from one environment and sediment type to the other. At a depth of 90 to 120 ft, backswamp clays (CH) frequently are in direct contact with the underlying, massive medium- to coarse-grained outwash deposits (SP). This abrupt change in depositional environment is believed to be a direct result of a sharp reduction in sediment supply and the Holocene sea level transgression which created a large drowned area or embayment in the lower part of the alluvial valley.

Selected properties of soils developed in backswamp areas of Louisiana are presented in Table 3. Elsewhere in the alluvial valley, soils consist primarily of the Sharkey and Alligator series where natural levee deposits from adjacent meander belts are thin or absent.

The tabulation and analysis of geotechnical properties presented in Figure A10 represents samples from both the backswamp and inland swamp environments and unfortunately does not differentiate between well-drained and poorly-drained conditions. Hence, there is a considerable range in the values for all parameters. A parameter by parameter comparison of backswamp properties to those of other fluvial and deltaic environments is presented in Figures A13 to A21. The relatively high typical water contents reflect to an appreciable extent the presence of varying amounts of organic matter. The relatively low cohesive strengths reflect the general absence of desiccation and oxidation except to a minor degree in well-drained swamp areas.

Abandoned channels

In a work of the scope of this synthesis, invariably there are surprises--both to the writer and the readers. One of these is the conspicuous lack of descriptive information on the nature of abandoned channel deposits. This is indeed surprising in view of the considerable attention that has been devoted by geotechnical engineers to "clay plugs" and the comprehensive investigation by Fisk (1947) in which hundreds of borings in such deposits were compiled and analysed. These subsurface data are adequate to define the general lithology of abandoned channel deposits, and there are large volumes of data from which to tabulate geotechnical characteristics (Figure A5), but sedimentologic data in published form are not available. Even reports on paleoecological interpretations, including radiocarbon dates, do not contain useful information of that type (Thorne and Curry 1983). One major reason for this information void is the absence of large excavations in this environment which, because of its weak soils, has been avoided in many construction activities. Another is the near-absence of abandoned channels in the deltaic plain area where most systematic sedimentologic studies have been conducted.

Two major lithologic units are present in most abandoned channels created by neck cutoffs and must be considered separately. The sand wedge or plug portion of the channel filling that forms mainly in the arms of a cutoff during early stages (Figure 20) consists predominantly of cross-bedded, fine to medium sands (SP) and silty sands (SM). They closely resemble and often cannot be differentiated from point bar deposits at a comparable depth; however, near the transition into the overlying clay plug, they are often interlayered with silts (ML) and silty clays (CL). This situation reflects the stage in channel filling in which slack-water sedimentation alternates with flood-stage sedimentation.

The overlying fine-grained or clay plug sediments are what most people regard as abandoned channel deposits. These consist predominantly of very soft to medium, gray, slightly organic, silty clays (CL) and clays (CH). Since the sediments are typically laid down in perennial water bodies or deep swamp environments and rarely exposed to oxidation or desiccation, they lack color mottling and nodules except in the uppermost portions of channels that are essentially filled (Table 4). Contrary to the impression created by the engineering concerns over the compressibility and high water contents of the deposits, infrequently are they truly massive clays. Most sequences of abandoned channel deposits exhibit distinct layering with individual beds generally not more than a few inches thick. Not unexpectedly, they exhibit many of the same characteristics as lacustrine deposits. Clay (CH and CL) sequences are frequently interrupted by thinner, planar layers of silts (ML) and fine sands (SM or SP) that mark major flood events. In terms of total volume, it is not unusual for 20 to 30 percent of the clay plug portion of an abandoned channel to consist of noncohesive sediments. The percentage is much higher in those channels that were filled rapidly and those in the upper part of the alluvial valley.

Deposits filling abandoned channels along smaller streams such as the Arkansas, Red, and White rivers are analogous to those of the Mississippi River, only proportionately finer grained. Silt and sand layers are less numerous, and organic contents tend to be significantly higher.

Clay plugs are considerably less well developed in abandoned channels originating from chute cutoffs than they are in those resulting from neck cutoffs. Because chute cutoffs result in channel segments that are much shorter and depart from the parent channel at a shallower angle, there is considerably less opportunity for lacustrine or slack-water conditions to develop that would be conducive to the deposition of clays and silts. Most chute channel filling consists of cross-bedded silty sands (SM) and fine to medium sands (SP) with gravel lenses and layers at depth that resemble the sand wedges in neck cutoff channels. Those clay plugs that do occur contain relatively large amounts of silt and sand in the form of laminated or layered clays and silts (CL and SM).

Abandoned courses

There is even less information available on the sedimentary characteristics of abandoned courses than on abandoned channels, but this is not surprising considering the limited geographic extent of the former. Relatively few borings have been drilled into this environment, and there are no published descriptions of the sedimentary sequences. Most of the information presented below comes from observations by this writer of the large excavation that was made for construction of the Jonesville Lock & Dam along the Black River in east-central Louisiana (Plate 10). This excavation was made into the abandoned course of the Walnut Bayou meander belt of the Mississippi River (M2C, Plate 1) (Saucier 1964).

All abandoned course sequences appear to have a thin, fine-grained topstratum overlying a much thicker, coarse-grained substratum. The topstratum may be of various origins. Where it represents slack-water sedimentation after complete abandonment of the course by the river, it probably consists of very soft to soft, gray, organic clays (CH and CL) and silts (ML). Where it represents point bar accretion by a small stream flowing within the confines of the larger channel, the deposits will be coarser grained with medium, gray, silty, or sandy loams (ML) and silty sands (SM) being the dominant soil types.

The substratum portion of abandoned course sequences represents channel-fill sediments deposited during the stage of waning discharge when an upstream diversion was taking place. These sediments consist of gray, fine to medium, well-sorted sands (SP) that exhibit large-scale, tabular cross-stratification with ripple-drift cross-laminations formed by migrating sand waves. Individual sets (at the Jonesville site) were 6 to 15 in. thick and separated by 0.5- to 1.0-in.-thick horizontal layers of clay (CH and CL). The sand layers probably represent deposition during a single, major flood event or a series of closely spaced events when a pulse of sand moved down the course: the clay layers must represent slack-water sedimentation during periods of little or no flow. The presence in the deposits of abundant armored clay balls (ripup clasts) indicates there were frequent periods when clay layers were scoured during a succeeding flood event. Toward the upper part of the substratum, the sand layers became thinner and the clay layers more numerous, but not noticeably thicker.

As is the case with sand wedges in abandoned channels, there are few or no lithologic or sedimentologic differences between abandoned course substratum deposits and laterally adjacent point bar deposits. Hence, mapping of abandoned courses must be accomplished using surface or nearsurface evidence.

Geotechnical characteristics of both the topstratum and substratum portions of abandoned courses are presented in Figure A6. Although there are important differences in sedimentary characteristics between abandoned channel and abandoned course deposits, it can be observed that the geotechnical properties are essentially the same. With regard to water contents and cohesive strengths, the values are intermediate between those of the natural levee and backswamp environments.

Crevasse splays and channels

Crevasse splays occur in most parts of the alluvial valley and upper deltaic plain, but "text book" examples occur in the False River area of Louisiana (Plate 11) and have been the subject of a detailed sedimentologic study by Farrell (1987).

The model developed by Farrell includes a lower sandy silt (SM) unit, a rooted clay unit (CH or CL), and an upper silty sand unit (SM) in a sequence that overlies backswamp deposits. The silty and sandy units are composed of multiple rhythmites which are defined as alternating beds or laminations of contrasting lithologies deposited during a single sheet flood event. Each rhythmite consists of well-sorted, very fine, ripple-laminated silts or sands which fine upward into clay, mud, or muddy sand. This sequence reflects conditions that existed during initial crevassing, full splay development, and and finally abandonment of the crevasse. With time, however, the primary stratification may become partially to completely obliterated because of biogenic reworking (worm burrows and root traces) or pedogenic processes.

The cyclic sedimentary sequence described above and the processes responsible are essentially the same as those that characterizes a natural levee. This is logical because most natural levees are largely the product of numerous episodes of crevassing and an aggregate of multiple overlapping and interfingering splays and sheetflow deposits. Consequently, in a maturely developed meander belt, the sedimentary characteristics of a splay, such as are indicated in Table 4, are discernible only for features of unusual size or those that developed toward the end of the period of maximum flow and are apparent at the surface.

Very little is known about the sedimentary content of the discrete, small, shallow channels on a splay that typically radiate from the point of crevassing. Because of their ephemeral nature, it is likely that they fill with silts and sands and, after a period of weathering and biogenic reworking, probably become indistinguishable from the sediments that are laid down by sheet flow. However, the larger and longer-lived channels from a major crevasse that develop into a distributary system (e.g., see Figure 17) are features with distinctive lithology. These are completely analogous to abandoned courses of full-flow streams and contain cross-bedded, fine sands (SP) with thin clay (CH or CL) layers indicative of episodic discharge.

Alluvial fans

Because of appreciable variations in the lithology of the upland formations and a considerable range in discharges and gradients (i.e., energy levels) in the streams that created alluvial fans, it is unwise to generalize regionwide as to their sedimentary characteristics. Based on the detailed work of Smith (1983), it is possible to generalize for those fans of roughly similar size and age that occur at the base of the bluffs in the Yazoo Basin area (Plates 8 and 9).

Smith (1983) recognized four types of alluvial fan deposits (i.e., channel, intersection point lobe, sheetflood, and splay) encompassing 10 separate sedimentary facies. These vary in lithology from lateral gravel bars that are pebble-size gravels in a sand matrix (GP or GW) through lateral sand bars that are well-sorted, cross-bedded, medium sands (SP) to fine-grained sheetflood deposits that are poorly sorted, massive sandy silts with scattered gravel (SM). Each facies constitutes a lobe whose thickness may vary in terms of feet (channel deposits) or inches (sheetflood deposits). Because the sedimentary properties of the facies frequently overlap, they are often hard to differentiate in borings or small outcrops. The primary differences in the facies relate more to their external geometry, thickness, and location on the fan than to their internal characteristics. For example, fine-grained sheetflood deposits may be distinguished from deposits of channeled sheetfloods by their location on the fan (the former on the proximal fan, the latter on the distal fan).

Contrary to popular belief, sediments comprising alluvial fans do not necessarily become finer grained with increasing distance from the fan apex because of the lobate pattern of growth and the multiple depositional processes involved. It has been observed, however, that proximal fan strata are the most uniform while those of the distal fan are the most diverse due to the downfan decrease in the thickness of individual strata and the increase in the number of processes involved. In a typical proximal fan setting, a majority of the total sedimentary sequence is composed of fine-grained sheetflood deposits.

The facies model developed by Smith (1983) for the Yazoo Basin fans does not apply to fans of medium to large size such as those of the Castor and Current rivers (Figure 9). Although the depositional processes are similar, the size differences in the fans are so great that the latter can best be viewed as delta-like systems wherein specific environments like distributaries, interdistributary lowlands, and even abandoned channels can be recognized. Environment for environment, the sedimentary characteristics are comparable to those of the Mississippi River delta, taking into consideration the overall finer grained sediment loads of the smaller fluvial systems.

Tributary valley fill

Valley-fill thicknesses typically are proportional to the size of the stream valleys and their drainage basin area. As with alluvial fans, the lithology of the upland formations through which small tributaries flow largely determines the character of the valley fill. In simplistic terms, most valley fills exhibit a fining-upward sedimentary sequence with a fine-grained topstratum and a coarse-grained substratum. The substratum, consisting of channel deposits, is usually distinct from the topstratum which is a veneer of overbank facies, but important exceptions occur.

One general type of valley fill, well illustrated by the conditions along several streams in north-central Mississippi, has been extensively studied by personnel of the U.S. Department of Agriculture Sedimentation Laboratory at Oxford, Mississippi (Grissinger, Murphey, and Little 1982). In a typical sequence, they recognize a basal unit of cross-bedded sands and gravels (SP and GP) with abundant woody debris laid down during the terminal Pleistocene or very early Holocene as channel lag deposits (lateral accretion). Bog-type sediments containing leaves, twigs, nuts, and stumps and logs are also present in the basal unit. The next higher unit, more than 10 ft thick, is a massive deposit of silt (CL and ML) that grades downward into silty sand or sandy silt (SM). This unit, dating to the early Holocene, may represent a low-energy fluvial deposit (vertical accretion) derived from the erosion of loess in the adjacent uplands. Above a well-developed paleosol is the upper unit which consists of late-Holocene meander belt alluvium, a heterogeneous mixture of sediments laid down in point bar, abandoned channel, and related environments.

A second general type of valley fill is especially well developed in west-central Mississippi and western Tennessee consists of several feet to 10 ft or more of mostly fine-grained postsettlement alluvium. These sedimentary sequences include interfingering, lenticular masses of mostly poorly sorted silts (ML), silty sands (SM), and sands (SP), often incorporating a variety of cultural materials of historic and late prehistoric age. The postsettlement alluvium overlies a thin, late Holocene, fine-grained topstratum which in turn overlies a coarse-grained substratum. Areas with appreciable thicknesses of alluvium often are discernible by the presence of an anastomosing network in the stream valley. This stream pattern closely resembles that of a braided stream, but the anabranches of the network are highly stable in position rather than frequently shifting as in a braided system.

A third general type occurs in fluvial systems that drain the Upland Complex (e.g., in southwestern Mississippi and the western Florida Parishes) where abundant coarse-grained sediments are available. Where floodplain destabilization has taken place because of historic-period land use practices, a fine-grained topstratum often is completely absent and the valley fill consists almost entirely of sands and gravels (SP and GP). Broad, unvegetated, sand and gravel flats and bars are widespread, and finer grained sediments such as silty and sandy clays (CL) occur only as thin layers in shallow chutes or swales.

Holocene Deltaic Environments

Distributary natural levees

Figure 45 shows the downstream trend in soil types in the natural levees of the present Mississippi River meander belt in the lower deltaic plain: the same general trend occurs in the natural levees that flank abandoned distributaries such as Bayou Lafourche and Bayou la Loutre (Plate 14). Along most distributaries, discrete crevasse splays are absent, and in the few cases where floodwaters became channelized, a subdistributary formed rather than a sheet-like splay. Consequently, natural levees are largely free of significant lenticular masses of silts or sands.

A typical distributary natural levee consists of firm to stiff, massive to weakly laminated, mottled gray and brown clays (CH) and silty clays (CL) with moderate amounts of wood fragments (Table 5). The more upstream (and consequently slightly older) and the higher natural levees exhibit intensive burrowing, oxidation, and iron nodules and concentrations. The more downstream (and deeper) levees are less oxidized and not as biogenically reworked; consequently, thin silt beds or lamina representing individual flood events are more conspicuous (Coleman 1966a). Because they are less weathered and desiccated, the distal distributary natural levees have noticeably lower consistencies.

Where distributary natural levees developed downstream over interdistributary marsh, the largely inorganic levee sediments unconformably overlie peat (Pt) or highly organic clays (OH). Where levees have progressively widened such as during the constructional phase of a delta lobe, thin layers of interdistributary marsh or inland swamp deposits often occur beneath and interfingered with the distal levee sediments. Geologists welcome these occurrences since they provide a means of obtaining meaningful radiocarbon dates that immediately precede distributary development and natural levee formation.

Where natural levees developed into shallow bays or Gulf waters as distributaries prograded seaward, the subtidal to near sea level portions (subaqueous natural levees) are well stratified with parallel and wavy laminations, ripple bedding, and occasional scour and fill structures. The sediments have lower consistencies than those in the higher, subaerial portions and generally lack indications of oxidation and desiccation such as nodules and color mottling (Coleman and Gagliano 1965).

In the case of the natural levees along the relatively older distributaries which have subsided beneath and/or have been encroached upon by interdistributary marsh, the sediments maintain their diagnostic oxidation and biogenic-reworking characteristics for long periods of time (at least millennia). Thus, buried levee sediments can normally be recognized when they are encountered in borings by way of their color, consistency, and nodule content.

TABLE 5.
OCCURRENCE OF MINOR SEDIMENTARY STRUCTURES IN HOLOCENE DELTAIC DEPOSITIONAL ENVIRONMENTS

Sedimentary Structures	Depositional Environments							
	Distributary natural levee	Abandoned distributary channel	Point bar accretion	Interdistributary	Inland swamp	Lacustrine	Prodelta	Intradelta
Bedding								
thick >7 in.				D	D	D	D	C-D
medium 2-7 in.	C	D				A		C-D
thin <2 in.	C	C		A				C
Morphology								
parallel laminations (texture)	A	A	A	C-A	C-R	C-R	C	A
parallel laminations (color)	R			R	C	C	A-C	R
lenticular laminations						C	C-R	
wavy laminations	C-R		C					
Cross-laminations								
simple		C						A
planar	C-R	C	A					C
trough		A	R				R	A
Ripple laminations								
current	C	A	D-A			R	R	C
ripple-drift	R		A					
wave			R			C-R		C
Scour and fill	C	C-A	D					C
Plant remains								
distinct particles	C-R	R	C	C-R	A			R
finely divided			A	C-A	C-A	C	C-R	C
bedded				A	C			
Shell fragments			R	C-R	R	A	R	
Clay inclusions	R	A	C					C
Load casts	R	C						
Distorted laminations	R							R
Burrows	A-C			A	C	C	A-C	R
Nodules								
ferruginous	A-C				R			
calcareous	C			C	C			
Desiccation Features	A							
Oxidation mottling	A		R					

D-dominant; A-abundant; C-common; R-rare

Abandoned distributary channels

Few significant differences exist in terms of either sedimentary or geotechnical characteristics (Figure A7) between an abandoned course (as described earlier) and an abandoned distributary channel because the depositional processes are essentially the same. The lower portion of both is marked by a sand wedge consisting of sets of well-sorted, cross-bedded sands (SP) separated by thin clay (CH or CL) layers (Table 5). This sedimentary sequence reflects the effects of discrete pulses of discharge during major flood events.

The few differences that exist are largely with regard to the topstratum. Most deltaic distributaries received no discharge after a delta lobe was abandoned, and the vast majority of them were not reoccupied later or subsequently occupied by a smaller stream that carried an appreciable sediment load. Therefore, most were characterized by a slack-water bayou or slough with a very small suspended load. Under these conditions, only very soft, high water-content, gray to black, organic clays (CH and OH), organic silts (OL), or muck were deposited in the channels. In the more interior portions of an abandoned delta lobe, the channels sometimes became filled with intratidal marsh or swamp vegetation: in the distal areas, open water persisted in the channels, and they eventually experienced tidal flow and enlargement through bank erosion and tidal flushing. As the effects of subsidence on a delta lobe became manifest, freshwater marsh or swamp vegetation (e.g., cypress) in the channels was killed by increasing salinities.

An examination of Figure A19 reveals the greater presence of relatively soft, slack-water-deposited clays in abandoned distributaries as compared to abandoned courses. In the former, there are relatively few samples that yielded cohesive strength values over 0.5 tons/sq ft, whereas values between 0.5 and 1.0 tons/sq ft are fairly numerous in abandoned courses.

During the past several decades, both the morphology of and deposits within many abandoned distributary channels has been impacted by dredging and boat traffic since most of the larger distributaries are occupied by navigable streams. For example, virtually every major distributary is frequented by a wide variety of recreation craft, shrimp boats, oyster luggers, and oil field service floating plant. Where boat traffic is infrequent, the channels typically become densely choked and clogged with floating, nuisance, exotic aquatic plants such as the notorious water hyacinth (*Eichornia crassipes*) and alligator weed (*Achyranthes philoxeroides*). These species add significantly to the volume of organic ooze that accumulates in the channels.

Point bar accretion

Point bar accretion deposits along the present Mississippi River meander belt are present only in small areas south of Baton Rouge and are generally absent along most of the river south of New Orleans (Plates 11, 12, and 14).

Occurrences along the larger distributaries like Bayou Lafourche are even more restricted because of the limited extent of meandering.

With consideration for the downstream decrease in grain size of the river's sediment load, point bar accretion in the deltaic plain area consists of silty sand (SM) or sandy silt (SM or ML) underlain at depths of 10 to 25 ft by clean, fine sands (SP) (Kolb and VanLopik 1958). The sands are cross-bedded with current and ripple-drift laminations and frequent scour and fill structures (Table 5) as they are everywhere in the Lower Mississippi Valley area. Significant amounts of finely divided organic matter and clay balls are also present.

Probably the biggest difference between point bar accretion deposits in the alluvial valley area and those in the deltaic plain is the general absence of distinctive swales in the latter area. Because of reduced flood heights and the slow rates of accretion on bars, major depressions between ridges do not form--the ridges are welded together in a closely spaced lateral sequence with minimal topographic variation. In contrast to the situation in the alluvial valley area, natural levees often completely obscure underlying ridge and swale topography such that the limits of river channel migration must be determined by borings rather than by aerial photos.

It has only recently been determined that the extent of point bar accretion along the Bayou des Familles and Bayou la Loutre distributaries, however small, is significantly larger than formerly believed and mapped (Saucier and Dunbar 1993). Along most deltaic plain distributaries, the total lateral extent of the subaerial natural levee is indicated by deciduous, hardwood, forest vegetation, and the extent of lateral channel migration is only a small percentage of that distance. In other words, the natural levees extend well beyond the limits of the point bar deposits. However, along the streams in question, the extent of forest vegetation is essentially coincident with the extent of point bar accretion. The thin (10- to 15-ft-thick) natural levees that once extended beyond the limits of channel migration have completely subsided below sea level because of the thick, soft, compressible underlying interdistributary and prodelta deposits. In contrast, the much thicker masses of point bar deposits (60 to 80 ft thick) have been relatively much less affected by subsidence and still maintain elevations above sea level.

Interdistributary

Interdistributary deposits, while characterized at the surface by the root mats, highly organic clays, and peats of extensive fresh- to brackish-water marshes, actually form masses of largely inorganic sediments up to several tens of feet thick as a consequence of regional subsidence. Black, fibrous peats and thin, discontinuous zones of organic detritus occur throughout the sequences, but most of the sediments beneath the marsh surface consist of soft to very soft, gray, clays (CH and CL) with frequent thin lenses or paper-thin partings of silt (ML) or silty sand (SM). While organic sedimentation is important, the greatest amount of sediments by volume is the fine-grained materials from floodwaters that spill over the distributary natural levees or get carried inland by wave action and currents from distributary mouths. These sediments settle out in true vegetated marshes or in the numerous shallow ponds and lakes and bayous that characterize an interdistributary lowland. Since biological activity is intense in the lowlands, interdistributary deposits exhibit evidence of moderate burrowing and contain abundant amounts of shells from freshwater or brackish water clams such as *Rangia cuneata.*

The highest sedimentation rates in interdistributary lowlands normally occur during the early stages of delta lobe construction and decline as the lobe reaches its maximum extent and starts to deteriorate. With time, organic sedimentation becomes relatively more important; therefore, the greater amounts of silt and sand as thin lamina occur toward the base while peats and highly organic clays appear toward the top of interdistributary sedimentary sequences.

Subtle differences exist in the sediments directly underlying the vegetative mats of interdistributary marshes depending on the marsh type (i.e., freshwater versus brackish). Recognition of the precise environment in buried deposits can best be accomplished by identifying the particular species represented in the peats or organic clays (from plant remains, seeds, pollen, etc.). In general, floating marsh mats are underlain by at least several feet of muck or organic ooze. Freshwater marsh mats are underlain by variable thicknesses of clays (CH) and organic clays (OH) with peat (Pt) layers. Brackish-water marsh mats are underlain by several feet of fibrous peat which in turn is underlain by soft, gray or blue-gray clays (CH) and silty clays (CL) (Kolb and VanLopik 1958). Brackish-water deposits tend to have the higher overall organic contents, but even these do not normally exceed 10 to 20 percent of the total volume.

All organic materials in the interdistributary environment are in varying states of decomposition. Since they lie mainly below sea level in permanently flooded settings, oxidation is largely precluded and decay is largely due to anaerobic bacteria. As a result, the sediments contain large amounts of methane and hydrogen sulfide gas that contribute to their poor geotechnical properties.

Interdistributary deposits are more susceptible to consolidation as a result of human activities than those in any other fluvial or deltaic environment. The inorganic sediments are soft and have high water contents; hence, they will shrink dramatically upon drainage: with desiccation, the organic layers and lenses will shrink to a fraction of their former volume as a result of oxidation. The dramatic subsidence that has occurred in the New Orleans area during historic times (see Chapter 3) is largely a result of the oxidation of up to 12 ft of brackish-water peats and organic clays in interdistributary areas. Such deposits can compact to as little as 10 percent of their original thickness.

The geotechnical characteristics of the lower, largely inorganic portion, of sequences of interdistributary deposits are shown in Figure A8, while the characteristics of the upper, largely organic portion (marsh), are shown in Figure A9. The exceptionally low cohesive strengths of the latter are readily apparent as are the often extremely high water contents (over 100 percent dry weight). The effect of large organic contents are also manifest in the significantly lower dry and wet densities of marsh deposits.

Inland swamp

The deposits in the inland swamp environment consist of soft, gray clays (CH) and organic clays (OH) with brown peat lenses and layers of decayed wood. Scattered wood occurs throughout the clays as fragments less than an inch in size to rotten stumps, logs, and root masses. Pockets of leaf litter, including seeds, nuts, and bark fragments, are often preserved in the clay matrix as are masses of finely divided wood fragments locally referred to as "coffee grounds." The latter are accumulated by wave action along the shores of lakes and ponds and are scattered throughout swamps during storms or periods of flooding. Despite the various types of plant materials present, the total organic content of swamp deposits generally is less than 30 percent and sometimes less than in interdistributary marsh deposits (Kolb and VanLopik 1958).

Identification of the environment of deposition of lenses of peat or organic debris buried within deltaic sediments is often relatively easy. Marsh deposits are fibrous and almost always black in color, whereas swamp deposits contain wood fragments and are as often brown or reddish brown in color as they are black. In swamp deposits, leaves as old as several thousand years are often preserved to the extent that tree species can sometimes be identified by visual examination.

In the relatively less organic portions of swamp sedimentary sequences, the clays exhibit crude stratification and evidences of biogenic reworking by burrowing organisms. The sedimentary environment is also conducive to the formation of calcareous nodules and amorphous globules of calcite, but iron nodules are sparse (Table 5).

Lacustrine and lacustrine delta

Thanks to detailed sedimentologic studies of lacustrine and lacustrine deltaic deposits in the Atchafalaya Basin, including the X-radiography of numerous undisturbed cores, there is an unusually large amount of information on their lithology and sedimentary structures (Breland 1988, Coleman 1966a, Krinitzsky and Smith 1969, Tye and Coleman 1989b).

The lacustrine environment is characterized by quiet water deposition, reducing conditions, burrowing organisms, and occasional wave and current action (Coleman and Gagliano 1965). Sediments deposited under these conditions typically consist of soft, dark gray to black, highly organic clays (mostly CH) with scattered silt lenses. Where not obliterated by burrowing, fine stratification is present and includes parallel and lenticular laminations caused by thin silt (ML) lenses (Table 5). Lenses of finely divided organic matter and shells of fresh- to brackish-water clams, mostly *Rangia* and *Unio*, are abundant: fragments of backswamp deposits (ripup clasts) are present as clay balls.

The deposits described above constitute what Breland (1988) refers to as the lacustrine facies of the lacustrine subenvironment. She also recognizes a fluviolacustrine facies of the same subenvironment which forms when a distributary or newly developing course introduces a significant amount of inorganic fluvial sediments into the lake environment. This constitutes a transition between the fine-grained lacustrine and relatively coarse-grained lacustrine deltaic environments which is characterized by an appreciable increase in the number and thickness of silt lamina and layers. Otherwise, the sedimentary characteristics are essentially the same.

The lacustrine-deltaic deposits can be subdivided into a series of subaqueous and subaerial facies, including delta front and distributary-mouth bar (Tye and Coleman 1989b); however, each of these will not be separately discussed. As a whole, the sedimentary sequence coarsens upward and consists of alternating layers of gray, cross-bedded, silty and sandy clays (CL and SC), silts (ML), and fine sands (SM or SP). Since the sediments were laid down under the influence of considerable river flow-induced currents, they exhibit ripup clasts, scour and fill structures, lenticular laminations, and sometimes ripple laminations. Also present are burrows; scattered, finely divided organic matter; and calcareous nodules. Shells and shell fragments are appreciably less than in lacustrine deposits.

No data are available on the composition of lacustrine deposits outside of the Atchafalaya Basin area in lakes such as Catahoula Lake (Plate 10) or Reelfoot Lake (Plate 5); however, it is presumed they consist of heterogeneous mixtures of clays, silts, and sands with many of the same sedimentary structures found in the Atchafalaya Basin area. In the other lakes, it is anticipated that the lacustrine deposits are much thinner and discontinuous and absent in many cases.

Prodelta

Prodelta deposits consist of the fine-grained suspended sediments carried seaward from delta lobes and deposited in relatively deep Gulf waters. It has often been commented that they are some of the most homogeneous and widespread of all fluvial depositional environments (Kolb and VanLopik 1958). The deposits include firm to medium stiff, gray, massive fat clays (CH) that only rarely incorporate silt laminations and only in close proximity to the river mouth (Table 5). Small amounts of fine organic fragments are widely dispersed in the clays, and occasional shell fragments can be encountered (Kolb and Kaufman 1967).

An appreciable amount of geotechnical data is available for deposits of the prodelta environment and is presented in Figure A11. In comparing the data to that of the interdistributary environment (Figure A8), two important differences are apparent. First, the greater lithologic homogeneity of the prodelta deposits is clearly evident in the narrower range of variation in all of the measured parameters. Second, the cohesive strengths of prodelta deposits are typically higher than those of interdistributary deposits with the mean values of the former being between 0.5 and 0.6 tons/sq ft and those of the latter between 0.3 and 0.4 tons/sq ft.

Intradelta

In sharp contrast to prodelta deposits, intradelta deposits are predominantly coarse grained and form at the immediate mouth of an advancing deltaic distributary. Of the three facies of the delta front environment that are considered to constitute intradelta deposits (see Chapter 5), the distal bar is the relatively finer grained, consisting of coarsely laminated, gray silts (SM and ML) and clays (mostly CL) (Coleman and Gagliano 1965). These materials are laid down in Gulf waters peripheral to a distributary mouth but under the dominance of riverflow. The coarser sediments are typically cross-laminated with current ripples and scour and fill structures being a common occurrence (Table 5). Organic (wood) and shell fragments are relatively numerous.

Sediments of the distributary mouth bar and distributary channel are deposited within or at the immediate mouth of an advancing distributary, and hence within what is probably the highest energy regime in the deltaic plain. Massive deposits of cross-bedded silts (ML), silty sands (SM), and sands (SP) accumulate under the near-constant influence of both stream currents and wave action. Sedimentary structures typically include abundant trough cross-laminations, some wave and current ripples, and scour and fill. In a manner similar to that in abandoned courses, some clay layers deposited during low river stages survive subsequent flood-stage scour events and are preserved in the sedimentary record. Some lenses or thin layers of finely divided organic matter are present, but overall the organic content of intradelta deposits is quite low.

The geotechnical properties of intradelta deposits are portrayed in Figure A12. Paramater for parameter, there is very little difference between intradelta and interdistributary deposits. This underscores the need, in this and other deposits, to consider *all* physical (and, if possible, biological and chemical) characteristics of the materials--lithologic, sedimentologic, geotechnical, paleontologic, etc.--in identifying the depositional environment from cores.

Mudlumps

As a consequence of diapiric uplift, the prodelta deposits of which mudlumps are composed have not changed lithologically from those in an undisturbed state, but they exhibit considerably different sedimentary structures. Deep cores in mudlumps (Morgan, Coleman, and Gagliano 1963) have revealed the presence of considerable distortion, fracturing, brecciation, and faulting of the massive clay deposits as a consequence of the tens to hundreds of feet of rapid uplift. This disturbance no doubt has also affected the geotechnical characteristics of the deposits, but data are not sufficient for tabulation in Appendix A.

Holocene Marine Environments

Nearshore Gulf

Deposits of this environment vary widely in lithology, ranging from coarsely laminated to interbedded masses of firm to medium stiff, gray clays (CH, CL, and SC), shelly clays, and silty clays (CL) to well-sorted, fine-grained sands (SP and SM) and silts (ML) with only minor amounts of clays. Because of this variability, it is seldom if ever possible to make a positive identification of nearshore Gulf deposits based on lithology: the most dependable recognition criterion is the presence of large quantities of shells, shell fragments, and even shell hash. The faunal remains are indicative of a prolific shallow shelf assemblage (Parker 1956) in which dozens of species of mollusks are represented (Rowett 1957, Hollander and Dockery 1977). The shells occur both scattered throughout the sandy or clayey matrices and in thick lenses and reef-like concentrations up to several feet thick.

Evidence of intense burrowing (Table 6) is probably the second most prominent sedimentary characteristic of nearshore Gulf deposits. Biogenically reworked sediments prevail but may be interrupted by zones of laminated silts and sands with wave and current ripples, indicating periods of reworking and winnowing under marine conditions. Thus, the deposits are indicative of an environment in which depositional and erosional processes were both active.

The lowermost, or strand plain, portion of the sedimentary sequence is the most lithologically diverse. It represents the zone in which subaerial deposits of the coastal plain were overridden and reworked or buried by a transgressing shoreline driven by rising sea level. Deposits in this zone are typically a heterogeneous mixture of shells, shell fragments, wood fragments, peat, organic clay, clay balls, and nodule concentrations in a silty or sandy matrix. Borings in the nearshore Gulf environment indicate that locally the zone may contain true shell reefs, buried forests (with preserved trees, stumps, and leaf litter), beach deposits, and sands deposited in tidal inlets or drowned erosional channels.

The widely varying lithologies of nearshore Gulf deposits and those of other marine environments (discussed in the next section) have precluded any attempt to characterize their geotechnical properties. No data on any parameter are presented in Appendix A.

TABLE 6.
OCCURRENCE OF MINOR SEDIMENTARY STRUCTURES IN HOLOCENE MARINE DEPOSITIONAL ENVIRONMENTS

Sedimentary Structures	Depositional Environments				
	Nearshore Gulf	Bay-sound	Beaches and barriers	Cheniers	Reefs
Bedding					
thick >7 in.	A	D-C	C	D	C
medium 2-7 in.	C	A	A	C	
thin <2 in.		C			
Morphology					
parallel laminations (texture)	C	C		C	
parallel laminations (color)		C			
lenticular laminations	A	A		C	
wavy laminations		R			
Cross-laminations					
simple	C		A	C	C
planar	C		C		R
trough		R	C		
Ripple laminations					
current	C	C			
ripple-drift			C-A		
wave	A		C		R
Scour and fill	A	R	A	C	
Plant remains					
distinct particles	R	C	R		
finely divided		C	C		
bedded					
Shell fragments	A	A	A	D-A	D
Clay inclusions			A	C	R
Load casts					
Distorted laminations	C				
Burrows	D	A-C		A-C	
Nodules					
ferruginous			C	C	R
calcareous					
Desiccation Features					
Oxidation mottling				C-R	
D-dominant; A-abundant; C-common; R-rare					

Bay-sound

Bay-sound deposits display many of the same sedimentary characteristics as nearshore Gulf deposits, including frequent burrows and abundant shells and shell fragments (Table 6). However, the deposits are finer grained and more finely interbedded, being intermediate in that regard between nearshore Gulf deposits and lacustrine deposits. Loose to moderately dense, gray sandy silts (ML) and silty sands (SM) are the predominant soil types, but silty and sandy clays (CL and SC) occur as thin interbeds. The coarser materials exhibit thin, lenticular, parallel laminations with ripple marks and scour and fill structures indicative of reworking by tidal currents and waves in shallow water bodies. Finely divided organic matter is present in moderate quantities.

The overall shell content of bay-sound deposits is a little lower than in nearshore Gulf deposits, and the species represented are indicative of lower salinities and higher turbidities. However, this environment is biologically highly productive, and reef-like concentrations of shell and shell hash are rather frequent occurrences.

Bay-sound deposits occupy an intermediate position in a spectrum of faunal-rich aquatic environments. In terms of decreasing salinity, energy levels, and sediment grain sizes, the spectrum ranges from nearshore Gulf to bay-sound to large lake (e.g., Lakes Pontchartrain and Borgne shown in Plates 12 and 14) to medium-size lake (e.g., Lake Salvador and White Lake in Plates 13 and 14) to small interdistributary lake and pond. In the case of each parameter, however, local conditions such as lake geometry and proximity to a distributary may be an overriding factor.

Beaches and barriers

The lithology of beaches and barriers is so highly varied (Coleman 1966b) that generalization is very difficult. Most workers (e.g, Kolb and VanLopik 1958) rely on a simple categorization of these features into sand and shell types which, with clarification and amplification, is adequate for this synthesis.

The most numerically abundant beach type is the thin, discontinuous admixture of gray or tan silt (ML) or sandy silt (SM), shell, organic debris, clay balls, and nodules that accumulates along the mostly eroding shores of deltaic plain lakes and bays. The sediments are locally highly variable and derived almost exclusively from the erosion of interdistributary marsh and natural levee deposits. The accumulations seldom are more than 1 or 2 ft thick and several tens of feet wide and occur mainly in the form of pockets or scallops in an actively eroding shore. Nearly all beaches of this type throughout the deltaic plain contain shell and shell fragments, dominantly *Rangia*, but heavy accumulations almost always indicate the destruction by erosion of a nearby Indian shell midden. Such accumulations are highly conspicuous because of their white color (sun-bleached shells) and abundance of wave washed pottery and other artifacts.

Beaches and barriers of consequential size (10 ft or more thick and hundreds of feet wide) that occur along the Gulf shoreline, e.g., Grand Isle and the Chandeleur Islands (Plate 14), are mainly massive accumulations of gray to tan, well-sorted silts (ML), sandy silts (SM), and silty sands (SM) with significantly smaller quantities of shell and organic matter. The deposits are typically cross-bedded with a variety of types of laminations (Table 6) caused either by wave action or wind action in the low dunes that flank the beach. As a consequence of wave energy that is considerably higher than in more interior water bodies, the shell is commonly in the form of small fragments rather than whole shells and species typical of open Gulf conditions prevail rather than *Rangia*.

Lithologically and mineralogically, the relict barrier islands and beaches that underlie the New Orleans area (such as the Pine Island Beach Trend discussed in Chapter 5 are of a distinctly different character, being Gulf Coast barriers rather than deltaic plain barriers of the type discussed previously. Containing relatively little sediment of Mississippi River origin, they consist of thick masses of light gray to tan, well-sorted, fine to coarse sands (SP) with occasional granule-size gravels in the lower portions and fine dune sands in the upper portions (Saucier 1963). Where exposed in pits, they have been observed to have ripple cross-laminations, trough cross-laminations shelly lenses, and wavy clay laminations (Miller 1983). When exposed at the surface as they sometimes have been where dredged for highway embankment construction, the sands bleach to a brilliant white color unlike any in beaches or barrier islands of the deltaic plain (i.e., Mississippi River sediments). The beach sediments closely resemble in color and mineral composition those of the Gulf Coast east of the Pearl River, and no doubt represent a non-Mississippi eastern source.

Cheniers

Cheniers are relict beaches only slightly older (geologically speaking) than the present active Gulf beach in the same area and in the same environmental setting. Therefore, it is expected that the two would be lithologically similar. This is generally the case; however, while there are no published data to verify it, this writer has observed that the typical chenier contains a much higher percentage of shell in the form of finely divided shell hash than the present beach. The matrix of a chenier is a tan to light brownish gray, coarsely stratified mixture of shell fragments and silt and sand. Some sedimentary structures such as simple cross-laminations and scour and fill features are present, but just as often the chenier deposits are apparently massive or biogenically reworked into a homogeneous mass.

The principal differences between active beaches and cheniers in terms of sedimentary characteristics involve the higher degree of weathering and oxidation of the latter because of increased age. The presence of a woody vegetative cover for centuries to millennia has led to the formation of a soil profile with a well-developed "A" horizon in undisturbed areas.

Reefs

The principal reef-forming organism of coastal Louisiana is the American oyster, *Crassostrea virginica*. Reefs are composed of mostly massive to roughly stratified, cemented masses of live and dead oyster shells up to several tens of feet thick. Shell normally exceeds 50 percent by volume with the remainder of the matrix consisting of gray sands, silts, and marly clays. Readers interested in field observations as to reef development and patterns of growth are referred to Coleman 1966b and Kolb and VanLopik 1958.

7 Quaternary Stratigraphy and Chronology

The discussions in this part of the synthesis are arranged chronologically, from oldest to youngest, by major geologic unit. The intent is to provide an interpretation of the sequential events that led to the formation of the present landscape and the present distribution of landforms and depositional environments. Discussions are closely linked to a series of paleogeographic maps constructed to illustrate conditions at certain key intervals in Lower Mississippi Valley geomorphic history (Plate 28).

The outline of the present alluvial valley and deltaic plain, as defined by the extent of Holocene and Wisconsin stage deposits, and the location of the present Mississippi River channel are shown in Plate 28 by thin, white lines as a means of reference for the reader. The limits of landforms and geologic units (e.g., meander belts) are shown only by color breaks. Black lines, either solid or dashed, intentionally are not used because they imply a degree of precision and accuracy that is unwarranted based on available data. Similarly, the locations of abandoned courses within meander belts are not shown since each paleogeographic reconstruction represents a range in time (rather than a precise date) during which the courses often changed location and configuration.

Considerable uncertainties exist in all aspects of the geologic history of the area; however, data are available to allow the reconstructions in Plate 28 to be considered more than merely illustrative or diagrammatic. Nevertheless, they cannot and should not be used for analyses of small areas or specific locations where precise determinations are important. For example, archeologists must refrain from enlarging the maps--as they often have with those of Fisk (1944)--in an attempt to define the associations of sites with particular stream channels. Separate, detailed, geomorphic studies of small areas are necessary for accurate site/landform determinations and will remain so for the foreseeable future.

Upland Complex

Pre-Pleistocene setting

Plate 28A portrays the hypothesized paleogeographic setting of the Lower Mississippi Valley area during the very latest Tertiary (Pliocene) before the onset of the first Pleistocene continental glaciation about 2.5 million years BP (Figure 4). At that time, the wide belt of coarse clastic sediments (mostly sands and gravels) that began several million years earlier in very late Miocene times (May 1980) had reached its maximum extent. The belt spread west and south from the southern Appalachians, the Nashville Dome area, and the continental interior as a broad alluvial fan. It probably consisted of a series of lobes attributable to ancestral, braided courses of the Tennessee River (from western Kentucky to southeastern Louisiana) (Brown 1967, Potter 1955a), the Cumberland and Ohio rivers (western Kentucky and southern Illinois), and the Mississippi River (southeastern Missouri and northeastern Arkansas). Traces of the actual braided stream channels do not exist but are inferred from present surface topography and the trends of the thickest belts of graveliferous deposits. In essence, these exist as broad ridges in the interfluves between the present, entrenched upland drainage. A few of the more probable and prominent trends are shown in Plate 28A.

By the late Pliocene, the western edge of the alluvial fan reached the near-central part of the Mississippi Embayment before the process(es) that supported the outward-moving wave of sediments ceased (Plate 28A). Shortly thereafter, in the upper part of the Lower Mississippi Valley, the drainage from the Ohio and Tennessee basins probably flowed southward in a narrow valley in a shallow depression between that fan and the one created by an ancestral Mississippi River along the trend of present Crowley's Ridge. Drainage from the already well-established St. Francis, Current, Black, and White rivers of the Ozark Plateau created a small valley just east of the Ozark Escarpment which they shared with a relatively small, ancestral preglacial Mississippi River flowing from the north. The two systems probably joined somewhere south of Helena.

The Arkansas and Red rivers were well established in their present positions across western and central Arkansas and Louisiana, respectively, at that time, and most likely they became tributary to the continuation of the systems previously mentioned somewhere in the alluvial valley area. However, the present upland drainage in western Kentucky, Tennessee, and Mississippi (e.g., the Hatchie, Yalobusha, Big Black rivers) probably existed in only incipient form as headward eroding systems in shallow depressions between the braided channel trends of the alluvial fan. Similarly, the Ouachita River system of southeastern Arkansas and northeastern Louisiana probably did not exist.

It should be emphasized that all of the river valleys that existed at that time were probably narrow and shallow and perhaps not even as wide as is shown in Plate 28A (exaggerated for clarity). Moreover, their floodplains were at considerably higher elevations than the present alluvial valley surface. Whereas large parts of the alluvial fan survive as the Upland Complex, no traces of the river valleys or their deposits are preserved in the sedimentary record.

Very little is known about the position or character of the Gulf of Mexico coastline at that time, but sea levels during the late Pliocene are believed to have been lower than at present and were falling, and the low-stand shoreline was somewhere on the Louisiana Continental Shelf (Self 1993).

Early Pleistocene stage

No direct evidence exists anywhere in the Lower Mississippi Valley area (except in the deep subsurface beneath the deltaic plain) of deposits or landscapes during the initial Pleistocene glacial stage. This is the case although more than 500,000 years were probably involved based on the eustatic cycle interpretations of Beard, Sangree and Smith (1982) (Figure 4). It can be safely assumed, however, that the prolonged period of relatively low sea level (acting in consort with climate) caused the existing drainage network (described previously) to become at least moderately entrenched, but not necessarily much wider than it had been previously. It was probably during that time that the system tributaries that form the present upland drainage east of the alluvial valley attained most of their present character and extent, incising into and often through the Plio-Pleistocene alluvial fan into the underlying Tertiary formations. In general, the uplands were probably moderately dissected for the first time and transformed from a gently rolling to a moderately hilly landscape. The highest range of accordant crest elevations of the hilly interfluve areas constitute the Williana terrace surface which Fisk (1944) and Krinitzsky (1949a) interpreted as a Quaternary depositional terrace. As discussed in Autin et al. (1991), the Williana is considered by this writer to be a landscape response involving the erosion, local reworking, and partial redistribution of the alluvial fan deposits to a lowered base level.

The most important event of the early part of the Early Pleistocene Stage to affect the Lower Mississippi Valley area was the glacial derangement of Midwest drainage such that the drainage basin for the lower river was significantly increased (see Chapter 4). This probably meant a modest increase in discharges, even during times of waxing glaciation, and contributed to stream entrenchment.

As the initial Pleistocene glaciation began to wane, the small, moderately entrenched stream system through the Lower Mississippi Valley area was required for the first time to accommodate a major load of meltwater and coarse-grained glacial outwash. It is presumed that the increased discharge and sediment load caused significant alluviation and widening of the valley through the lateral shifting of braided streams (Plate 28B). A shallow valley up to several tens of miles wide and filled with several tens of feet of sands and gravels likely was present by the time the first major glacial stage came to a close. However wide the valley was at that time, it was still considerably narrower than the Wisconsin/Holocene valley. The complete absence in the uplands of any loess deposits (reworked or otherwise) correlated with these stages could be interpreted as evidence that the valley was not wide enough to allow the formation of a broad outwash plain from which large quantities of silt could be deflated.

The maximum lateral extent of the enlarged valley cannot be ascertained, but it is interpreted that in some cases it must have extended beyond the present limits of the alluvial valley. In those cases, such as north of Jonesboro and in the Vicksburg-Natchez area, inliers of glacial outwash valley-fill deposits lie stratigraphically adjacent to and only slightly lower in elevation than the dissected alluvial fan. Because of the complications created by subsequent erosion and the lithologic similarity of the two types of coarse-grained deposits, they cannot be easily differentiated. Their limits have not been mapped, but samples from scattered outcrops and pits have verified the presence in the area of distinctively different glacial and nonglacial (Appalachian) mineral suites (Rosen 1969).

There is no specific information about what conditions were like during the presumed period of basic deglaciation coincident with the high sea level between about 2,000,000 and 1,800,000 years B.P. (Aftonian Stage) (Figure 4). However, it is speculated that the floodplain of the main river system remained stable and the ancestral Mississippi and Ohio rivers developed meander belts. Where the meander belts were located within the overall valleys can only be guessed at. In the uplands, base levels of erosion slowly rose as alluvium, and colluvium accumulated in lowlands between hills and small stream valleys.

Sea level is estimated to have been at least as high as at present about 1,800,000 years B.P. and possibly considerably higher (Figure 4). However, no shoreline positions or features have been identified either at the surface or in the subsurface. Based on interpretations by McFarlan and LeRoy (1988), deposits of that age (Aftonian), overlying coarse-grained deposits of the Upland Complex, reach a thickness of at least 1,000 ft at the present Louisiana shoreline and thicken to over 3,000 ft at the edge of the continental shelf. Nothing is known regarding the depositional environments of the sediments, but they probably include materials laid down under deltaic to shallow marine conditions in a transgressive sequence. The position of the meander belt in the deltaic plain area is not known, but it is presumed that it followed the general trend of the earlier entrenched valley as identified in the subsurface by McFarlan and LeRoy (1988).

The portion of the Early Pleistocene stage between about 1,800,000 and 1,300,000 years B.P. (Figure 4), the classic Kansan Stage, represents a time of major glacial advance and low sea level. This is a period for which there is no known tangible surface evidence whatsoever in the Lower Mississippi Valley. Presumably, it was a time of major stream entrenchment, renewed landscape degradation, and shoreline regression.

It is hypothesized that the lowered base levels of erosion that were associated with the stream entrenchment and valley degradation were the impetus for the formation of the lower of the two discrete ranges of accordant crest elevations of hills in the uplands that Fisk (1944) and Krinitzsky (1949a) identified as the Bentley terrace. Like the Williana terrace, the Bentley is interpreted herein as an adjustment of the landscape, involving both erosion and deposition, but primarily the former. The deposits of the terrace, now considered a part of the Upland Complex, include surviving remnants of the Plio-Pleistocene alluvial fan and areas where the eroded materials were redeposited at relatively comparable elevations. On a regional basis, the depositional component of the terrace is poorly developed in the dissected upland areas and somewhat better developed as a coast parallel element.

In summary, the Upland Complex appears to represent a unit which, from initial deposition through two major cycles of erosion and modification, involves possibly as much as 1,700,000 years or about 60 percent of the entire Quaternary. It is the sole surviving evidence of events and processes that occurred in the entire Lower Mississippi Valley area. That long period no doubt was much more complex in terms of processes than is suggested herein, but neither the deposits nor the landscape has yet yielded useful clues. At best, the complex is a poor and indirect indicator of what was taking place along the ancestral Mississippi River.

Intermediate Complex

No direct evidence exists on which to make a glacial stage assignment for the Intermediate Complex. Herein, it is designated on the basis of deductive reasoning as late Early Pleistocene and correlated with the eustatic cycle of Beard, Sangree, and Smith (1982) that has traditionally been correlated with the Yarmouthian (Figure 4). This is a conclusion reached from an attempt to achieve a best fit between the total sequence of conceptual glacial stages (Figure 4) and the mapped terrace complexes and offshore sedimentary sequences. If it is assumed that the stratigraphically older Upland Complex represents two major eustatic cycles and the stratigraphically younger Prairie Complex is Sangamon and younger, by process of elimination the Intermediate Complex is attributable to the interglacial stage that occurred between about 1,300,000 and 800,000 years B.P. Obviously, the age assignment would be negated if either of the two assumptions prove incorrect.

The principal outcrop areas of the Intermediate Complex in the Lower Mississippi Valley proper occur as narrow bands across the central part of the Florida Parishes and north of the Great Southwest Prairies in Louisiana where they comprise the Montgomery terrace (Fisk 1938) (Plate 2). The two smaller areas located in central Arkansas (Saucier 1967) are tenuously correlated with this complex based only on relative topographic position and elevation. A similar tenuous correlation is made between the Intermediate Complex and the Humboldt and possibly the Henderson terraces of the Obion, Forked Deer, and

Hatchie rivers in the uplands of western Tennessee (Saucier 1987). Despite this limited distribution, it is suggested that the Intermediate Complex constitutes the first (oldest) widespread interglacial-stage valley-fill sequence in the area of the type envisioned by Fisk (1938) and Russell (1938) as depositional terraces.

It is presumed that the initial (and basal) portion of the valley-fill sequence (of which no evidence is apparent) consisted of outwash deposits laid down by braided streams during full glacial conditions. Those became mixed through complex scouring and filling with older Early Pleistocene outwash, but since the valley became wider and presumably deeper, much of the earlier sediments may have been flushed out and replaced with the younger ones. Subsequently, with landscape stability resulting from a relatively high sea level and interglacial climate, major rivers adopted a meandering regime and a basically fine-grained topstratum was laid down in at least the lower portions of the alluvial valley and into its tributaries. As shown in Plate 28C, it is hypothesized that areas of valley train characterized at least portions of the alluvial valley landscape as they have in later interglacial periods. Once again, readers are cautioned that the actual locations and limits of the valley trains shown in Plate 28C, along with those of the probable meander belts and flood basin areas, are largely hypothetical since no direct evidence exists.

The actual fluvial environments in which the sediments were deposited have not been determined for any of the outcrop areas. It is surmised that those in Arkansas represent valley train or alluvial apron environments while those in Louisiana represent alluvial aprons composed of sediments eroded from the Upland Complex and deposited at the edge of the Coastal Plain. The Humboldt and Henderson terraces represent undifferentiated valley-fill sequences along those small rivers that are graded to the higher-than-present base level that existed in the Mississippi alluvial valley at that time.

Several important inferences can be drawn from the overall distribution of the Intermediate Complex, albeit limited. It appears that for the first time during the Quaternary, the floodplain level in the ancestral Mississippi alluvial valley, while higher than at present by several tens of feet, was well below the average elevation of the uplands on both sides. This implies that erosion, degradation, and stream entrenchment during the preceding major glacial stage had been appreciable and extensive. Crowley's Ridge probably became a prominent topographic feature for the first time. Additionally, the possible presence of fluvial terraces along upland streams east of the alluvial valley would certainly substantiate that dissection of the uplands was well advanced with the present drainage pattern essentially established. In this regard, it should also be noted that the Intermediate Complex is well represented along the Ouachita and Saline rivers in southern Arkansas, indicating the advanced state of development of that fluvial system as well (Saucier and Snead 1989).

As in the case of the Upland Complex, no interglacial-stage shoreline positions or features have been discerned indicating the maximum extent of transgression or the extent to which a deltaic plain formed during the still stand.

Beneath the deltaic plain, McFarlan and LeRoy (1988) recognize a Kansan-Yarmouthian Stage sedimentary sequence (Figure 4) consisting of a basal, coarse-grained regressive unit (substratum) and an upper, fine-grained transgressive unit (topstratum) representing coastal plain, deltaic, and/or shallow marine environments. The sequence progressively thickens southward from the outcrop areas to the edge of the Louisiana Continental Shelf and has a thickness of about 800 ft at the Louisiana shoreline in Terrebonne Parish.

The extremely limited extent of Intermediate Complex outcrops suggests that stream erosion and entrenchment and valley degradation were intensive and extensive during the subsequent Illinoian Stage (Middle Middle and Early Middle Pleistocene Stages, Figure 4). The stage is estimated to have lasted from about 800,000 to perhaps as recent as 125,000 years B.P. (Figure 5). This is not unexpected if as indicated in Figure 4, the stage involved three separate major glacial cycles, each with a significant eustatic fluctuation. On the other hand, it cannot be concluded that this glacial stage necessarily was more complex than the preceding ones: it may be that the older ones were equally as complex, but the details have yet to be resolved.

It is nevertheless reasonable to conclude that at the time of the maximum extent of glaciation, the alluvial valley had been degraded with most if not all of the fine-grained topstratum (representing multiple fluvial environments) destroyed by lateral stream migration and vertical incision. At that time, the western side of the alluvial valley locally exceeded its present limits (e.g., the Grand Prairie did not exist as a terrace), whereas the eastern side was not as far east as it is today. Much of the floodplain surface may have been a braided stream- dominated sandy plain which was at or slightly below the level of the Holocene surface.

If tenuous loess sheet identifications and age estimates of various workers are correct (as reported by McCraw and Autin 1989), the early Illinoian Stage marked the deposition in the uplands of the first loess sheet for which there is surviving evidence. Various names have been assigned to this oldest loess sheet, but according to the interpretation used herein, it is the Marianna loess which thus far has been identified only on Crowley's Ridge. According to the same source, the second oldest loess sheet would correlate with the latter part of the Illinoian Stage. That deposit, designated herein as the Crowley's Ridge loess but referred to others as the Loveland loess, has been identified on Crowley's Ridge and also in the uplands near Vicksburg. Considerable work will be necessary before the total distribution of these loess sheets is known, but at least some present interpretations suggest the presence of modest-sized valley trains in the Western Lowlands and the central parts of the alluvial valley area at that time.

Prairie Complex (Sangamon Phase)

Heretofore in this part of the report, a simple, logical sequence of discussion based on major geologic units (e.g., the Intermediate Complex) has been possible because of their relative temporal relationships. However, at this point the approach breaks down because as discussed in Chapter 5, the Prairie Complex spans several major glacial stages. It is necessary therefore to interrupt the discussion of this unit to introduce others in their proper chronological sequence. For example, the Early Wisconsin valley trains are younger than the initial Prairie Complex deposits but older than the later ones.

For the first time in the Quaternary history of the Lower Mississippi Valley, there is ample evidence of the environments, landforms, and processes representing a glacial cycle. With this evidence and by analogy from later cycles, it is possible to reconstruct a sequence of events with *reasonable* confidence although major questions still remain.

Initial Prairie Complex deposits

During the probable time of major waning of the Illinoian glaciation about 160,000 to 150,000 years B.P. (Figure 5), the upper part of the alluvial valley experienced the deposition of large volumes of glacial outwash from both the ancestral Mississippi and Ohio river systems in the Western and Eastern Lowlands, respectively (Plate 28D). A valley train developed in both lowlands to an elevation 15 to 30 ft or more above the present surface (Wisconsin-age valley trains) (Smith and Saucier 1971, Saucier 1987). The highest levels of the valley train in the Western Lowlands occurred immediately west of Crowley's Ridge where several small remnants have been identified based on elevation and surface morphology (Plates 6 and 7). If a recent identification of both Peoria and Loveland loess sheets is correct (Rutledge, West, and Omakupt 1985), the entire extent of the adjacent highest level of the Early Wisconsin valley trains (Pve3 level, Plates 6 and 7) would probably represent glacial outwash deposited during the waning of the Illinoian glaciation. Accordingly, the small outcrops, designated herein as part of the Prairie Complex, instead may actually be part of the Intermediate Complex.

West of that area, lower (and presumably slightly younger) valley train levels extended across the Western Lowlands, beneath the Grand Prairie at a depth of several tens of feet below the present surface, and into the Arkansas Lowland. From there, the surface of the valley train dipped southward, lying at a depth of about 80 ft below the surface of the Prairie Complex near the Arkansas-Louisiana state line, and about 150 ft below the surface of the complex near Lafayette, Louisiana. Beneath the central part of the deltaic plain and onto the continental shelf, outwash deposits (coarse-grained substratum) have been identified in wells at depths of 500 to 1,000 ft (McFarlan and LeRoy 1988).

As a direct consequence of the rapid deposition of outwash in the upper part of the alluvial valley, gradients on most of the larger upland rivers shallowed appreciably just upstream from the bluff line. In western Tennessee, this led to the formation of the Hatchie terrace which is believed to be a temporal equivalent of the Prairie Complex (Saucier 1987). No counterparts have been identified on Crowley's Ridge or in the Ozark Uplands but may not have formed in the latter area because of the steep stream gradients.

Late transgressive phase deposits

Major outwash deposition in the upper part of the Lower Mississippi Valley is assumed to have markedly declined or ceased during the final waning stage of the Illinoian glaciation from about 150,000 to 135,000 years B.P. (Figure 5). In the lower part of the valley, rising sea level caused a reduction in the gradient of the ancestral Mississippi River, and it can be presumed that it changed from a braided to a meandering regime with overbank sedimentation becoming dominant. If sea level rise was rapid, the extreme lower part of the alluvial valley may have even become a shallow embayment.

Widespread overbank sedimentation at that time is evident by the presence of extensive backswamp deposits directly overlying what had been the lower levels of glacial outwash beneath the Grand Prairie and the Prairie Complex surface along the western side of the valley between Pine Bluff and Monroe (Plates 2, 7, 8, and 9). These deposits eventually attained a thickness of 60 to 80 ft, representing slow overbank deposition that probably continued uninterrupted up to or slightly beyond the time of the maximum sea level high stand.

By 135,000 years B.P., the ancestral alluvial valley had a mature floodplain that was wide enough and gradients no doubt shallow enough for multiple meander belts to have formed. However, no direct evidence exists for where either the Mississippi or the Ohio Rivers was located. The meander belt configurations shown in Plate 28D are largely just illustrative. In contrast, there is a clear record of where the Arkansas River was flowing at that time. A series of well-preserved distributaries, complete with natural levees, abandoned courses, and abandoned channels (Plate 7), indicate that that river, rather than being confined to a narrow valley, had developed an extensive alluvial fan out into (and overlying) backswamp deposits. That development is compatible with the view that the alluvial valley was a mature and stable landscape. In the Red River valley, no meander belts of this age have been identified, but Russ (1975) believes that the Sangamon Stage floodplain is represented by the higher of the two Prairie Complex terraces that have been recognized and mapped.

In the uplands adjacent to the Lower Mississippi Valley, the Sangamon Stage is regarded as a period of prolonged stability. As was the case throughout much of the midwestern United States, that stage is evidenced by the formation of a distinctive, characteristically reddish soil that is generally referred to as the Sangamon soil or the Sangamon geosol. That horizon has proven to be a valuable stratigraphic marker for use in the correlation of loess and terrace deposits.

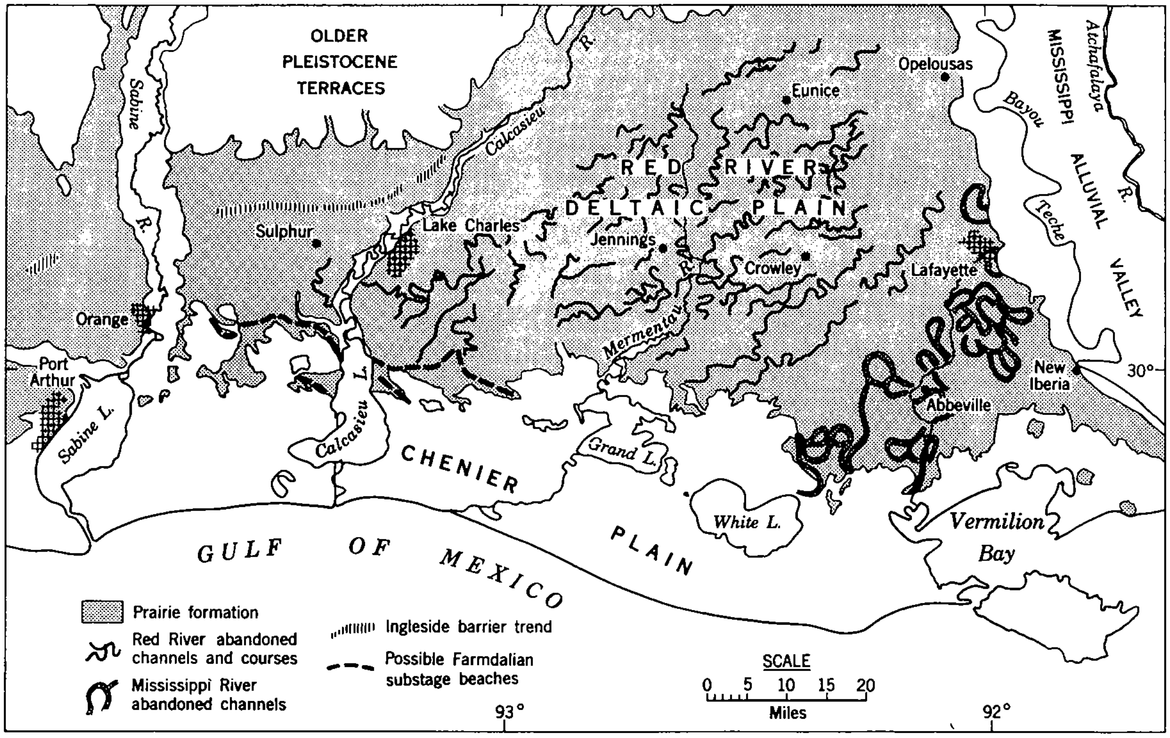
With regard to the coastal zone, the high sea level stand that is believed to have occurred between about 135,000 and 130,000 years B.P. (Figure 5) attained a level (10 or more feet above the present) that would have driven the Gulf shoreline well inland from its present position. No shoreline features have been identified that relate to this stage, but they probably exist just south of the outcrop of the Intermediate Complex where they are buried beneath more recent alluvial and colluvial deposits. Seaward from them, a thin layer of sediments was laid down in nearshore Gulf and related marine environments, thickening toward the edge of the continental shelf. These are evidenced by a basal, fossiliferous zone of the Prairie Complex which has been detected beneath the Great Southwest Prairies and in the Florida Parishes area between Baton Rouge and the Pearl River -- the Biloxi formation of Otvos (1991).

If the glacioeustatic sea level record illustrated in Figure 5 is correct, the high sea level stand was followed by a brief fall of several thousand years duration. Its effect on the coastal area is not apparent in the sedimentary record but presumably would have caused some shoreline regression and shallowing of offshore areas and possibly some minor stream entrenchment. These conditions would have quickly reversed when sea level rose again and attained an elevation even higher than its penultimate level. Although the second highstand about 120,000 years B.P. is believed to have been quite brief, it is very likely the event (in Oxygen Isotope Stage 5e) that caused the formation of Houston Ridge and the related beaches and barrier islands of the Ingleside Barrier Trend in southwestern Louisiana (Autin et al. 1991) (Figure 47). A stratigraphic equivalent of the Ingleside Trend has not been positively identified in southeastern Louisiana; however, thick masses of sand revealed in borings at a depth of 100 ft and more in the Lake Pontchartrain area may represent a barrier trend that has been strongly displaced vertically by faulting and subsidence (Kolb, Smith, and Silva 1975). The probably orientation of the trend is shown in Plate 28D. A shallow sound may have existed between the offshore barrier and a mainland shoreline to the north: the location of the latter has not been detected.

"Eowisconsin" Stage Events and Deposits

Background

More than 25 years ago, this writer (Saucier 1968) recognized that the Lower Mississippi Valley area had responded to two rather than just one significant glacioeustatic cycle since deposition of the earlier Prairie Complex units during the Sangamon Stage. That realization was what prompted assigning various components of the complex to both the Sangamon and the



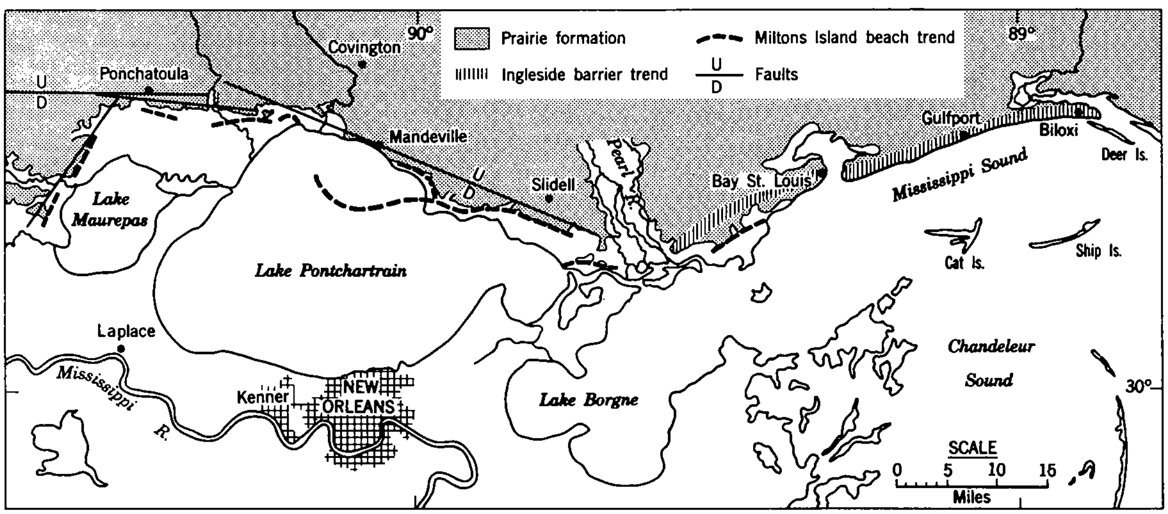


Figure 47. Locations and early interpretations of the ages of fluvial and marine features of the Prairie Complex in southwestern and southeastern Louisiana (from Saucier 1977b)

Wisconsin stages and eventually to differentiation of the complex into various alloformations. It also gave rise to considerable emphasis on eustatic events of the Middle Wisconsin as a means of reconciling the Prairie Complex stratigraphic sequence and its cut and fill cycles with the sea level curve that was most widely accepted at the time. More recently, the stratigraphic relations among the various components of the complex have been somewhat more clearly defined and new components discovered, but they have not been more precisely dated. It can only be stated that they definitely appear to be attributable to the 100,000-yr-long period between about 120,000 years B.P. and the last glacial maximum about 18,000 years B.P. Compounding the problem has been the development of a sea level curve that is far more complex than what was envisioned even 20 years ago (Figure 5). Thus, the temporal positioning of several units and features of the Prairie Complex are still very much argumentative.

A brief review of the units and some of the relevant stratigraphic relations is in order and is presented in the following paragraphs. Afterwards, this writer offers for the first time a new and significantly different, tentative chronological concept. The proposed scenario is certain to be highly argumentative because it is not compatible with all observed evidence. Two lines of evidence--cross-cutting relations on one hand and loess stratigraphy and pedogenesis on the other--are presently unreconcilably in conflict. However, this writer introduces the new scenario for consideration and debate because he feels the interests of science will not be served by avoiding the issue.

One component of the Prairie Complex of questionable age includes the Red River deltaic plain in the area known as the Great Southwest Prairies (Figure 47 and Chapter 6). The plain was formed when the Red River, flowing separately from the Mississippi River, apparently rapidly extended a series of courses and distributaries out of the mouth of its valley through the Ville Platte/Opelousas, Louisiana, area and thence southwestward at least as far as the present Gulf shoreline in southwestern Louisiana. The pattern of relict deltaic channels has been mapped in detail in part of the area of this study (Plates 11 and 13), and the overall picture is presented by Saucier and Snead (1989). In terms of morphology, lithology, and causal processes, the Red River deltaic plain is highly analogous to the broad alluvial fan which the Arkansas River created across the Grand Prairie region during the Sangamon Stage. The deposits of the plain directly overlie assumed Sangamon-age nearshore Gulf and related marine deposits (Biloxi formation) and surround and bury eastern extensions of the Ingleside Barrier Trend.

Other major units of the Prairie Complex in question are the several segments of Mississippi River meander belts in south central Louisiana. One is the Avoyelles Prairie, also known as the Avoyelles Hills or Marksville Hills, located east of Alexandria (Plates 2 and 10). The elevation of the surface of the unit is below that of the Prairie Complex outcrop just to the northwest, suggesting that it formed later than the time of the Sangamon high sea level stand. In turn, it is higher and, based on firmly established cross-cutting relationships, is *older* than an area of Early Wisconsin Stage valley train just to the north. That relationship is perhaps the single most important one in terms of developing the new scenario as noted in the following discussion. In addition, the unit is veneered with 10 ft or more of Peoria loess, but only one loess sheet.

Another major meander belt unit in question is the Lafayette Meander Belt in the Lafayette-Abbeville area of Louisiana (Figure 47, Plates 11 and 13). That meander belt segment, formed by the Mississippi River, is part of a fluvial trend which is inset into Sangamon Stage marine deposits of the Prairie Complex and which is stratigraphically younger than the Red River deltaic plain unit to the west. It also is overlain by 10 to 12 ft of Peoria loess. Although separated by nearly 50 mi, the Avoyelles Prairie and the Lafayette Meander Belt apparently are part of the same trend. They have similar lithologies and soil profile development, and the downvalley gradients are consistent.

The third meander belt unit of the Prairie Complex in question is manifest by the outcrop located just downstream from Mt. Pleasant Bluff north of Baton Rouge (see Chapter 6 and Plate 11). Its lithology is similar to that of the other two meander belt units, it is covered with a thin veneer of Peoria loess, and it is flanked by and underlain by Sangamon Stage deposits. There can be no doubt that it is of the same origin and same general age as the other units.

In addition to the deltaic plain and meander belt units, there are certain features of nearshore marine origin on the Prairie Complex which, although not units (alloformations) in the same sense as those described above, are relevant to this discussion. This writer (Saucier 1977b) first called attention to a series of small beach ridges or shoreline positions along the edge of the Prairie Complex in several parts of south Louisiana (Figure 47). South of Lake Charles, two trends occur 20 to 30 mi south of the Ingleside Barrier Trend and obviously formed at a later time and at a slightly lower sea level position. North of New Orleans, the Miltons Island trend includes a series of well-defined beach features between the mouth of the Pearl River and Ponchatoula, Louisiana, and continues into the area west of Lake Maurepas. The features south of Lake Charles could have formed during either a sea level transgression or a still stand in a regression but definitely are stratigraphically older than the Red River deltaic plain unit of the Prairie Complex. The features were originally attributed to the high sea level stand about 30,000 years B.P. (Middle Wisconsin).

Tentative new concept

All of the units/features described above apparently formed during a relatively high sea level stand (within a few feet above to a few tens of feet below present level) and reasonably are associated with a single (but not necessarily simple) major eustatic cycle. At one time or another, *all* have been attributed to the Middle Wisconsin (e.g., Saucier 1977b, Autin et al. 1991), but that interpretation is no longer viable in terms of presently discerned stratigraphic relations. By process of elimination, their age now appears bracketed between about 120,000 and 70,000 years B.P., or the "Eowisconsin" (Figure 5).

During Oxygen Isotope Stage 5d, a major eustatic sea level fall must have caused the Gulf shoreline to recede well offshore from the Ingleside Barrier Trend and related shorelines, particularly in southwestern Louisiana where the continental shelf is wide and shallow. It was perhaps during brief still stands slightly above present sea level in the early part of that eustatic cycle when the small beach ridges and shoreline features south of Lake Charles and some of the ones *north* of Lake Pontchartrain formed in southern Louisiana. In the latter area (Figure 47), the part of the Miltons Island Trend actually *beneath* Lake Pontchartrain (Saucier 1977b) is excluded because it lies incorporated in a sedimentary sequence which, as will be discussed later, is still considered to be of Middle Wisconsin age. If the sea level curve shown in Figure 5 is correct, the early part of Stage 5d would have been the only time since the Sangamon that sea level was high enough to allow those features to form.

Figure 5 does suggest, however, that an alternative hypothesis may be possible provided that a variation of a few feet is permissible in the interpretation of the actual high sea level stand elevations reached during the late Illinoian and Sangamon stages. It is possible that the Ingleside Barrier Trend correlates with the high stand between about 135,000 and 130,000 years B.P. and the younger and smaller beach ridges, and some of the Miltons Island Trend correlate with the high stand that occurred about 120,000 years B.P. That possibility would be stronger if the earlier sea level high stand was slightly higher than the later one rather than slightly lower as is now indicated.

During the subsequent Oxygen Isotope Stages 5c and 5a (Figure 5), sea level rose significantly but not to elevations attained earlier during the Sangamon. Nevertheless, it is hypothesized that the eustatic rises were sufficient to drive the coastline inland across the continental shelf and to shallow the gradients of Coastal Plain streams plus the Red and Mississippi rivers. During one of those two stages, which must have involved rapid alluviation in the lower alluvial valley area, the Red River was able to spread its sediments as a broad, thin sheet across what, at that time, was a low, flat, emergent coastal plain composed of Sangamon Stage nearshore Gulf and related marine deposits, including the features described previously. Hence, the Red River deltaic plain was formed.

Slightly later, the Mississippi River, flowing just to the east, created at least two meander belts, one including the Avoyelles Prairie and Lafayette Meander Belt segments and the other the Mt. Pleasant Bluff segment (Plate 28E). The Mississippi River was graded to a sea level perhaps a few tens of feet lower than at present, and the delta must have been offshore from the present Louisiana coastline well south of Lafayette. Both meander belts could represent the same stage (either 5c or 5a) in a situation analogous to the Holocene when several discrete meander belts have formed in a several-thousand-year period. However, because of unresolved stratigraphic interrelationships, it is not known if the Red River deltaic plain and the Mississippi River meander belts represent the same or different stages. The former could correlate with Stage 5c and the latter with Stage 5a, or they both could correlate with either.

The hypothesis presented above is apparently compatible with known geomorphic and subsurface stratigraphic evidence, but as stated previously, it is in conflict with one aspect of loess stratigraphy. If the Red River deltaic plain and the meander belt segments are as old as the "Eowisconsin," they should be veneered with two loess sheets rather than just one. The Roxanna loess does not extend far enough south in the alluvial valley area for its presence to be expected, but the Sicily Island loess should be present along with the Peoria loess if the landforms are older than the Early Wisconsin. However, only the Peoria is present. A possible explanation for this situation is that the Sicily Island is actually pre-Early Wisconsin in age and therefore too old to be present on the Prairie Complex, a possibility that is being considered by some.[1]

Before this discussion ends, it is important to note that this tentative concept or model is based entirely on observations of landforms and deposits in southern Louisiana. At present, there is no direct evidence whatsoever of what conditions may have been like elsewhere in the alluvial valley area during the "Eowisconsin," and this writer has not had the opportunity to test the model against stratigraphic evidence from elsewhere in the Gulf Coast area. Therefore, the conditions outside of the deltaic area portrayed in Plate 28E must be viewed as entirely hypothetical and illustrative.

Onset of Early Wisconsin glaciation

Following Oxygen Isotope Stage 5a, sea level is believed to have begun a steady net decline in elevation, culminating with a major low stand about 70,000 years B.P. during the peak of the Early Wisconsin glaciation (Stage 4). That eustatic fall no doubt caused another major recession of the Gulf shoreline well offshore onto the continental shelf and an episode of weathering, oxidation, and stream entrenchment near the coast. Using an extensive, high-resolution, seismic reflection profiling data set, Berryhill and Suter (1986) have recognized two stratigraphically separated networks of incised, fluvial channels off central and western Louisiana to a distance of 200 mi from the present shoreline. They interpret the deeper of the two networks to be entrenchment of local streams into the Prairie Complex surface during the Early Wisconsin Stage, although Coleman and Roberts (1988) suggest that it occurred at an earlier time (Illinoian Stage?). This writer proposes that the deeper network represents Early Wisconsin Stage entrenchment into the "Eowisconsin" deposits, which would include the Mississippi River delta that formed south of the Lafayette Meander Belt.

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[1] Personal Communication, 1994, Whitney Autin, Assistant Professor, Louisiana State University, Baton Rouge.

In several major studies and delineations of the entrenched Pleistocene surface beneath New Orleans and the Lake Pontchartrain area (Kolb and Van-Lopik 1958, Saucier 1963, Schultz and Kolb 1950), the surface of the shallowest, oxidized, weathered, entrenched horizon underlying the Holocene deposits has traditionally been considered the top of the Prairie Complex. It was regarded as simply the subsided and downfaulted seaward continuation of the Pleistocene surface exposed immediately north of the Pontchartrain Basin (Plate 12). However, Saucier (1977b) postulated that the equivalent of the Sangamon Stage portion of the Prairie Complex was actually the second (deeper) of two buried, weathered, erosional surfaces underlying the basin area. Later detailed interpretations of a large subsurface data set (Kolb, Smith, and Silva 1975) appear to substantiate that view. Although the control on the second horizon is limited because of its considerable depth, it slopes seaward from an elevation of about -60 ft msl along the north shore of Lake Pontchartrain to about -160 ft msl along the Mississippi River near New Orleans. Shallow stream entrenchments are indicated by lines of closely spaced borings, but contouring of their trends has not been possible. The second horizon has also been identified in wells farther south beneath the deltaic plain and has been mapped at depths of 1,000 to 1,200 ft at the present Louisiana shoreline (McFarlan and LeRoy 1988).

It is proposed herein that the second weathered horizon occurs on the top of "Eowisconsin" Stage deposits rather than Sangamon Stage deposits and was formed during the Early Wisconsin Stage sea level low stand. If this new interpretation is correct, still a third horizon within the Prairie Complex sedimentary sequence is conceptually possible--one that would have formed on Sangamon Stage deposits during Oxygen Isotope Stage 5d. There are only a very few engineering borings in the greater New Orleans area that are sufficiently deep (greater than 200 ft) to encounter still a deeper horizon, and they indeed suggest that one may be present.

At the southern terminus of the alluvial valley, some dramatic paleogeographic changes took place. Possibly in direct response to eustatically lowered base levels, the Mississippi River abandoned the Avoyelles Prairie, Lafayette Meander Belt, and Mt. Pleasant Bluff meander belt trends, changed from a meandering to a braided regime, adopted a course through the area of the present Atchafalaya Basin, and began downcutting and degrading its valley. As a consequence, it began forming the low escarpment that marks the eastern side of the Prairie Complex outcrop west of and parallel to the present Teche Ridge (Plates 11 and 12). By the time of the Early Wisconsin glacial maximum, the escarpment must have been at least several tens of feet higher than it is today.

Outside of the deltaic plain area, there is no direct evidence in the Lower Mississippi Valley area of any events, processes, or deposits that correlate with waxing Early Wisconsin glaciation. However, it is presumed that both valley degradation and aggradation were involved as the area underwent major changes in climate and base levels. By inference, it is believed that large volumes of deposits of the Prairie Complex were removed by lateral and vertical stream cutting over most of the area of the present (Holocene and Wisconsin Stage) alluvial valley and the floodplain was slightly lower than at present. Yet, a large volume of valley fill (substratum) probably still remained. Locally, small areas of the Prairie Complex survived the planation and entrenchment and remain as "islands" surrounded by more recent deposits such as those near the eastern edge of Macon Ridge (Plates 8 and 9). Pulses of outwash must have periodically entered the area; however, the net effect on the landscape was erosional rather than depositional. It is evident that most of the thick topstratum of overbank sediments (e.g., backswamp and natural levee) was destroyed by fluvial action with only areas such as the Grand Prairie remaining as terraces above the lowered floodplain level. As has been indicated previously, it is likely that coarse-grained deposits of Sangamon age constitute at least part of the present graveliferous substratum but are indistinguishable from similar deposits of other ages.

A post-Sangamon (post-"Eowisconsin"?) Stage episode of stream incision is apparent in virtually all Lower Mississippi Valley tributaries of all sizes since they responded directly to lower base levels. Gullies formed and eroded headward into Prairie Complex deposits along the margins of the main valley while drainage systems in older formations widened, deepened, and lengthened. If multiple entrenchment events actually took place during the 50,000-year period between the Sangamon and the Early Wisconsin, which is certainly plausible, they have not been resolved in the stratigraphic record.

Early Wisconsin Stage Events

Until recently, it has been convenient to consider that the transition from Oxygen Isotope Stage 5a to 3 represented a steady change from essential nonglacial (major interstadial?) to glacial and back to nonglacial conditions with a smoothly falling and then rising sea level curve. However, it now appears from a variety of continental and oceanographic evidence that the 40,000-year-long period from about 70,000 to 30,000 years B.P. involved a number of stadial/interstadial-scale events (Figure 5). In reality, the response of the Lower Mississippi Valley to these glacial and glacioeustatic events was probably much more complex than what is presented next, but the evidence in the stratigraphic record, if present, has not been commensurably resolved.

Valley train formation

The primary manifestation of the waning of the Early Wisconsin glaciation was the deposition of large volumes of glacial outwash into a valley that had previously been degraded and entrenched to some extent following deposition of the earlier phases of the Prairie Complex. Outwash deposition affected practically all of the alluvial valley area except the Arkansas Lowland and spread onto an exposed continental shelf beneath the present deltaic plain. Valley trains that mark the maximum extent of valley aggradation are extensively preserved, forming the present surface of Sikeston's Ridge, a part of the western St. Francis Basin, the vast majority of the Western Lowlands, Macon Ridge, and several small areas in western Tensas Basin (Plate 1).

As part of his determination that all of the valley trains (braided-stream terraces or surfaces) of the alluvial valley area were not formed immediately prior to the Holocene during a single phase of waning glaciation, Saucier (1968) discovered that Macon Ridge and valley trains near Catahoula Lake (Plates 9 and 10) had been truncated and incised by a major episode of valley degradation that occurred prior to the deposition of Holocene overbank deposits (primarily backswamp). In that area, the stratigraphic relationships cannot be explained by a simple change in river regime from braided to meandering at the end of the Pleistocene or by events such as lateral stream migration during the Holocene. Assignment of the older valley train segments to the Middle Wisconsin Stage (Farmdalian Stage, Figure 4) was supported by two radiocarbon dates on freshwater shells from a braided channel fill on Macon Ridge (31,200 to 29,100 years B.P.).

There has been much discussion in the literature that radiocarbon dates in the 35,000- to 30,000-year B.P. range, especially those on shells, are anomalously old because of contamination with "dead" materials. However, irrespective of the dates, compelling evidence is now available indicating that all of the valley trains designated herein as Early Wisconsin are separated by at least one major glacial cycle (the Woodfordian or Late Wisconsin Stage, Figure 4) from the younger ones. The presence of a well-developed Peoria loess sheet on Macon Ridge and in portions of the Western Lowlands is one chronostratigraphic indicator that cannot be disputed. All of the evidence substantiates a minimum age of at least 25,000 years for the valley trains but does not preclude them being as old as 60,000 years, however.

As discussed in Chapter 5, the valley trains have multiple levels, with three being easily recognizable in the Western Lowlands and four on Macon Ridge. This writer has postulated that the cyclical downcutting by braided streams (valley degradation) that led to the formation of the discrete levels was due to an overall decrease in the ratio of outwash to meltwater, acting through pulses of sedimentation related to irregular glacial withdrawal and decay (Saucier 1974). As glacial chronological reconstructions become more precise, it is becoming more tempting to believe that the valley train levels may be a direct reflection of glacial advances and retreats in the Midwest that are related (but not directly so) to the sea level variations shown in Figure 5. If this is the case, the valley trains may represent much longer periods of outwash deposition than previously believed. Readers are cautioned, however, that it is *not* presently suggested that a particular level can be correlated with a specific sea level high stand as indicated in Figure 5.

It is postulated herein that the valley train comprising Sikeston's Ridge and the area adjacent to the eastern side of Crowley's Ridge was deposited during the *earlier* part of Oxygen Isotope Stage 3 (Figure 5) by the Ohio River as it discharged from the Cache Lowland (Plate 1) into the St. Francis Basin. This tributaries east of the alluvial valley. In western Kentucky and Tennessee, the Finley terrace has been correlated with Early Wisconsin Stage outwash deposition in the St. Francis Basin area (Saucier 1987). Like the Lake Monroe Complex, the Finley terrace has a nil gradient for at least 20 mi upstream from the edge of the alluvial valley. That, along with the presence of shoreline features, indicates the existence of an open lake for a considerable period of time.

The Early Wisconsin Stage Mississippi River valley train extended across the mouth of the Arkansas Lowland (Plate 1), but there are no indications that lacustrine conditions developed in that area. The appreciable discharge of that stream, in contrast to the streams mentioned above, was probably sufficient to maintain an efficient outlet for the discharge. Nevertheless, it is highly probable that the gradient of that river was shallowed to some extent as a result of the outwash deposition, but an environment no wetter than a backswamp probably persisted through the period.

Middle Wisconsin Stage Events

The decline in outwash deposition and valley degradation that was apparently taking place late in Oxygen Isotope Stage 3 (Figure 4) (e.g., after the highest levels of Macon Ridge were deposited) could be indicative of the onset of a period of waxing glaciation and relative sea level fall rather than a trend toward deglaciation, but that interpretation is not made herein. The end of the Early Wisconsin Stage is considered to have been quite different than the end of the Late Wisconsin Stage wherein sea level continued to rise for several thousand years after outwash deposition ceased.

This writer believes there is varied and pervasive evidence in the Lower Mississippi Valley area which supports the view that a relative sea level low stand about 35,000 years B.P. was quickly followed by a brief but significant interstadial high stand (Farmdalian Stage) about 30,000 years B.P. (Figure 4). That view is held despite cited evidence to the contrary from various parts of the world (Bloom 1983) and the strong dissenting opinion of at least one vocal critic (Otvos 1991, Otvos and Howat 1991). While it is conceded that sea level did not quite reach its present level at that time (see Chapter 3), the eustatic rise nevertheless was apparently sufficient to cause an appreciable shallowing of the gradient of the Mississippi River and its floodplain south of the latitude of Memphis. The following discussion is based on that paradigm.

There is no apparent manifestation of the Middle Wisconsin Stage in the upper and central part of the alluvial valley area. The Mississippi River definitely did not change from a braided to a meandering regime anywhere in the Western Lowlands area, and there are no relict meander belt segments anywhere in the St. Francis or Yazoo Basin areas that could date to this period. As shown in Plate 28F, it is hypothesized that the Ohio River started meandering in the latter basin area, but the Mississippi River did not do so until appreciably farther south. Drainage from the uplands probably flowed for the most part separately in small meander belts that followed the lowest parts of the alluvial plain but must have eventually become tributary to the larger river systems.

It is hypothesized that the initial effect of the Middle Wisconsin transgression in the coastal area probably was to force the Gulf shoreline inland in the lower part of the alluvial valley, creating an estuarine like embayment or a backswamp/lacustrine environment like that of the present Atchafalaya Basin. A Mississippi River deltaic plain probably did not exist, and in southwestern Louisiana, the Gulf shoreline transgressed across and truncated (ravinement) the distal parts of the "Eowisconsin" Stage Red River and Mississippi River deltaic plains (Plate 28F).

There is abundant evidence in southeastern Louisiana of a seaward-thickening wedge of nearshore Gulf deposits that overlies the weathered, entrenched, erosional surface (Saucier 1977b) which herein is interpreted to have formed on "Eowisconsin"-aged deposits. In the New Orleans-Lake Pontchartrain Basin area, the sedimentary sequence occurs between the first and second weathered Pleistocene horizons (geosols) according to Kolb, Smith, and Silva (1975). Based on hundreds of borings that penetrate the sequence, it consists of a complex, interfingered mass of clays, silts, and sands which in its lower portions is strongly fossiliferous, indicating deposition in a nearshore marine environment. Finite radiocarbon dates ranging from 29,300 to 27,000 years B.P. have been obtained on marine shells from the deposit (Saucier 1977b).

The wedge of marine deposits varies from over 100 ft thick in parts of the New Orleans area, where it fills entrenchments in the underlying surface, to as thin as 20 ft beneath the northern part of Lake Pontchartrain. Massive beach and barrier island deposits, including the Miltons Island Trend, apparently accumulated in locations similar to those of the earlier Sangamon Stage. This is not unexpected since the general configuration of the offshore profile did not significantly change and the sediment source remained the same, i.e., large quantities of sands were brought to the coast by the Pearl River and carried westward by prevailing longshore currents.

This same sedimentary sequence has been identified off southwest Louisiana (Berryhill, Trippet, and Mihalyi 1982) and elsewhere beneath the deltaic plain using geophysical logs and other techniques (McFarlan and LeRoy 1988). In the former area, it probably consists primarily of the sediments that were eroded and reworked from the Red River and Mississippi River deltaic plains. The sequence has been interpreted as having a maximum thickness beneath the present shoreline of about 600 ft and thickens several hundred feet more to the edge of the continental shelf. Overall, this is considerably thinner than older glacial/interglacial sequences, possibly a reflection of the relatively short period of time involved during the interstadial.

The most problematic aspect of the sedimentary sequence is its shallow elevation in the Lake Pontchartrain area. Although nowhere does it lie above present sea level, extensive areas occur as shallow as -20 ft msl (or even less, depending on the interpretation) south of where it pinches out completely against the Baton Rouge Fault Zone (Kolb, Smith, and Silva 1975). With an allowance for the effects of subsidence south of the fault zone (Saucier 1963) this would indicate a sea level very close to its present level, a possibility that is strongly contradicted by widespread evidence from other parts of the world (see Chapter 3). That is the strongest argument presented by Otvos (1991) to support his contention that the sedimentary sequence in question must be Sangamon rather than Middle Wisconsin in age. A suggested, but *very* tenuous, resolution of this dilemma lies in the observation that none of the unequivocal beach features and obviously marine components of the sedimentary sequence are shallower than about -40 ft msl. Therefore, the possibility cannot be dismissed that the upper 20 ft or so of the sedimentary sequence represents non-marine alluvial apron or colluvial deposits from the uplands to the north which were laid down during the subsequent Late Wisconsin Stage low sea level stand.

Late Wisconsin Stage Valley Entrenchment

Onset of the Woodfordian glacial stage (Figure 4) and initial development of the Laurentide ice sheet in North America perhaps began as early as about 38,000 years B.P. but certainly was well advanced by 25,000 years B.P. That date marked still another episode of climatically induced valley degradation, as well as stream entrenchment and shoreline regression in the coastal area due to a major fall in sea level.

As in other glacial stages, the largely erosional episode is marked in most areas by a break (hiatus) in the stratigraphic record. Although there was possibly a slight net decline in the mean discharge of the Mississippi and Ohio rivers due to large volumes of water being trapped in the continental glacier, periodic pulses of meltwater and outwash and increased flows from tributaries (see following section) were probably sufficient to cause the rivers to maintain or change to a braided regime as far south as the Gulf of Mexico. Irrespective of regime, it is obvious that significant volumes of preexisting outwash were flushed from the Lower Mississippi Valley area, especially in its southern portion.

In the alluvial valley area, there is some direct stratigraphic evidence that the rivers shifted laterally and locally impinged against and eroded the valley margins. For example, a considerable area of Early Wisconsin outwash between Greenville and Sicily Island (Plates 8 and 9) was removed, creating the slightly scalloped eastern margin of Macon Ridge. By the time of the last glacial maximum about 18,000 years B.P., the ridge margin was probably an escarpment at least 50 ft high since the level of the active floodplain was significantly lower than it is today (Saucier 1968).

In the coastal area, the Middle Wisconsin nearshore marine sedimentary sequence emerged as a coastal plain and became progressively more exposed to subaerial weathering and erosion as the Gulf shoreline rapidly regressed southward. These events marked the beginning of the formation of the first horizon (shallowest Pleistocene paleosol) that now underlies the Holocene deltaic and chenier plain sedimentary sequences, a process that proceeded from north to south as sea level dropped. The erosional surface and paleosol are almost always easily discernible in borings beneath the upper deltaic plain but become progressively less distinct and harder to identify toward the south where the surface was exposed to subaerial weathering for a much shorter period of time.

With consideration for the effects of subsidence, the contours on the top of the Pleistocene surface (the suballuvial surface) as shown in Plates 24 to 27 provide an indication of the configuration of the surface up to the time of maximum glaciation. Beneath the chenier plain, the surface is so flat and gently sloping that the elected 25-ft-contour interval does not adequately reflect its character, but this is all that available data will allow. Beneath most of southeastern Louisiana, the situation is much worse since the extremely sparse borings and wells allow only a 50-ft contour interval and the contours are highly generalized. However, thanks to the extraordinary density of subsurface control in the greater New Orleans area and recent, large-scale mapping by this writer (Saucier et al. 1984, 1991b), a much better indication of the true character of the Pleistocene surface can be obtained from Figure 48. This figure, prepared as an example of what conditions are probably like beneath much of the deltaic plain, contains a 10-ft-contour interval and is based on an evaluation of thousands of borings.

Figure 48 clearly reveals the pattern of drainage that developed on the emerging surface as local streams extended their courses southward in pace with the regressing shoreline although the pattern is somewhat complicated by the more recent entrenchment of the modern Mississippi River channel. It is important to note that the entrenchments are quite narrow and steep sided with an overall dendritic pattern but were sufficiently wide to allow the formation of narrow floodplains underlain by coarse-grained alluvial fill. Interfluve areas are broad and flat to gently rolling, typically with not more than 5 ft of local relief. In general, the buried surface is similar to and perhaps even less dissected than presently exposed portions of the Prairie Complex immediately to the north of the deltaic plain.

Entrenchments that can be mapped in detail in the New Orleans area, where borings are typically only hundreds of feet apart, no doubt extend to the edge of the continental shelf. However, the entrenchments cannot be discerned much less mapped where borings are generally many tens of thousands of feet apart. Nevertheless, there are indications that the drainage does not become progressively more deeply incised toward the south as was once believed (Fisk 1944). Rather, most streams probably were no more deeply entrenched than about 50 ft into the Pleistocene deposits with their thalweg profiles paralleling that of the general trend of the coastal plain surface. Thus, because of course lengthening, stream gradients changed very little rather than sharply steepening as once believed.

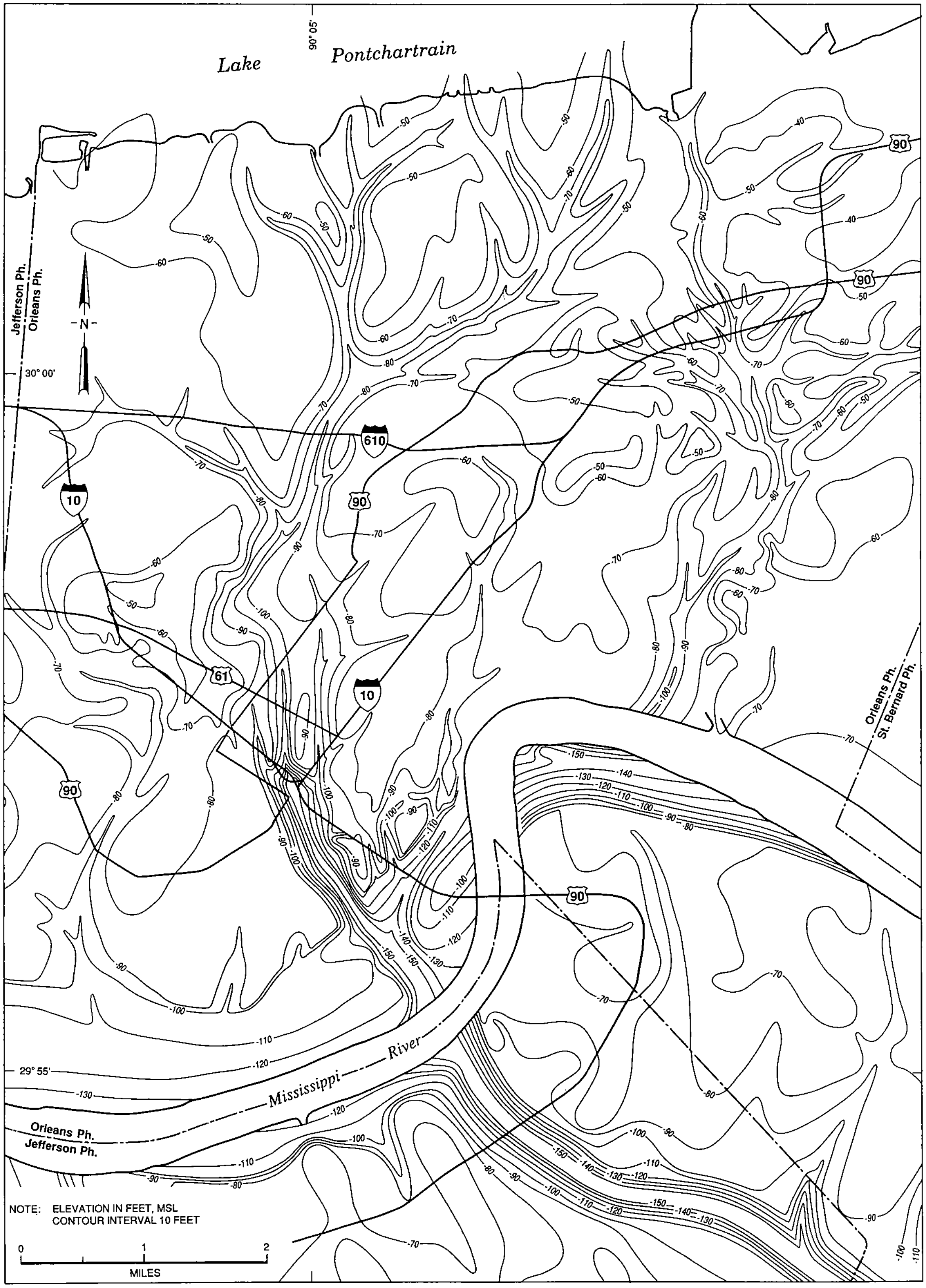

Figure 48. Configuration of the entrenched surface (first horizon) on Wisconsin-age deposits of the Prairie Complex as mapped on the New Orleans East and Spanish Fort 1:24,000-scale quadrangles

The behavior of the Mississippi River (which included the discharges of the Ohio River and tributaries) during that time was strikingly different from the local streams described previously. Flowing in a braided regime, the river shifted widely within its wide, box-shaped, trough-like alluvial valley, constantly reworking the coarse-grained substratum deposits and occasionally cutting into and truncating the suballuvial surface. South of the present "mouth" of the alluvial valley along the so-called Donaldsonville-Franklin line (Plates 12 and 13), the entrenched valley trended southward and southeastward and flared out as it trended past the present shoreline and into the Gulf. The western margin of the now-buried entrenched valley is well defined and can be traced by closely spaced contours from near Franklin to the western side of Point au Fer Island (Plate 26). The eastern side of the entrenched valley south of Donaldsonville apparently is less well defined and unfortunately occurs in an area with very few borings; however, it appears that it lies just east of the present course of Bayou Lafourche (Plate 27).

Mapping of the configuration of the suballuvial surface within the Mississippi entrenchment (Plates 26 and 27) has been difficult, and the indicated contours are extremely tenuous because of a general lack of subsurface control. Unfortunately, the vast majority of available data are from petroleum exploration wells, and many of these are clustered over and around salt domes where the suballuvial surface often has been significantly uplifted by diapirism. Other forms of data used were the geophysical logs of wells where it was only possible to identify the probable base of the substratum rather than the erosional surface *per se*. As indicated in Chapter 5, it is nevertheless apparent that the surface of the entrenchment does not exhibit the integrated, dendritic pattern that was postulated by Fisk (1944) (Plate 15) and is present on the shallower portions of the Pleistocene surface to the east and to the west. Rather, the surface is highly irregular with several reasonably well-documented enclosed depressions.

In due consideration of the uncertainties of the limited data and allowing for subsidence, it appears that the mean maximum depth (elevation below sea level) of the Mississippi entrenched valley at the Louisiana shoreline was only about 300 ft rather than more than 700 ft as postulated at one time (Fisk and McFarlan 1955). Depths as great as 400 ft occur (Plates 15, 26 and 27) but appear to be limited to enclosed depressions (scoured areas) inland from the present shoreline. To assume that the valley contained an appreciable alluvial fill even at the time of the last glacial maximum will mean that the floodplain within the valley was well above the eustatically lowered sea level. From the Louisiana shoreline, the floodplain must have sloped rather steeply southward, and the Mississippi system discharged into the Gulf near the head of the Mississippi submarine canyon at the edge of the continental shelf.

Deweyville Complex

As discussed in Chapter 3, this writer believes the Deweyville Complex is principally the manifestation of climatic conditions peculiar to certain atmospheric circulation patterns that existed under late waxing to near-full glacial stage conditions. Accordingly, he bases the discussions that follow on the premise that all occurrences of the geological unit throughout its geographic range are of approximately the same age. This is in contrast to the positions of others who have implied a lack of contemporaneity as a means of defending their widely varying age estimates. The presence of this complex on streams of a particular size (i.e., distance inland of the headwaters) and its absence on others appear to be indicative of the causal processes responsible for this extraordinary Gulf and Atlantic Coastal Plain phenomenon.

Extrapolating from paleoecological studies (e.g., Royall, Delcourt, and Delcourt 1991), it is reasonable to believe that beginning no later than about 25,000 years B.P., the more inland portions of the Lower Mississippi Valley area experienced the beginning of a trend toward higher water tables, increased soil moisture, and a significant increase in effective runoff. Those conditions were directly attributable to changes in the amount and/or seasonality of precipitation and a decrease in mean annual temperatures. This trend intensified with time and reached a maximum development probably about 18,000 years B.P. Pollen and plant macrofossil remains indicate that the sand flats of the alluvial valley itself and the adjacent uplands to at least as far south as the Tunica Hills of the western Florida Parishes of Louisiana were vegetated with white spruce (*Picea glauca*) and larch (Delcourt and Delcourt 1977, 1981). A recent reevaluation of the fossil locales, however, indicates that the *Picea* remains are not typical of any extant species and probably represent either an extinct species or an extinct varity or subspecies (Jackson and Givens 1994). Therefore, it is possible that the climate of the area was cooler than present but not necessarily as cool as implied by *Picea glauca* or other "boreal" taxa.

Streams like the Arkansas, Red, Sabine, and Pearl rivers have headwaters that are at least several hundred miles inland from the present coast. The Deweyville Complex on these streams has been interpreted as evidence that discharges greatly exceeded those of the present during an interval believed to be coincident with maximum glaciation and for several thousand years thereafter. Those streams meandered widely and created a series of fluvial terraces as they episodically degraded their entrenched valleys in response to lowering base levels. On the latter two streams, the Deweyville Complex slopes seaward and is apparent on both sides of the valleys to as far south as the inner margins of the chenier and deltaic plains where it disappears beneath Holocene intratidal marsh deposits. Evidence is available from high-resolution seismic surveys and cores that the complex is present to at least 10 mi offshore at a depth of 60 ft below present sea level along the continuation of the Sabine River valley (Coastal Environments, Inc. 1986). There is no reason to doubt that the Deweyville Complex is present along the entrenched valley possibly to near the edge of the continental shelf at depths in excess of 300 ft.

On streams whose headwaters are closer to the present coast, such as the Calcasieu in southwestern Louisiana and the Amite and Tangipahoa rivers in southeastern Louisiana (Plate 12), the Deweyville Complex is very poorly developed or completely absent. This is a possible indication that the full-glacial climate within a few hundred miles of the coast at that time was no wetter than at present and probably appreciably drier. Coastal zone aridity would also explain the complete absence of the Deweyville Complex along the entrenchments of local streams beneath Lake Pontchartrain and New Orleans. A dissenting opinion on the origin of the complex has been offered by Alford and Holmes (1985) who believe that the complex was formed during interstadial or interglacial times and that the large discharges were the product of increased tropical storm activity.

With the coastal area still under consideration, it is important to point out that there is no recognized, exposed coast parallel equivalent of the Deweyville Complex. If the chronologic positioning of the complex by this writer is correct, none should exist. At one time, Kolb, Smith, and Silva (1975) postulated that the Wisconsin-age, nearshore marine deposits beneath the first weathered Pleistocene horizon in the New Orleans area was the coastal equivalent of the Deweyville Complex. This argument was based in part on radiocarbon dates from the deposits as old as 30,000 years B.P. which would place formation of the complex at a time of relatively high sea level. Although a few dates that old (and some even older) indeed do exist, the ones with the more reliably established provenience cluster in the 25,000- to 18,000-year-B.P. range. Regardless, stratigraphic evidence is very strong that the Deweyville Complex formed during a time of waxing to near full-glacial conditions and a falling sea level: consequently, it must be younger than the Middle Wisconsin Stage.

In and adjacent to the alluvial valley area, the Deweyville Complex is best developed along the Ouachita River where it is definitely younger than the Lake Monroe Complex but older than the initial post-glacial valley aggradation. Along that river system, the complex correlates best with full-glacial conditions (e.g., about 18,000 years B.P.), and archeological site distribution suggests a cessation of significantly higher-than-present stream discharges no later than about 14,000 years B.P. (Weinstein and Kelley 1984).

The only other significant manifestation of the Deweyville Complex is along the western margin of the Grand Prairie where the Arkansas River meandered into and truncated a part of the Prairie Complex (Plate 7). Nothing is known about the trend of the more downstream course of the river while flowing in the larger-than-present meander belt, however. Based on conditions upstream in western Louisiana, it is apparent that the Red River similarly occupied an oversized (as compared to present) meander belt at that time.

All three of the streams mentioned previously are assumed to have maintained a widely meandering regime into the alluvial valley proper and probably up to the points where they became tributaries of the Mississippi River. However, the latter stream was braided at the time and probably assimilated the flow of the tributaries without a change in character. The braided nature of the

Mississippi River at the time is the inferred reason why there is no manifestation of typical, Deweyville-like cutoffs or meander scroll topography anywhere in or adjacent to the alluvial valley or associated with the entrenched valley beneath the deltaic plain.

Valley-wall scallops caused by streams with larger-than-present meander radii have been observed along some Mississippi River tributaries in western Tennessee (Saucier 1987), but their correlation with the Deweyville Complex is tenuous. Elsewhere in the northern part of the Lower Mississippi Valley, there are no fluvial terraces of comparable age. However, a stratigraphic equivalent may be present in the Western Lowlands area. The large alluvial fan of the Current River, lying between the Ozark Escarpment and the present course of the river northeast of Pocahontas, Arkansas (Plate 5), displays a series of six paleochannels as shown in Plate 8. Archeological sites associated with the channels suggest that the third oldest is at least 12,000 years old, but no dates are available for the two oldest ones (Price et al. 1981). It is interesting to note that those two (and only those two) display morphometric parameters suggesting considerably higher-than-present discharge. Stratigraphically, the alluvial fan overlies the youngest level of the Early Wisconsin valley train and theoretically could be nearly 30,000 years old; however, a much younger age is more likely. It is hypothesized that the two oldest paleochannels formed between about 18,000 and 14,000 years B.P. during a time of higher-than-present stream runoff from the Ozark Plateau and hence are temporally correlated with the Deweyville Complex.

Detailed mapping of Quaternary deposits along rivers within the Ozark Plateau area has been accomplished only along the Current River (Saucier 1983). Therefore, it is not possible to make a definitive statement about the absence or presence of terraces of the Deweyville Complex in that region. It is known only that along the Current River, the process of terrace formation is fundamentally related to alternating episodes of floodplain stability and instability. If a Deweyville Complex equivalent existed, it has been destroyed by more recent valley flushing. Unfortunately, it has thus far been impossible to correlate events within the valley to features on the alluvial fan.

Late Wisconsin Stage Valley Trains

18,000 to 12,000 years B.P.

A date of 18,000 years B.P. is generally accepted by most workers as the time of the last eustatic sea level low stand (Figure 5) and probably, but not necessarily, approximates the time of the maximum extent of the last glaciation in North America (i.e., the Laurentide ice sheet). In the Lower Mississippi Valley area, it was the time of the maximum exposure and entrenchment of the Middle Wisconsin deposits on the continental shelf and the maximum degradation of the alluvial valley area. Although the climate was cooler and significantly wetter than at present, it was also a time of extensive seasonal loess deposition and sand dune formation (Saucier 1978). Valley trains with at least ribbons of fresh outwash as sandy plains apparently were sufficient to provide the silt that became incorporated as the lower portion of the Peoria loess sheet in all parts of the area.

Beginning about 18,000 years B.P., the Laurentide ice sheet began a phase of wasting and rapid retreat with a major increase in the amount of meltwater and outwash coursing southward through the alluvial valley into the Gulf of Mexico. The volume of glacial runoff apparently increased progressively but in an episodic manner for several thousand years thereafter. Based on estimates using entirely different lines of evidence (Emiliani, Rooth, and Stipp 1978; Teller 1990), by about 14,000 years ago the volume of flow through the Mississippi Valley area probably exceeded that of the present by a factor of at least five.

High-resolution evidence has only recently become available for how the alluvial valley area responded to an episode of outwash deposition. A pollen core has been obtained from the Powers Fort Swale which is located in the Western Lowlands about 15 mi southwest of Poplar Bluff (Plate 5) (Royall, Delcourt, and Delcourt 1991). The strategic location of this site allows the reconstruction of a series of events that affected both the Western Lowlands and the St. Francis Basin area (Saucier, in press).

Between 18,000 and about 16,000 years B.P., most of the outwash being transported by the Mississippi River moved from the central Mississippi Valley area through the Drum and/or Advance Lowlands (Plates 1, 4, and 5) into the Western Lowlands. Some also probably spilled through a narrow gap in Crowley's Ridge into the Morehouse Lowland. The braided course of the river at the time was primarily along the route of the Cache River, but some flow was probably following the route of the present Black River (Plates 5 to 7). It appears as though the braided channels were rather narrow, and most of the outwash was being funnelled through the Western Lowlands rather than being deposited on a broad floodplain in that area. Not much is known about the course of the Ohio River at that time, but by inference it can be stated that its braided course trended through the Cache Lowland of southern Illinois into the alluvial valley area and turned southward and stayed east of Sikeston's Ridge (Plates 4 and 5). The area of mapped braided channel, with large areas of sand dunes, shown immediately east of the ridge may reflect the location of the river at that time. As far south as the latitude of Memphis, the Ohio River channel probably closely followed the base of the valley wall more or less along the route of the present river. In both parts of the alluvial valley, the floodplain, where vegetated, supported a forest characterized by boreal species such as spruce (*Picea*), northern pine (Diploxylon *Pinus*), and fir (*Abies*) (Royall, Delcourt, and Delcourt 1991). Dry-season winds were strong and contributed to the formation of sand dunes and heavy loess accumulations in the uplands within and east of the alluvial valley.

An initial impact of the influx of glacial outwash must have been the buildup (aggradation) of the floodplain in the upper part of the Western Lowlands. By an extrapolated date of about 16,300 years B.P., meltwater flow past the Powers Fort Swale area markedly declined, suggesting the floodplain aggradation caused the initial partial diversion of the river through the rapidly widening Bell City-Oran Gap (Plates 1 and 4). That process must have proceeded swiftly because by an estimated 14,700 years B.P., meltwater flow into the Western Lowlands became quite sporadic with that area thereafter serving only as an ephemeral sluiceway. Although his age estimates were wrong, Fisk (1944) (Plate 21) correctly reconstructed and nicely illustrated the sequence of events that led to the diversion of the river through the gap into the Morehouse Lowland area.

As the volume of flow through the Bell City-Oran Gap into the upper part of the St. Francis Basin rapidly escalated, the river reworked a large area of Early Wisconsin Stage outwash and constructed a new valley train at a slightly lower elevation than had existed previously. The pattern of braided channels that were active at that time (i.e., about 14,000 years B.P.) is well preserved on the upper of the two valley train levels as mapped in Plates 5 and 6 (Smith and Saucier 1971; Saucier, in press). Although the pattern is quite evident, it is not possible to say which of the several discrete channels were active at any given point in time.

Based on inferential evidence, Esling, Hughes, and Graham (1989) believe that the Ohio River abandoned the Cache Lowland at about that time and adopted its present course past Cairo (Plates 4 and 5). No valley train deposits in the St. Francis Basin can be specifically attributed to the Ohio River; however, it is evident that the braided river flowed near the base of the bluffs and probably merged with the Mississippi River in the valley constriction between about Memphis and Helena. South of Helena, the valley train of the combined rivers probably formed first along the eastern side of the Yazoo Basin where the highest remnants remain. Subsequently, the locus of deposition shifted westward to the vicinity of the present Mississippi River meander belt (Plates 7 and 8) and then back eastward to near the center of the basin. This sequence is largely conjecture based on slight differences in elevation and inferences from later meander belt positions.

The distribution of archeological sites in the Yazoo Basin area provides only an estimate of the minimum ages of the valley trains in that area. Sites dating to the late Paleo-Indian and Early Archaic periods are associated with the outwash deposits exposed in the northeastern and west central portions of the basin area (Plates 7 and 8) (Brain 1970, Lauro 1980). These sites establish only that the surfaces existed and were suitable for permanent habitation by about 11,000 years B.P. The sites do not establish how much earlier these surfaces may have been created.

Both Midwestern stratigraphic evidence and oceanographic evidence indicate that glacial runoff through the Lower Mississippi Valley area escalated in the form of a series of increasingly stronger pulses until a peak was reached about 12,000 years B.P., plus or minus about 500 years. Thereafter, the runoff declined abruptly. The most likely position of the rivers at that time is indicated in Plate 28G. It should be noted that whereas climate had begun to ameliorate a few thousand years earlier, the date of 12,000 years B.P. can be considered the last manifestation of basically glacial conditions. The local precipitation regime that gave rise to the Deweyville Complex had probably ended by about 14,000 years B.P., but the climate was still cooler and wetter than present and major constituents of a boreal forest were still present in the alluvial valley area. A date of 12,000 years B.P. can also be viewed as the effective end of loess deposition, and no sand dunes are known to have formed after this time.

South of the latitude of Vicksburg, the positions of the braided channels of the major rivers are unknown due to burial beneath Holocene deposits. However, the 12,000-year-B.P. floodplain slopes southward and lies at a depth of 100 to 120 ft opposite Baton Rouge. The actual mouth of the river within the entrenched alluvial valley was probably only a relatively short distance south of this point.

Meltwater flow and outwash deposition were the dominant processes and they heavily flavored the character of the landscape in the St. Francis, Yazoo, Tensas, and Atchafalaya basin areas between 14,000 and 12,000 years B.P., but sharply different conditions existed only short distances away. For example, upon eventual complete cessation of glacial runoff through the Western Lowlands, the combined flow of the Black and St. Francis rivers occupied (or remained as the only flow within) the abandoned braided channel along the route of the present Cache River. Therein, it flowed in a tightly meandering regime, the meander belt of which now constitutes the Cache River terrace (Plates 5 and 6). From the distribution of archeological sites dating to the Paleo-Indian and Early Archaic periods (Figure 2), it is apparent that the meander belt was abandoned no later than about 12,000 years B.P. and the rivers shifted to their present courses.

In the Arkansas Lowland, it is hypothesized that the Deweyville Complex floodplain was abandoned by at least 14,000 years B.P. with the Arkansas River undergoing a metamorphosis as it adapted to a smaller discharge. It is probable that prior to 12,000 years B.P., the river was forced to flow southward west of Macon Ridge because of the blockage at the mouth of the lowland area by Early Wisconsin Stage outwash (a northern extension of Macon Ridge). According to chronologic reconstructions based heavily on archeological site distribution, the seven delineated and numbered meander belts shown in Plates 7 to 9 all postdate 12,000 years B.P. (Autin et al. 1991). However, at several locations, such as west of Dermott, Arkansas (Plate 8), there are faint traces of short segments of older meander belts that are heavily veneered with backswamp deposits. If the age estimates of the later meander belts are correct, it is likely that the largely buried ones date to the 14,000- to 12,000-year-B.P. period and possibly even earlier.

The character of the lower Red River valley south of Alexandria during the Late Wisconsin is a complete mystery. There is no possibility of extrapolating backward from Holocene meander belts since all identified ones are almost certainly no older than 5,000 years. Because the Red River responded to the appreciable glacial-stage degradation in the Mississippi River entrenched valley, its floodplain prior to 12,000 years B.P. was 40 ft or more lower than at present, and no traces remain of former channels. All that can be surmised is that the Red River never flowed in a braided regime because of a load of glacial outwash, and "Deweyville conditions" ceased by about 14,000 years B.P.

12,000 to 10,000 years B.P.

Climate-wise, a date of 12,000 years B.P. effectively marks the end of glacial conditions and the beginning of the postglacial period. At that time, a rather sudden shift in atmospheric circulation patterns initiated a strong trend toward significantly warmer and drier conditions in the Lower Mississippi Valley area. Correspondingly, that date marked the effective demise of boreal forest species and their replacement with oak-hickory-dominated, mixed, deciduous hardwood forests in upland areas and cypress-gum swamps in lowland areas in the southern part of the valley. However, it did *not* mark the end of glacial runoff into the alluvial valley area and the deposition of outwash since the Laurentide ice sheet had not yet retreated from the Great Lakes region.

If the interpretations of some Midwestern geologists are correct (Teller 1990), glacial runoff into the Lower Mississippi Valley declined dramatically after 12,000 years B.P. and ceased by 11,000 years B.P. This was caused by meltwater from the Laurentide ice sheet being diverted from the Mississippi River valley to the St. Lawrence Valley outlet. It is likely that during a several-hundred-year period, the discharge to the Gulf of Mexico was no greater than at present and possibly even less.

Because of an ice readvance in the Great Lakes area about 10,500 years B.P., however, the St. Lawrence Valley outlet closed and meltwater flow returned for the last time to the Lower Mississippi Valley, albeit for a short period of time. There are differences of opinion as to whether the return flow lasted until 10,000 or 9,500 years B.P. (Saucier, in press), but nevertheless there were important geomorphic consequences.

The return of flow to the Lower Mississippi Valley after the brief lull was probably not in the form of a catastrophic flood event (such as would have been caused by the bursting of an ice margin lake), but flow increased sharply and was probably the event that triggered the Thebes Gap Diversion. In the manner described and illustrated by Fisk (1944) (Plate 21), meltwater flow was able to spill through and rapidly widen and deepen a small existing upland drainage feature in the Commerce Hills (Plate 1), thereby diverting the full flow of the Mississippi River into the portion of the St. Francis Basin lying east of Sikeston's Ridge.

Formation of the Charleston Fan immediately downstream from Thebes Gap (Plate 1) was an initial and direct effect of the diversion. This small but classically shaped alluvial fan of only about 150 sq mi in extent was first described by Ray (1964) and more recently investigated and cored by Porter and Guccione (1994). Radiocarbon dates obtained on the fan do not precisely date its formation (a short time prior to 10,590 years B.P.), but it is unlikely that its total life span exceeded several hundred years.

Initiation of full Mississippi River flow through Thebes Gap (and the formation of the Charleston Fan) is probably more accurately dated by way of a strategically located Dalton-culture archeological site situated on a low rock bench or ledge on the eastern side of the gap. Known as the Olive Branch Site (Gramly and Funk 1991), this cultural deposit could not have accumulated as it did unless the jointed rock of the ledge on which it is located had been flushed clean of residual soil by high flows through the gap. Furthermore, the deposit would not have been preserved unless the prehistoric occupation took place after the flows subsided to approximate present volumes. Radiocarbon dates indicate initial site occupation occurred about 10,000 years B.P.; consequently, it is hypothesized by this writer that formation of Thebes Gap and the Charleston Fan occurred between about 10,500 and 10,000 years B.P.

There is no indication whatsoever of the effects, if any, the post-10,500-year-B.P. discharge had farther south in the alluvial valley area. It is quite likely that the event was so brief and of such small magnitude that the outwash was assimilated by the braided channel regime and caused no channel changes or separately recognizable depositional units beyond the limits of the St. Francis Basin. Outwash pulses of this and even considerably greater magnitude probably were a common if not typical occurrence during the entire period of postglacial runoff.

At about 10,000 years B.P., the St. Lawrence Valley outlet reopened and was the sole route to the sea of the final runoff from the Laurentide ice sheet. In the Lower Mississippi Valley area, it is well established that the oldest backswamp deposits date to about 9,900 to 9,800 years B.P. in the St. Francis Basin area (Guccione, Lafferty, and Cummings 1988), indicating that the Mississippi River quickly adopted a meandering regime once meltwater flow ceased and the discharge of the river dropped to its present volume.

Archeological sites also provide verification of the age of some of the older valley train deposits in the St. Francis Basin area. The highest level of Late Wisconsin outwash (the Pvl2 surface, Plates 5 and 6) contains dozens of Dalton culture (Paleo-Indian period) sites, especially along its eastern edge (Saucier, in press). These sites are unequivocal proof that that valley train level had been abandoned as a site of active outwash deposition by at least 11,000 years B.P. and probably more than 12,000 years ago. In contrast, the lowest valley train level (Pvl1), deposited after the Mississippi River broke through the Bell City-Oran Gap, contains no Dalton-culture sites whatsoever. This absence of sites is interpreted as an indication (but not proof) that the lowest level was the site of active outwash deposition and therefore unsuitable for permanent human habitation between about 11,000 and 10,000 years B.P. It should be noted that an absence of sites on a particular level can also denote that they are present and have not been found, or that they were present and were later destroyed: in this instance, neither of these alternative explanations appears to be viable.

Holocene Transgression

Buried Pleistocene horizon

In the alluvial valley area, the generally accepted date of 10,000 years B.P. for the termination of the Pleistocene epoch and the beginning of the Holocene agrees extremely well with the well-defined stratigraphic break caused by the cessation of outwash deposition and the regime change of the Mississippi River from braided to meandering. However, in the coastal area, as pointed out so well by Russell (1940), that chronologic boundary is transparent in the sedimentary record. The entire period from about 18,000 to at least 4,000 years B.P. was characterized by a progressive rise in sea level, interrupted by brief still stands or even minor falls. It is possible, but unlikely, that any of the reversals in the overall trend coincided with either the abrupt change in climate at about 12,000 years B.P. or the termination of glacial runoff to the Lower Mississippi Valley at about 10,000 years B.P.

Starting about 18,000 years B.P., the Gulf of Mexico shoreline began moving inland (transgressing) across the weathered, eroded, and entrenched surface formed on deposits of Middle Wisconsin age. By about 12,000 years B.P., sea level is estimated to have been about 140 ft below present (Stright 1990), and the shoreline about in the position indicated in Plate 28G. Sea level was probably rising at the rate of several feet per century at that time; therefore it is deemed unlikely that any beaches or other strand line features formed on the surface. Rather, the shoreline probably exhibited only a small beach ridge which was retreating inland at a rate of tens to hundreds of feet per year. Existing upland forests were being destroyed directly by coastal erosion while riverine forests on the floodplains of the entrenched valleys were being killed by saltwater intrusion. It has been speculated by many that under these conditions, the coastal zone of Louisiana essentially lacked intratidal marshes and reefs and associated biologically productive ecosystems that are unable to adapt to rapidly changing water levels. If true, this has important archeological implications since the first human populations had already arrived in the region.

However swift the shoreline retreat was across the low and flat portions of the Coastal Plain, such as in southwestern Louisiana, the planation that took place in the littoral and shallow-water offshore zones was not sufficient to remove the several-tens-of-feet-thick weathered zone that developed on the Pleistocene deposits. In shallow depressions and across small entrenched valleys, a thin veneer of nearshore Gulf and/or strand plain deposits was laid down on top of the weathered horizon. These facts explain the exceptional preservation in the stratigraphic record of the pattern of stream entrenchment discussed earlier.

As indicated previously, it is likely that at about 12,000 years B.P., Gulf waters extended well inland into the Mississippi River entrenched valley but not as far as the latitude of Baton Rouge. It is improbable that any type of deltaic plain existed, and the mouth of the Mississippi system more closely resembled a drowned river mouth/estuary like the present Alabama-Mobile River mouth in upper Mobile Bay, Alabama. Extensive backswamp and lacustrine environments are known to have already been in existence in the upper part of the Atchafalaya Basin at that time, indicating that the rapidly shallowing river gradients had resulted in a regime change from braided to anastomosing or meandering.

Pine Island Beach Trend

At this point in the discussions, the writer departs briefly from the heretofore basically chronological presentation of events to avoid potential confusion. With a focus on southeastern Louisiana, the discussions jump ahead by as much as 7,000 years, following to a logical conclusion a consideration of landforms and deposits directly related to the final stages of the Holocene transgression. Readers should be aware that in central Louisiana, Mississippi River delta lobes were forming during this 7,000-year period: these are discussed later in the section dealing exclusively with delta complexes and the evolution of the deltaic plain. This break in continuity of discussion is possible and logical because the portion of southeastern Louisiana under consideration was not affected by deltaic sedimentation until after 5,000 years B.P.

Between about 12,000 and 6,000 years B.P., rising sea level caused the Gulf shoreline to continue to retreat inland in southeastern Louisiana at a moderate to rapid rate. It is likely that the shoreline was so unstable that a beach of significant size did not have an opportunity to form. Various workers have hypothesized that there were significant still stands during that period, such as a stand at a depth of about 50 ft below present sea level between about 9,000 and 8,000 years B.P. (Penland, Boyd, and Suter 1988), but no erosional or depositional evidence of this stand has been discerned in southeastern Louisiana. However, at about 7,000 years B.P. when sea level was about 30 ft lower than at present, Gulf waters first transgressed across interfluves on the Pleistocene surface in the New Orleans area (Miller 1983). About 6,000 years B.P. when sea level had risen to within 10 to 15 ft of its present level, a decline in the rate of rise coincided with a relatively steeply sloping area on the Pleistocene surface to cause an appreciable slowing of the rate of shoreline retreat. Since there was an unusually large amount of sand present in the area from longshore drift from the mouth of the Pearl River, a large, linear sand shoal formed and soon emerged into a beach ridge along a northeast-southwest trend (Saucier 1963). Within a few hundred years, the ridge developed accretion ridges, indicating at least a temporary cessation of shoreline retreat, and started prograding southwestward. By about 5,000 years B.P., sea level had risen at least several feet higher and Gulf waters formed a lagoon or sound north of the beach trend, converting it into a true barrier spit that was anchored to the mainland near the present The Rigolets (Plate 12). Because of a rapid influx of Mississippi River deltaic sediments shortly after that time, the beach trend became isolated from the Gulf, segmented by tidal passes, and eventually surrounded and buried. As a result of this fortuitous circumstance, the trend has become a part of the sedimentary record and is called the Pine Island Beach Trend. Thousands of borings were recently used (Saucier et al. 1984, Saucier 1991b) to update and refine previous mapping (Corbeille 1962, Saucier 1963) of the configuration of this marine feature: a generalized version is shown in Figures 49A and 49B.

Virtually all of the trend now lies beneath sea level because of subsidence, although 5,000 years ago much of it must have been 5 to 10 ft above sea level and closely resembled the barrier islands now located off the coast of Mississippi and Alabama (Plate 12). At one time, the feature probably stretching for more than 35 mi without a break. It must have been extensively used and probably occupied during the Late Archaic period at least seasonally by prehistoric populations, but no artifacts or evidence of human habitation have been detected (or at least reported) in any of the thousands of borings or several borrow pits that have penetrated its surface. Numerous shell middens of the Tchula period (Tchefuncte culture) (Figure 2) occur along the beach trend (Ford and Quimby 1945, Shenkel 1974), but those were occupied after the trend was isolated by interdistributary marshes or are associated with later, small lake beaches which formed along the south shore of Lake Pontchartrain.

At one time, the problematic Milton's Island Beach Trend, which occurs along the north shore of Lake Pontchartrain (Figure 47), was thought to be contemporaneous with the Pine Island Trend (Saucier 1963). However, as discussed earlier, a slightly more viable argument can be made in favor of the former trend being a Middle Wisconsin Stage feature.

As a Holocene-age, relict barrier island or barrier spit, the Pine Island Beach Trend can be considered a unique feature in Louisiana: there are no known or probable chronostratigraphic equivalents. Its formation was due to a fortuitous set of circumstances in a favorable setting. In southwestern Louisiana, there was neither a topographic rise on the Pleistocene surface to slow the rate of shoreline retreat nor an abundant supply of sand to nourish a beach trend. Conditions as they existed elsewhere in the deltaic plain area about 5,000 years B.P. are discussed later in this chapter.

Mississippi River Meander Belts

The writer has found this to be the most difficult section to write of the entire volume. To reiterate what was mentioned in Chapter 1, it is feared that

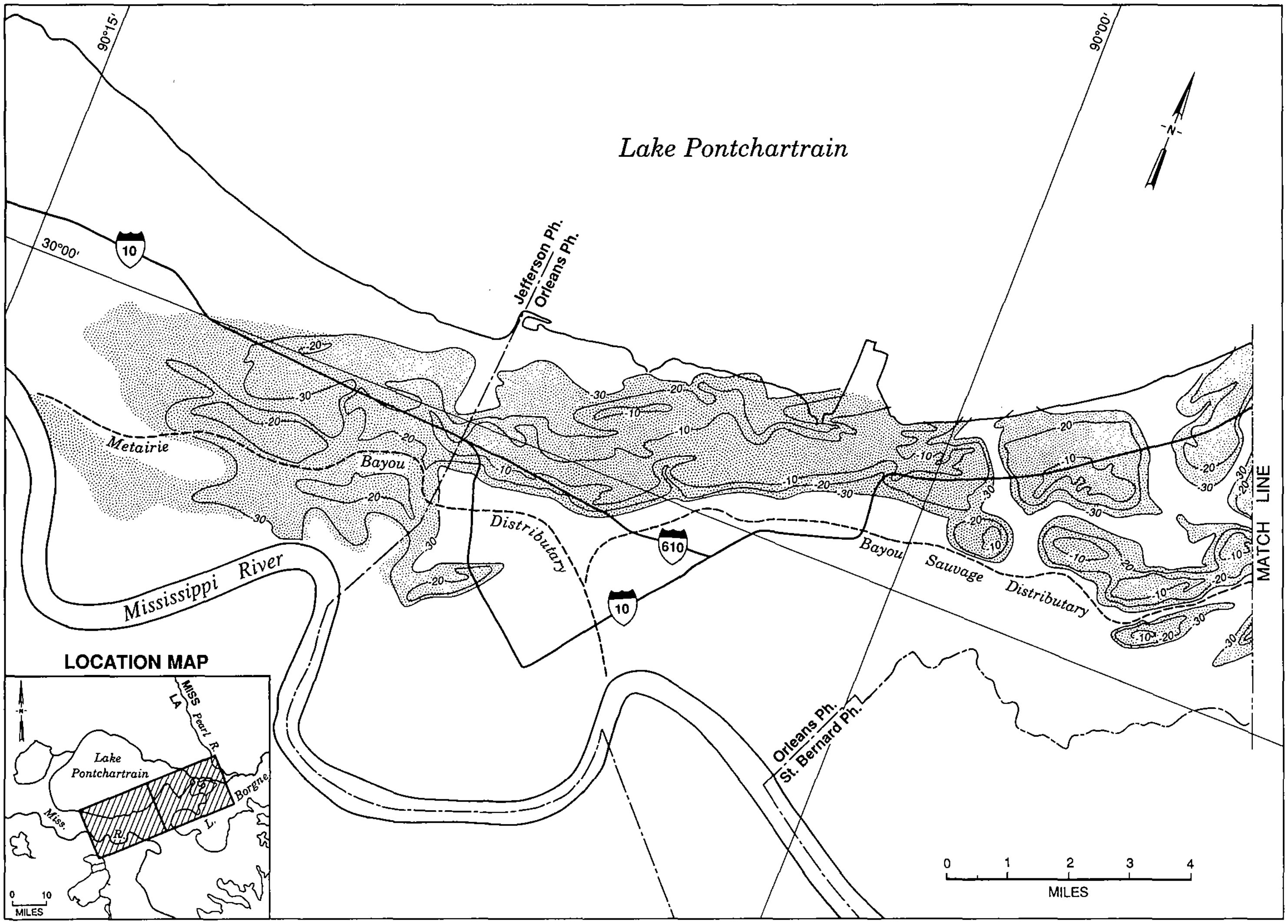

Figure 49A. Configuration of the top of the western half of the Holocene-age Pine Island Beach Trend buried beneath deltaic deposits in the greater New Orleans area

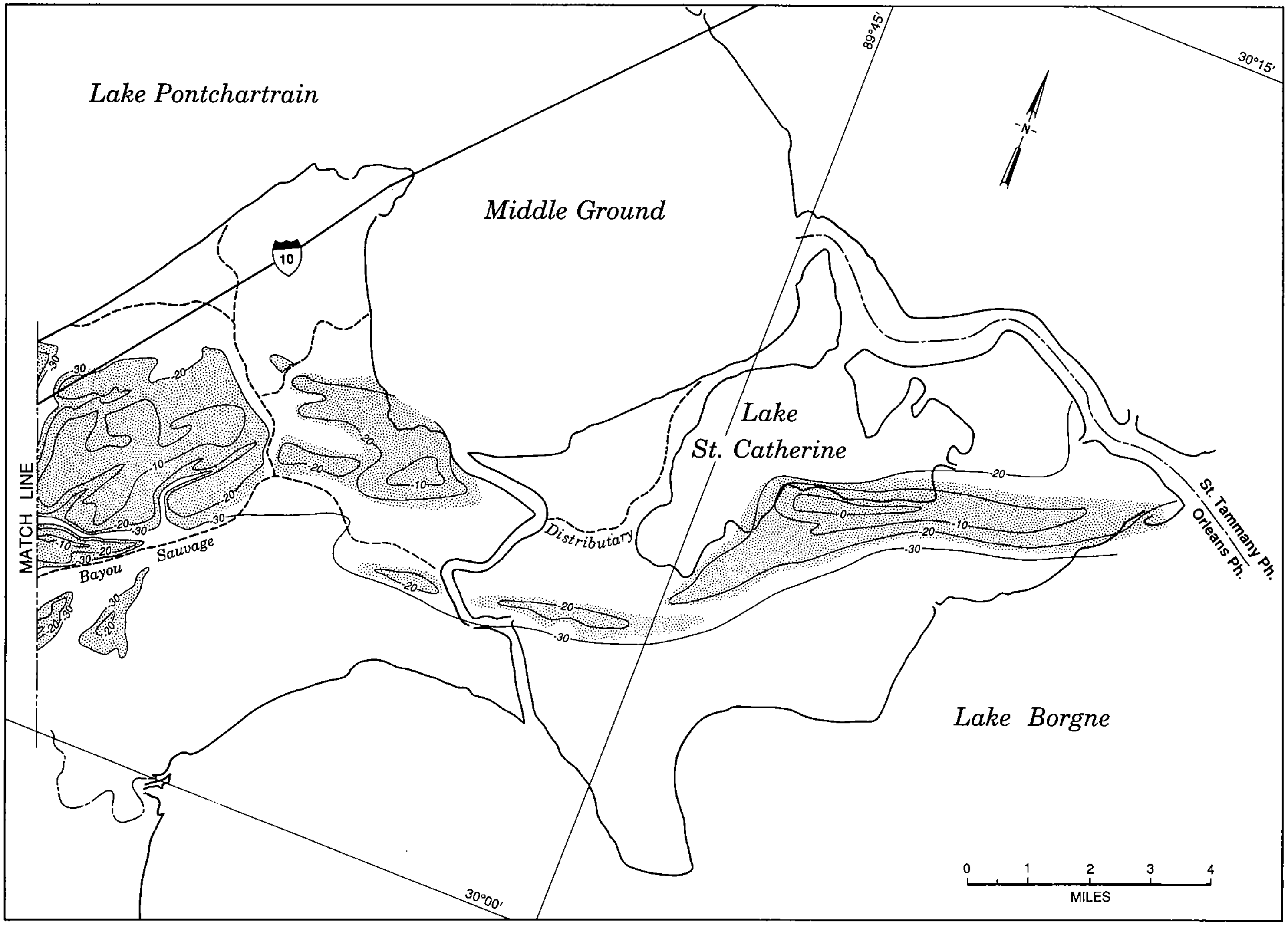

Figure 49B. Configuration of the top of the eastern half of the Holocene-age Pine Island Beach Trend buried beneath deltaic deposits in the greater New Orleans area many readers will be strongly disappointed that so little can be stated with certainty about the chronology of Holocene Mississippi River meander belts. After the detailed reconstructions of meander belt and abandoned channel ages by Fisk (1944), it is difficult if not even embarrassing for geologists to admit that during the past 50 years, we have taken *apparent* major steps backward rather than forward. The logical high expectations for even greater advances and more detail as a result of 50 years of more work cannot be met. Once again, it is essential to mention that there is currently *no* definitive evidence to establish even the relative ages of whole meander belts (or distributaries) much less the approximate ages of individual cutoffs. For that reason, no attempt has been made in the paleogeographic reconstructions (Plates 28H to 28M) to show the positions of abandoned courses within the meander belts for the various time periods.

Since the reasons for the current poor state of the art have already been elaborated, it should be pointed out that the interpretations presented below, which may ultimately prove to be no more correct than those of Fisk, do involve factors that were not previously considered. For example, 50 years ago, the chronological reconstructions were influenced by a knowledge of meander belt processes (such as cutoff formation), but not by why, when, and where whole meander belts formed. While meander belt relationships are equivocal, thought *has* been given for the first time to the probable influences of preexisting topography and local drainage. Readers are cautioned that the reconstructions are also heavily biased by untested concepts of this writer as to why certain meander belts are apparently less than full-flow channels. If these concepts are valid, important advances in our knowledge of processes have emerged that hopefully will more than compensate for the lack of progress in chronological reconstructions.

The focus of attention in this chapter of the synthesis necessarily is on the Yazoo Basin, for it is there that there is the greatest spatial separation of meander belts. Interpretations of that area depart significantly from previous ones (e.g., Saucier 1981, Saucier and Snead 1989) in certain key aspects and include concepts and information not previously published. For example, a sixth meander belt is now recognized rather than five and only three carried the full flow of the Mississippi River and only two of those for any significant period of time, the others being only partial-flow distributaries.

Between the mouth of Thebes Gap and a point about 35 mi north of Memphis (Plates 4 to 6), the Mississippi River has occupied one meander belt continuously since about 10,000 years B.P. In only one small area west of Caruthersville, Missouri (Plate 5), is there an indication of a segment of an abandoned course. Without any chronologic indicators, it is not possible to correlate this segment with any of the separate meander belts farther south in the alluvial valley. The size and configuration of the abandoned course segment suggest that it was no more than a distributary channel that carried only a portion of the total discharge of the river.

Stage 6

No particular significance has been given in previous meander belt chronological schemes to a 18-mi-long, north-south trending band of point bar deposits with several related abandoned channels lying near the eastern side of the alluvial valley due east of Helena (Plate 7). The small town of Marks, Mississippi, is situated on the meander belt, and portions of it are occupied by the Coldwater River. Because of the small size of the channels, it was assumed that they must be related to some unknown small stream and they were designated as undifferentiated alluvium (Saucier and Snead 1989). However, it is presently believed that no local drainage could have carried a sufficient sediment load to create features of that size (meander radii and channel width); consequently, they must have formed along a major Mississippi River distributary channel. Accordingly, the meander belt is designated as the Stage 6 channel position (Figure 25A) and has been given the name South Lake Meander Belt (M6, Plate 1).

If this proposed mode of origin is correct, two key questions arise. How old is the meander belt? Since it obviously did not carry the full discharge of the river, where was the main river channel at the time? The relative stratigraphic relationships in the area indicate only that the meander belt *could* be the oldest in the area, at least in the Yazoo Basin, but not that it necessarily is. No radiometric dates are available, and although it is highly likely that archaeological sites are present, their ages are not known.

In the tenuous scenario that is offered herein, the main channel and meander belt of the Mississippi River was located about 10 mi to the west and trending north-south (Plate 28H). It was flowing in essentially the same location as the river was when it rather suddenly changed from a braided to a meandering regime about 10,000 years B.P. when glacial runoff abruptly ceased. It is hypothesized that the South Lake Meander Belt formed as a crevasse off the main channel sometime between about 10,000 and 9,000 years B.P. and persisted for only a few hundred years (Figure 50). Perhaps the distributary developed to the extent that it did only because floodwaters entered and found an easy path to follow along local drainage located in a flood basin environment between the river and the uplands.

It is not possible to determine where the meander belt began or how far downvalley it flowed, but it is likely that it reentered the main river channel as a tributary within a few tens of miles. No chronostratigraphic equivalent to this Stage 6 meander belt has been identified elsewhere in the alluvial valley area: if one exists, it may be buried beneath backswamp deposits.

In terms of landscapes, when the Stage 6 channel was active, the Mississippi River was flowing in a meandering regime throughout most of the alluvial valley area. The climate was comparable to that of today, all valley train areas were probably heavily vegetated, boreal species had disappeared from the alluvial valley area, loess was no longer being deposited, and backswamp areas vegetated with cypress-gum forests were becoming widespread. Natural levee

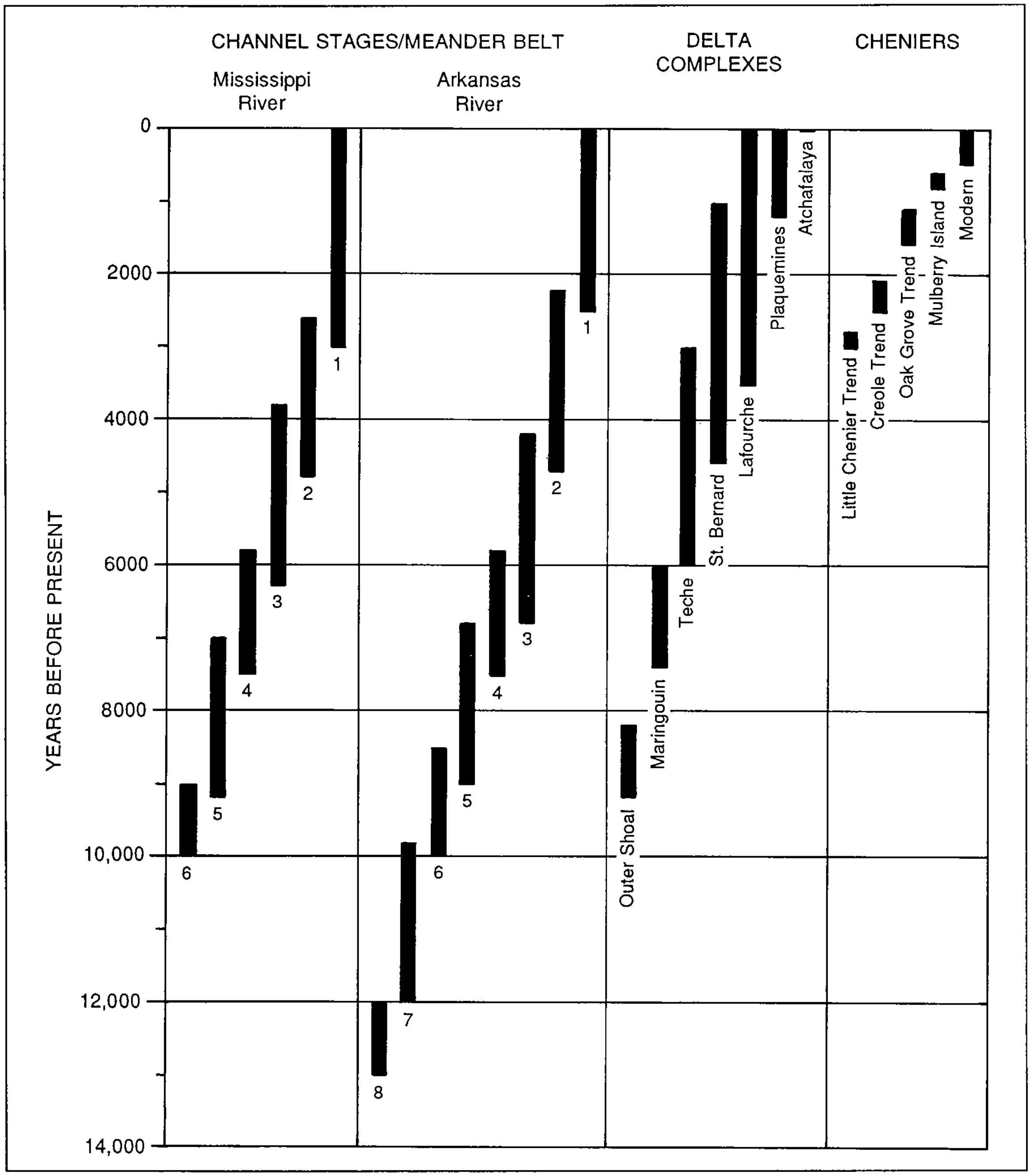

Figure 50. Estimates of ages of channel stages (meander belts) of major rivers, delta complexes, and cheniers (modified from Autin et al. 1991)

ridges were starting to become significant landscape features but were probably not important elements to the prehistoric inhabitants. Their economy apparently still centered around the hunting of large Pleistocene megafauna, and they focused their nomadic activities on the valley train surfaces where a savannah-like environment (upland grasslands with riverine, deciduous hardwood forests) may have existed rather than a riparian setting.

Stage 5

Evidence of the Stage 5 channel occurs in several areas along the eastern side of the Yazoo Basin from about 20 mi south of Memphis to about 30 mi south of Greenwood (Figure 25A). The principal components are what has been designated herein as the Coldwater Meander Belt (M5, Plate 1) and several isolated occurrences of abandoned channels farther to the south (Plates 7 and 8). The delineation and designation of this meander belt are the same as has been used by this writer in recent publications (Saucier 1981, Saucier and Snead 1989), but different from that used in Saucier (1974) and earlier works.

The Stage 5 channel may actually represent where the main channel of the Mississippi River was located during the preceding Stage 6, but it is just as likely that the river adopted it as a new course through a diversion from somewhere to the west (Plate 28I). Between Helena and Greenwood, the Stage 5 channel may have occupied a portion of the Stage 6 meander belt; however, between Greenwood and Vicksburg it adopted an entirely new channel. Based on the large size of the abandoned channels which are closest to the margins of the zone of channel migration (i.e., the zone of point bar accretion) and which are presumably among the oldest, full-flow conditions were attained rather quickly after the meander belt was first occupied. Morphometrically, there are three well-preserved abandoned channels along the eastern side of the meander belt north of Greenwood (Plate 8) which are comparable to those along the present Mississippi River meander belt.

No direct evidence of any type exists; however, the positioning of Stage 5 within the overall chronologic scheme suggests that the full-flow conditions occurred for a few hundred years sometime between about 9,000 and 8,000 years B.P. (Figure 50). Two archeological sites with Middle Archaic-period components are known to be associated with the Stage 5 meander belt about 15 mi southeast of Clarksdale, Mississippi (Plate 7) (Connaway 1981). However, these sites, which date to perhaps as long ago as 7,000 years B.P. (Figure 2), are associated with a small stream (Opossum Bayou) that formed after the meander belt was abandoned. Therefore, the sites establish only a minimum age for the feature.

One concerted effort has been made to systematically obtain radiocarbon dates on abandoned channels and large swales in the Stage 5 (and other) meander belts along the eastern side of the Yazoo Basin (Thorne and Curry 1983). A total of 12 dates were obtained from soil cores at five locations in Stage 5 features. Of the total, four were considered to be anomalously old (e.g., more than 16,000 years B.P.) because of inferred contamination with lignite and coal particles (see Chapter 1). The remainder of the dates vary from 120 to 4,890 years B.P., with four being younger than 1,000 years B.P. The latter are valid dates, but they obviously date very recent organic sedimentation in the features, some of which are accumulating materials today. The date of 4,890 years B.P. establishes only a minimum age for the meander belts and is not believed to be diagnostic of the period of active discharge.

After the episode of full river discharge, there was apparent progressive diversion of flow to another meander belt because younger abandoned channels and segments of abandoned courses are distinctively smaller in size. A statistical analysis of the smaller cutoffs verifies they represent conditions of divided flow, but an actual quantitative estimate of paleodischarge volumes has not been possible (Smith 1989).

The smaller size of the later meander belt features is most apparent in the Coldwater Meander Belt segment (Plate 7). A well-developed abandoned course can be traced continuously over an airline distance of about 40 mi and probably represents flow conditions as they existed just prior to total abandonment. This writer estimates that river discharge at that time was not over 50 percent of the total.

South of the Yazoo Basin area, there is no indication of the location of the Mississippi River during Stage 5. The absence of direct evidence suggests that it was in the same or approximately the same location as the later Stage 3 meander belt, but the evidence has been destroyed by subsequent meandering. Since sea level was several tens of feet lower than at present at about 8,000 years B.P., natural levees and other features associated with the channel would have formed considerably below present floodplain level south of the mouth of the Red River. It is probable that the meander belt had few or no cutoffs and perhaps was even transitional between braiding and meandering (anastomosing?) due to the rapidly rising base level.[1] In backswamp areas, sedimentation rates were probably relatively high and deep swamp to lacustrine conditions prevailed.[2]

Stage 4

The paleogeography of the alluvial valley is especially confusing and uncertain during Stage 4. The only aspect that seems reasonably certain is that the Mississippi River flowed in a single meander belt to as far south as Helena. A segment designated the Big Creek Meander Belt (M4, Plate 1) is preserved in the lower St. Francis Basin area (Plate 6) only because the river later diverted

[1] Personal Communication, 1994, T. E. Tornqvist, Geographer, Utrecht University, Utrecht, The Netherlands.

[2] Personal Communication, 1994, Andres Aslan, Geologist, University of Colorado, Boulder, CO.

to other courses in that area: otherwise, there is no manifestation of the specific location or character of the river at that time.

Probably somewhere very close to Helena, the river began about 7,500 years B.P. to divert a portion of its flow from the Stage 5 channel southward through the approximate center of the Yazoo Basin along a route immediately west of the inferred position of the main Stage 6 meander belt (Plate 28J). The new distributary enlarged over a period of probably a few hundred years, creating what has been designated the Little Mound Bayou Meander Belt (M4C, Plate 1) in the Mound Bayou-Shelby, Mississippi, area (Plates 7 and 8). Based on abandoned course segments and abandoned channels in that area, and in other areas near Indianola, Mississippi (Plate 8), the meander belt never carried more than perhaps about half of the total river discharge.

South of the vicinity of Indianola, a Stage 4 meander belt has not been identified in the Yazoo Basin, but it is believed that it is present but so heavily veneered with backswamp deposits as to be indistinguishable at the surface. No borings exist in this area to ascertain its presence in the subsurface, however.

A probable downstream equivalent of this Stage 4 channel is the well-preserved Tensas Meander Belt (M4D, Plate 1) in the upper end of the Tensas Basin west of Vicksburg. The morphometry of the abandoned channels and abandoned course in that meander belt (Plate 9) is highly similar to that of the Little Mound Bayou Meander Belt, once again indicating a major distributary status rather than a full-flow channel.

While the Little Mound Bayou-Tensas Meander Belt system was carrying perhaps half the flow of the Mississippi River, the remainder was still being conveyed by the Stage 5 channel. However, before that flow ceased, between Minter City, Mississippi, on the north and Belzoni, Mississippi, to the south (Plate 8), the river created a new and distinctively separate distributary channel just to the west. The result is what is designated as the Bear Creek Meander Belt (M4B, Plate 1). Thus, for an appreciable period of time centered about 7,500 years B.P., the discharge of the river was more or less equally divided between two channels over an airline distance of about 150 mi.

The two channels apparently merged a short distance north of Natchez, and from there southward the full-flow channel shifted to the western side of the alluvial valley to near Marksville, Louisiana (Plate 10), and thence followed the trend of the later Teche Meander Belt (M3D, Plate 1). No discrete meander belt marks the course of the river at that time, but it apparently evolved in place into the Stage 3 meander belt which can be followed at scattered locations north of Marksville and from there southward becomes the Teche Meander Belt *per se*. Until several years ago, no separate Stage 4 meander belt was believed to exist south of Marksville; however, detailed mapping in the area (Smith and Russ 1974) revealed an abandoned course and meander belt features which were interpreted as being separate from and chronostratigraphically older than the Teche Meander Belt (Saucier and Snead 1989). In this synthesis, the writer identifies them as having been formed during Stage 4 (Plate 11) and designates them as the Bayou Portage Meander Belt (M4E, Plate 1). Readers are cautioned, however, that the meander belt more often than not is indistinguishable from the Teche ridge in terms of being a distinctively separate physiographic entity.

South of a point about midway between Opelousas and Lafayette (Plate 11), the pattern of abandoned channels and meander scroll topography is so complex that it is often not feasible to assign a feature to a particular meander belt (i.e., either Stage 4 or Stage 3). Moreover, based on work by Saxton (1986) and further interpretations by Goodwin et al. (1991), it is possible there are features other than those discussed herein in a trend west of Bayou Teche that correlate with the Stage 4 channel. In the latter reference, those features are described and designated the Lake La Pointe Meander Belt. This writer believes that explicit designation of each feature in that area on Plate 11 would be premature and is avoided.

Archeological evidence bearing on the age of the Stage 4 channels is clouded with uncertainty. There are several late Paleo-Indian or Early Archaic-period sites with Dalton culture artifacts that occur in the central Yazoo Basin in close proximity to Stage 4 meander belt segments (Brain 1970), but in no case has the provenance been established with certainty. Some of the sites suggest association with Stage 4 natural levees, but there is reason to believe the sites are actually located (initially occupied) on Late Wisconsin Stage valley train deposits and are partially surrounded by later natural levee deposits. As indicated earlier, however, there were probably sociopolitical and/or economic reasons why Archaic and earlier cultures may not have favored meander belts for permanent habitation. Consequently, archeological evidence (including the lack thereof) may be of considerably less value in landform dating than in later (Formative Stage) cultures.

Several sites dating to the Poverty Point period are located on the Bear Creek Meander Belt, including the famous Jaketown Site near Belzoni (Plate 8) (Ford, Phillips, and Haag 1955). Radiocarbon dates from this site establish its occupation between about 2,800 and 2,100 years B.P. (Ford and Webb 1956) which is well after the estimated time of Mississippi River flow through the meander belt. This situation is verified by the site stratigraphy which shows cultural deposits overlying a considerable sequence of channel filling deposited by the Yazoo River after the system was abandoned by the Mississippi River. The association of Poverty Point period sites with underfit, slack-water streams within the confines of larger abandoned channels is a characteristic pattern throughout much of the Yazoo Basin area.

Radiocarbon assays on samples from Stage 4 meander belt features during the Thorne and Curry (1983) investigation provided six valid dates and seven that were interpreted as anomalous for various reasons, the former ranging in age from approximately 7,700 to 520 years B.P. The oldest of those dates, obtained from the Sky Lake cutoff about 7 mi north of Belzoni (Plate 8), is about 500 to 1,000 years earlier than what has been estimated by this writer. When the precision of the types of evidence used is considered, there is no legitimate basis on which to question its validity.

Prior to the formulating of the concept that the Stage 4 meander belts in the Yazoo Basin area represented divided riverflow (Saucier 1974), extensive consideration was given to the possibility that the stage involved only a single channel that was responding to a climatically induced reduction in river discharge (see Chapter 3). Supporting that possibility of an Altithermal response was a pollen profile from the Old Field site in the Morehouse Lowland area which indicated a decline in water levels and an expansion of grassland between about 8,700 and 6,500 years B.P. (King and Allen 1977). Those dates coincide well with the estimated age of the Stage 4 channels (Figure 50), but they do not agree with the generally accepted age range for the Altithermal (7,500 to 4,000 years B.P.) when considered from a regional rather than local perspective. The time-span during which the Mississippi River drainage system would have most likely responded to a reduction in runoff coincides better with Stage 3, which unquestionably indicates a discharge regime similar to the present.

Stage 3

Termination of flow in the two separate Stage 4 meander belt trends probably occurred simultaneously because of an event in the southern St. Francis Basin area. Probably starting as a major crevasse near the Wilson-Joiner area in east-central Arkansas (Plate 6), the Mississippi River diverted increasing flow into a narrow lowland between Crowley's Ridge and the natural levee ridge of the Stage 4 channel until the distributary eventually captured the full flow of the river (Plate 28K). The resultant meander belt has been designated the St. Francis Meander Belt (M3A, Plate 1) and is clearly evident by its abandoned course and numerous abandoned channels (Plate 6).

From that area southward, the river reoccupied a short stretch of the existing meander belt past Helena into the northern end of the Yazoo Basin before once again diverting to a new course. The new course developed south of the Clarksdale area through the narrow lowland or basin between the Stage 4 meander belt ridges, creating the clearly evident Sunflower Meander Belt (M3B, Plate 1; Plates 7 to 9). That feature forms a prominent, 95-mi-long (airline distance), broad, alluvial ridge to as far south as about 25 mi northeast of Vicksburg.

When the sizes and shapes of the abandoned channels along the Sunflower Meander Belt are considered in overview, there are relatively few that indicate less than full-flow conditions. This indicates that both the processes of meander belt formation and abandonment were relatively rapid and that no significant amount of flow was diverted through a distributary at any time.

In the area between Natchez and the Red River near Marksville, the Mississippi River Stage 3 channel occupied the Cocodrie Meander Belt (M3C, Plate 1; Plates 10). With consideration given to the possible obscuring effect of later backswamp deposits that have encroached upon the flanks of the meander belt, it is evident that the river formed very few cutoffs in that area. The morphology of the abandoned course indicates that it carried less than the full flow of the river. Further, the Cocodrie Meander Belt is only one-third to one-half the width of the Sunflower Meander Belt. A possible explanation for the limited amount of meandering was the presence of thick and relatively erosion-resistant backswamp deposits flanking the meander belt in that portion of the Tensas Basin, but the suggested partial flow implies that other channels were active. If the channels were active, they evidently did not develop meander belts and may have been part of a complex anastomosing-channel network.

South of the Red River, the Teche Meander Belt marks the continuation of the Stage 3 channel into the deltaic plain. As that system proceeded southward, it (and its underlying Stage 5 and 4 predecessors) encountered a thick and continuous backswamp sequence that significantly inhibited channel meandering and cutoff formation (Plates 11 and 13).

In an attempt to establish the chronological position of the Stage 3 channel (Figure 50), this is one of the few instances in the alluvial valley area in which geological evidence apparently is more diagnostic than archeological evidence. Austin (1986) recently obtained eight radiocarbon dates (two of which she interpreted as anomalous) from a sediment core in abandoned channel deposits in Larto Lake, an oxbow lake located west of Black River south of Jonesville, Louisiana (Plate 10). The oldest date, 5,870 years B.P., from near the base of the sedimentary sequence indicates that channel cutoff took place shortly before that time. For unknown reasons, it appears that this dating attempt did not incur difficulties with lignite and coal contamination of the type which proved so troublesome to Thorne and Curry (1983) in their Yazoo Basin study.

It should be noted that in previous meander belt correlations (Saucier and Snead 1989), Larto Lake has been included as part of Stage 2. However, when the radiocarbon dates and the physiographic evidence that is nondefinitive and allows multiple interpretations were considered, that feature was redesignated as related to the Stage 3 channel.

Numerous archeological sites are associated with the Stage 3 channel, especially along the Sunflower Meander Belt. The oldest appear to include at least 15 sites of the Poverty Point period (Webb 1977), the age of which has been well established in the region as being between about 3,500 and 2,500 years B.P. (Figure 2). All of the sites are believed to have been initially occupied well after the meander belt was abandoned and therefore provide no clue as to the time of its origin. Radiocarbon dates are available from cultural deposits at three other sites situated on Stage 3 meander belts, but all of these are younger than 1,200 years and similarly are geologically undiagnostic.

Stage 2

Mississippi River flow through the Stage 3 channel ended abruptly when a major crevasse formed in the Wilson-Joiner area of the lower St. Francis Basin (Plate 6) and the river created a new meander belt along the eastern edge of the alluvial valley past Memphis (Plate 28L). Some of the abandoned channels in the present (Stage 1) meander belt between the point of diversion and the vicinity of Helena may actually date to Stage 2, but in an absence of any chronological indicators, all such features are designated as part of Stage 1. The only meander belt segment that can be reasonably correlated with Stage 2 is the Fifteen Mile Bayou Meander Belt (M2A, Plate 1) which impinged against Crowley's Ridge in the vicinity of Marianna, Arkansas (Plate 6). The size of the features in this segment indicates it carried the full flow of the river with no major distributaries functioning upstream of this point.

Just downstream from that meander belt segment, probably very close to Helena, the river apparently divided its flow between two channels quite early in Stage 2. This was not a case of the river occupying one channel and then creating and eventually shifting its discharge to a major distributary. Rather, it appears that two channels formed simultaneously and each carried no less than 40 percent and no more than about 60 percent of the total discharge throughout their active lives.

The eastern channel is clearly marked by the distinctive Yazoo Meander Belt (M2B, Plate 1) which can be traced without interruption for about 150 mi from just south of Helena past Greenwood to near Vicksburg (Plates 7 to 9). Although the location of that meander belt is largely coincident with the earlier Stage 5 channel, the two can be differentiated in most cases on the basis of the size of meander belt features and the fact that the younger Stage 2 abandoned course is continuous while that of Stage 5 is quite discontinuous. In the vicinity of Holly Bluff, Mississippi (Plate 9), the Stage 2 channel reentered the Sunflower Meander Belt and flowed to near Vicksburg.

In contrast to the meander belt of the eastern channel which is so evident, the location and size of the western meander belt has been difficult and uncertain. No separate meander belt (as a distinctive physiographic feature) is present, and evidence of its existence lies solely with a relatively small number of abandoned channels in a cluster located west of Cleveland, Mississippi (Plates 7 and 8). Further, there is nothing about the abandoned channels other than their relatively small size that indicates their association with Stage 2. They have been identified as Stage 2 only because the presence of a less than full-flow meander belt has been theorized as being present somewhere in this area. Simply put, a channel complementary to the Yazoo Meander Belt must be present somewhere, and by process of elimination, there is no other place that it can be. Hence, this is a very tenuous identification.

Occasional abandoned channels of relatively small size occur at scattered locations along both flanks of the modern (Stage 1) meander belt south of the Cleveland area. While these have not been identified as Stage 2 features, this writer would not be surprised if future work produced radiocarbon dates or archeological evidence (e.g., Late Archaic period sites) that verify the cutoffs are old enough to warrant such correlation.

Just northwest of Vicksburg in the vicinity of Eagle Lake, Mississippi (Plate 9), the eastern and western channels merged, and from there southward a full-flow channel created the Walnut Bayou Meander Belt (M2C, Plate 1). Although the number of cutoffs that took place in that meander belt are relatively small, the river meandered widely and there can be no question that it carried the full discharge of the river for at least a thousand years. In some areas, such as near Tallulah, Louisiana (Plate 9), the Walnut Bayou Meander Belt simply was a continuation of occupation of the Stage 3 channel. Over most of the distance of about 95 mi to the Red River, however, that meander belt developed in a slightly different location than the Stage 3 Cocodrie Meander Belt, actually cutting across it in one instance (Plate 10).

A very significant event occurred just south of the Red River early in the development of the Walnut Bayou Meander Belt. For the first time during the Holocene, the river created a distributary and quickly diverted its full flow toward the eastern side of the alluvial valley. South of the Old River area (Plates 10 and 11), no separate meander belt marks its course to the Gulf, but subsurface data indicate that one underlies the modern meander belt between Donaldsonville and New Orleans (Plates 11 and 12). That meander belt development marked the first time that any consequential amount of flow was being discharged through a meander belt into the Atchafalaya Basin area. As will be discussed later, that event led to extremely important developments that affected the entire deltaic plain.

There is no readily apparent explanation for why more than 4,000 years passed before the Mississippi River was able to affect a diversion out of the Teche and its predecessor courses along the western side of the alluvial valley to take advantage of an obvious gradient advantage. Possible factors include the inhibiting influence of thick backswamp deposits in the river's bed and banks, a relatively finer grained sediment load than farther upstream, and reduced stage variations along the lower reaches of the river.

Relatively more chronological data are available for the Stage 2 channel (Figure 28) than for any of the other five in the alluvial valley area, most of it being related to the Yazoo Meander Belt. There are 15 radiocarbon dates on cores from abandoned channels (Thorne and Curry 1983) of which six have been interpreted to be valid. Five of those are on organic materials deposited during later stages of channel filling and are not geologically significant. However, one date, approximately 4,000 years B.P., appears to accurately establish the time of cutoff of an abandoned channel located about 6 mi south of Greenwood. The location of the abandoned channel within the meander belt indicates that it formed relatively early in the life of that system. That date, therefore, is interpreted as suggesting that initial flow through the Stage 2 channel probably does not predate that time by more than 500 to 1,000 years.

Archeological evidence from the Yazoo Meander Belt is considerably more abundant and even more indicative of the age of that feature. Nineteen radiocarbon dates are available on cultural deposits from 10 sites, nine of which are from the Teoc Creek Site, a Poverty Point period site located 8 mi northeast of Greenwood (Plate 8). The nine dates are all deemed to be valid and tightly cluster between 3,700 and 3,000 years B.P. This archeological site firmly establishes a minimum age for a Stage 2 abandoned channel which is compatible with the date from the geological context, but it has far greater significance. The Teoc Creek Site provided this writer a rare opportunity to reconstruct in unusual detail the paleogeography of a segment of a meander belt using information from field investigations, soil cores, archeological excavations, and radiocarbon assays (Connaway, McGahey, and Webb 1977). Since that reconstruction provides an indication of the complex patterns of local drainage that can develop when a Mississippi River meander belt is abandoned, and how extreme care must be exercised in interpreting site/landform relationships, a summary follows as a valuable case history and example.

The Teoc Creek Site is unusual in the Lower Mississippi Valley because it contains sound evidence that human occupation took place during a time of active natural levee formation. In the vast majority of cases, sites are initially occupied and inhabited where and when overbank flooding and active sedimentation has ceased. As explained later, the Teoc Creek Site was initially occupied on an inactive natural levee, but habitation, albeit reduced, continued uninterrupted through a subsequent period of levee formation.

The preoccupation setting of the Teoc Creek Site vicinity is shown in Figure 51A. The eastern Stage 2 channel had meandered to within 2 mi of the eastern edge of the alluvial valley and in close proximity to an abandoned channel dating to Stage 5. Sometime prior to 4,000 years B.P., a neck cutoff took place on the Stage 2 channel, creating a narrow, linear oxbow lake. Shortly thereafter, Teoc Creek, a small upland stream, began flowing through the lower arm of the abandoned channel, largely filling that portion of the oxbow lake with sediment (Figure 51B). Shortly before 3,700 years B.P., Indians first occupied the Teoc Creek Site, settling on the crest of an inactive natural levee along the banks of the abandoned channel where it was still an oxbow lake.

Within several hundred years (at most), the Yalobusha and Tallahatchie rivers, which had been flowing in a combined channel through the Stage 5 abandoned channel to the north, diverted southward and occupied a course past the Teoc Creek Site (Figure 51C). Those streams began building natural levees within the confines of the larger abandoned channel, but this presumed slight deterioration in living conditions apparently did not preclude occupation of the site. The accompanying reduction in the size of the oxbow lake may have been a factor instrumental in the initiation of habitation at the Neill Site in a slightly more attractive setting about a mile to the west, however.

A short time after initial occupation of the Neill Site, the Yalobusha-Tallahatchie River changed direction within the Stage 2 abandoned channel

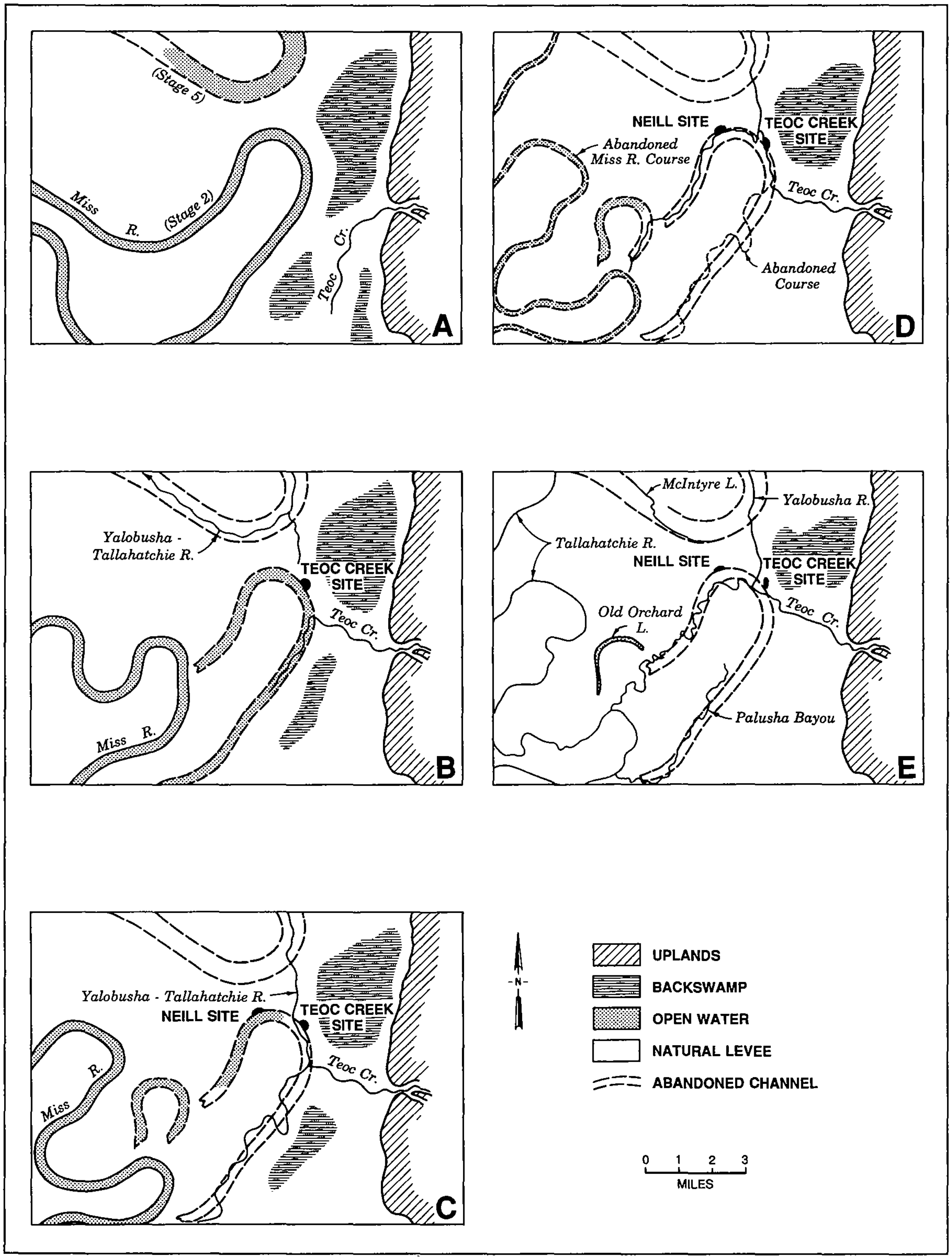

Figure 51. Evolution of a part of the Yazoo Meander Belt (Stage 2) as reconstructed using geomorphological and archeological evidence from the Teoc Creek and Neill Sites (modified from Connaway, McGahey, and Webb 1977)

(Figure 51D). This event marked the beginning of a period of active natural levee growth at that site but once again did not preclude continued occupation. The radiocarbon chronology suggests that although occupation continued at the Neill Site, the Teoc Creek Site was abandoned at or slightly after the river shift which is estimated at about 3,000 years B.P. Since the river shift could have occurred only after the Stage 2 channel was abandoned by the Mississippi River, this date establishes a minimum age for the Yazoo Meander Belt.

The two final events in this scenario were a diversion of the Tallahatchie River to a new channel to the west, leaving only the Yalobusha River flowing through the Stage 2 abandoned channel, and a channel shift in Teoc Creek which resulted in the destruction of a significant part of the Teoc Creek Site (Figure 51E). Those events, probably occurring prior to about 2,000 years B.P., resulted in what is essentially the modern setting.

Thus, the total package of chronological evidence supports the interpretation that the Yazoo Meander Belt began forming about 4,800 years B.P. and was abandoned by the Mississippi River about 3,000 years B.P. or slightly earlier. No radiocarbon dates or archeological evidence is available from the western channel in the Yazoo Basin area, and similar data from the Walnut Bayou Meander Belt are all too young to be indicative of the period of active sedimentation in that system.

Stage 1

When the Mississippi River abandoned the eastern Stage 2 channel in the Yazoo Basin area, full flow was shifted to the western channel. In essence, that event marked the beginning of the Stage 1 channel and initiation of the modern meander belt: no upstream avulsion took place to create a channel in a new location. North of Helena, the Fifteen Mile Bayou Meander Belt was abandoned in favor of a route slightly farther to the east (Plate 7) but did not necessarily involve an avulsion. Rather, it was probably only a complex neck or chute cutoff that resulted in the abandonment of a series of bends rather than a single one.

Since attaining a full-flow status between Memphis and the vicinity of Vicksburg about 3,000 years ago (Figure 28), the Stage 1 meander belt has developed in the unusual manner discussed in Chapter 5; i.e., creating an exceptionally large number of cutoffs considering the amount of time involved (see Figure 25B). The relatively great extent to which the river has meandered in that stretch is likely due to the prevalence of coarse-grained glacial outwash deposits in the river's bed and banks (as compared to backswamp), as well as the probability that the river is following the former route of the White River system which carried a considerable volume of sandy deposits.

The unusual width of the modern meander belt between not only Memphis and Vicksburg but also Cairo and Memphis, and the large number of abandoned channels, many of which contain or used to contain oxbow lakes, were important factors in influencing prehistoric settlement patterns. As documented in Phillips, Ford, and Griffin (1951) and Phillips (1970) and by numerous other archeologists, those abandoned channels are known to contain hundreds of archeological sites, many of which are large Mississippian-period villages or towns with clusters of earth mounds. It is quite apparent that during the Formative Stage (Figure 2), the broad natural levee ridges of the Mississippi River were favored locations for the practice of maize-based agriculture. Attempts to use that extensive database in establishing a chronology of abandoned channels have been largely unsuccessful, primarily because of the lingering influences of Fisk. However, with the use of a different paradigm, at least a "first cut" is technically feasible for small areas and awaits only a comprehensive effort. In addition to potentially hundreds of dates that can be inferred from artifact assemblages, this writer is aware of 60 valid radiocarbon dates from 23 archeological sites that span the period from 2,000 to 300 years B.P.

The most significant event in the development of the Stage 1 channel occurred north of Vicksburg in the vicinity of Eagle Lake about 2,800 years B.P. (as determined from events in the deltaic plain to be discussed later). It is hypothesized that a crevasse along the eastern bank of the river rapidly developed into a major distributary (because of a gradient advantage?) that flowed in the rim swamp along the base of the bluffs past Vicksburg and Natchez to as far south as the Old River area (Plates 9 and 10). Within a few hundred years (at most), the distributary captured the full flow of the river and began developing a meander belt, perhaps facilitated in its rapid expansion by the existence in the swamp area of the "Yazoo-type" channel of combined small valley tributaries such as the Big Black River, Bayou Pierre, and the Homochitto River (Plates 9 and 10). In the vicinity of Old River, flow in the new Stage 1 channel was forced to reenter the Stage 2 channel and proceed southward into the deltaic plain.

Little River Lowland/St. Francis Basin

The Mississippi River first adopted a meandering regime between Cairo and Memphis about 10,000 years B.P., and since then it has repeatedly developed major crevasses and formed distributaries along its western bank. The distributaries extended anywhere from a few miles to nearly 100 mi southwestward through the Little River Lowland on the lowest level (Pvl2) of Late Wisconsin glacial outwash (Plates 5 and 6). Small but well-developed meander belts with natural levees and occasional abandoned channels attest that each carried appreciable flow (5 to 10 percent of total discharge?) and suspended sediment for probably hundreds of years. However, for unknown reasons, none ever threatened to cause an avulsion.

Archeological sites, relationships to Mississippi River meander belts, and a study by Guccione, Lafferty, and Cummings (1988) that involved soil coring and radiocarbon dating provide some clues to the ages of the distributaries. The two southeasternmost systems are apparently the oldest. One of them extends along and just west of I-55 in northeastern Arkansas near the community of Keiser, and another lies a short distance farther west and trends through the community of Dyess (Plate 6). Both meander belts are truncated on their upstream ends by some of the earliest Mississippi River abandoned channels and on their downstream ends by the Stage 3 St. Francis Meander Belt. Thus, they must be at least 6,000 years old.

The next oldest distributary system, relatively long but significantly smaller in size than the others, originated near Portageville, Missouri, and is occupied by Little River past the Hornersville, Missouri, community (Plate 5). It then becomes the route of the Right Hand Chute of Little River through the communities of Big Lake and Rivervale, Arkansas (Plate 6). Extrapolations from radiocarbon dates (Guccione, Lafferty and Cummings 1988) suggest that the system developed about 5,400 years B.P. while the St. Francis Meander Belt was the course of the Mississippi River.

The youngest and best developed of the distributaries is occupied by Pemiscot Bayou and the Left Hand Chute of Little River (Plate 6). It originated east of Blytheville, Arkansas, from one of the more recent Mississippi River abandoned channels and thence flowed through the Arkansas towns of Lepanto, Marked Tree, and Parkin. Between the latter two towns, the meander belt is occupied by the St. Francis River. Below Parkin, the meander belt developed within and reworked portions of the St. Francis Meander Belt, indicating that it developed after the latter was abandoned and must be younger than about 4,000 years B.P. Several archeological sites dating to as early as the Tchula period (Figure 2) are situated on the natural levees of the distributary system, indicating that it must be older than 2,500 years B.P.

Because of the physiographic setting of the Little River Lowland, all of the distributaries mentioned previously continued to carry some flow long after the periods of active natural levee growth. Before artificial levee construction, all overflow from floods along the western bank of the Mississippi River between New Madrid, Missouri, and Blytheville entered the broad and extremely flat lowland and had to eventually find its way back into the river via the distributaries and eventually the St. Francis River.

Immediately north of the Right Hand Chute of Little River lie the St. Francis and Big Lake "sunklands." Those well-known paludal/lacustrine features traditionally have been attributed to tectonic deformation accompanying the 1811-1812 New Madrid Earthquake series (Fuller 1912) (see Chapter 8). This writer has cited indirect geological and archeological evidence, indicating a considerably greater antiquity for the sunklands, and attributed their formation to alluvial damming by the Left Hand Chute of Little River of the lower ends of relict braided channels on the higher Late Wisconsin stage valley train (Saucier 1970). Guccione, Lafferty, and Cummings (1988) and King (1978) disagree with that interpretation, pointing out that organic materials from the lacustrine sediments have yielded radiocarbon dates of only 180 years B.P. On the other hand, the Zebree Site (Morse and Morse 1975), a Mississippi-period settlement dating between about 1,300 and 900 years B.P., is situated on the western edge of Big Lake and probably would not have been occupied had it not been for an ecologically-rich swamp and lake environment in the sunkland. A compromise between the two positions apparently is in order because remains of forest stands and historical accounts indicate an extensive swamp environment existed in the areas prior to 1811.

White/Black River Lowland

Within the context of Lower Mississippi Valley geomorphology and Quaternary chronology, the White and Black rivers and their tributaries have been largely overlooked and forgotten. Essentially nothing is known about the Holocene evolution of these systems--they constitute the "new frontier" wherein the potential for paleogeographic reconstruction using geoarcheological and related approaches is unlimited.

Apparently rather soon after all glacial runoff through the Western Lowlands area ended about 12,000 years B.P., the lowland portions of all of the rivers draining the eastern Ozark Plateau (from the St. Francis River on the north to the Little Red River on the south) developed meandering regimes. Plates 6 and 7 reveal that cumulatively they created hundreds of cutoffs and on several occasions abandoned whole meander belts and created new ones. The most conspicuous abandoned meander belt is that of the White River (just west of Cotton Plant, Arkansas) which is now occupied in part by the Cache River between Augusta and Clarendon, Arkansas. Because of the number and sizes of abandoned channels, the older meander belt was active for a relatively longer period of time than the later one and carried a higher discharge. That probably represents the time when the St. Francis River was tributary to the Black River rather than flowing in a separate channel through a gap in Crowley's Ridge as it does today.

There is currently no basis whatsoever for making *any* estimate of when the various meander belt segments were active or abandoned. Archaic-period and early Formative Stage archeological sites are known to occur on the natural levees of the abandoned channels and courses, but no attempt has been made to reconstruct channel sequences based on their times of occupation. This writer is not aware of any radiocarbon dates on materials from these river systems. Thus, chronostratigraphic investigation within these fluvial systems is a virgin field of endeavor.

Consideration of the implications of the probable southward continuations of the White River system south of the Arkansas River during Stages 3 through 6, hypothesized herein for the first time, is equally as exciting. As shown in Plates 28H to 28K, it is interpreted that that river system flowed in a discrete, separate channel and meander belt to well south of Greenville and, at one time, to beyond Vicksburg. Except along the Black Bayou Meander Belt east of Greenville, all traces of the system have been obliterated by subsequent meandering by the Mississippi River. No direct archeological evidence thus remains in that area, but scientists interested in regional patterns of technology, raw materials distribution, and the diffusion of cultural traits should consider whether the hypothesized river system may have been a corridor conducive for their southward movement.

Arkansas River Meander Belts

In the two relatively recent attempts to establish an Arkansas River meander belt chronology (Saucier 1974, Saucier and Snead 1989, Autin et al. 1991), seven separate meander belts (the present and six abandoned ones) are recognized and believed to represent approximately the past 12,000 years. Both the number of meander belts and their estimated ages have been revised herein based on a more careful examination of cross-cutting relationships and a broader consideration of archeological evidence (Figure 50).

In the previous interpretations, a short east-west trending meander belt segment east of Altheimer, Arkansas (Plate 7) was designated as Stage 5. However, reinterpretation reveals that segment is truncated by a Stage 7 channel and obviously must be older. Accordingly, the segment is herein assigned a Stage 8 designation and is referred to as the Boggy Bayou Meander Belt (A8, Plate 1). No chronometric data exist in that area, and there is no basis for an accurate age estimation. If the estimated ages of the younger meander belts are correct, the Stage 8 channel was probably active between about 13,000 and 12,000 years B.P. (Figure 50). It is highly unlikely that it could be older than that because the Arkansas River was probably responding to climatically-induced higher discharges and flowing in a "Deweyville-like" regime to possibly as recent as about 14,000 years B.P.

The Stage 7 channel is manifest over a distance of about 40 mi by the Bayou Meto Meander Belt (A7, Plate 1) which trends along the eastern side of the Arkansas Lowland at the edge of the Grand Prairie. Near the community of Bayou Meto, Arkansas (Plate 7), the meander belt is truncated by the modern meander belt, and its route past that point is unknown. A complete absence of a possible continuation south of the modern meander belt (Stage 1) indicates that it must have followed the course of the present river and has been destroyed by it to as far east as the junction with the White River. At that point, it became tributary to the White River system and flowed south to eventually join the Mississippi River.

The appreciable width of the Bayou Meto Meander Belt and the relatively large number of cutoffs suggest that it was active for a considerable length of time. It is hypothesized that initial occupation took place about 12,000 years B.P. and continued to about 9,800 years B.P. (Figure 50). Radiocarbon dates are not available to firmly establish its time span, but an archeological site may be of help in determining the time of abandonment. According to House (1980), there is a Paleo-Indian or Early Archaic period site (Dalton culture) near Bayou Meto that appears to be associated with the Bayou Meto Meander

Belt. If that association is verified, it would be confirming evidence for abandonment of the system by no later than 9,800 years B.P.

Evidence for the Stage 6 channel consists of the small Bayou Macon Meander Belt (A6, Plate 1) which trends southward from the vicinity of the community of Winchester, Arkansas (Plate 8), to near Eudora, Arkansas. Between Winchester and Pine Bluff (Plate 7), the channel must occupy the position of the later Stage 4 channel, and a short distance south of Eudora it is hypothesized that the Stage 6 channel became tributary to the White River system (Plate 28H).

An examination of Plates 7 and 8 reveals that the Bayou Macon Meander Belt is considerably smaller than, for example, the Bayou Meto Meander Belt. Two explanations are possible, i.e., that the former was occupied for only a short period of time, or that it was a less than full-flow system. Although neither radiocarbon dates nor archeological evidence is present, this writer estimates that Stage 6 channel was active from about 10,000 to 8,500 years B.P. (Figure 50). If that was the case, a considerably larger meander belt should have formed. Therefore, it is more reasonable to attribute the smaller size to a condition of divided flow. The remainder of the flow may have still been in the earlier Stage 7 channel, or it may have been in the newly forming Stage 5 channel.

Two short meander belt segments, one just southeast of England, Arkansas (Plate 7), and the other just west of Gould, Arkansas (Plate 8), have been identified in the Arkansas Lowland and tenuously attributed to the Stage 5 channel. The more certain evidence involves the Crooked Bayou Meander Belt (A5A, Plate 1) and its continuation the Gourd Bayou Meander Belt (A5B, Plate 1). Together, the two meander belts extend for more than 120 airline miles and mark the first well-developed ones west of Macon Ridge in the Boeuf Basin and continuing into the Ouachita River Lowland (Plate 1). Considerably older and mostly buried meander belt segments occur but cannot be correlated with any particular stage.

Considerable longitudinal variation exists in the geometry of the two meander belts, but it appears certain that the channel carried the full flow of the Arkansas River for an appreciable period of time. Variations in width and numbers of cutoffs from one meander belt segment to another correlate extremely well with the nature of the bed and bank materials. For example, the zone of point bar accretion is noticeably wider and the number of cutoffs greater where the channel flowed through the Early Wisconsin Stage valley train deposits of Macon Ridge than where it flowed through the backswamp deposits of the Boeuf Basin and Ouachita River Lowland.

Once again, no radiocarbon dates or diagnostic archeological sites are present to help establish the age and duration of this stage. By inference, its chronological position is placed between about 9,000 and 7,000 years B.P. (Figure 50). Thus, the Arkansas River Stage 5 would be coincident with the Mississippi River Stage 5 (Plate 28I), and it is hypothesized that the two fluvial systems joined a short distance south of where the former flowed through the gap between Sicily Island and the upland area to the west (Plate 10).

As a consequence of an apparent avulsion in the vicinity of England, the Arkansas River established the Stage 4 channel and proceeded to form another extensive system marked by Bakers Bayou, Choctaw Bayou, Joe's Bayou, and Cross Bayou Meander Belts (A4A, A4B, A4C, and A4D, Plate 1). In doing so, the river abandoned the Boeuf Basin and once again flowed east of Macon Ridge. It was, however, the first and only time that the river flowed as far south as to enter the Tensas Basin in a separate channel.

Considerable uncertainly exists regarding the inferred relationship between the Stage 4 channel and the White River meander belt from about 30 mi north to about 50 mi south of Greenville. As shown in Plate 28J, the preferred interpretation is that the two river systems flowed close together but separately at that time. That interpretation is based on the absence of distinctive Arkansas River sediments along the Black Bayou Meander Belt east of Greenville, and on an abandoned channel in the Cross Bayou Meander Belt (Plate 10) that is too small to have accommodated the discharge of both rivers.

It is apparent in Plates 7, 8, and 9 that the Joe's Bayou Meander Belt is considerably larger than the others of this stage. Because of conditions farther downstream, this cannot be explained by the presence of White River discharge. The most viable explanation appears to be that when the Mississippi River adopted its Stage 2 meander belt about 4,800 years B.P. (see Plate 28L), for a short time one of two events (or possibly both) occurred. One was that the then-abandoned Joe's Bayou Meander Belt received a brief pulse of Mississippi River sediments from upstream near Lake Providence, Louisiana (Plate 9) or that it received similar sediments from downstream backwater flooding from near Sicily Island (Plate 10). Either or both of these events would have meant a modest reactivation with attendant renewed channel migration, natural levee growth, and cutoff formation.

Archeological sites, several with radiocarbon dates, are present to affirm and/or refine previous estimates of the age of the Stage 4 Arkansas River channel. Along the Bakers Bayou Meander Belt, House (1980) has reported sites dating to the Middle Archaic Stage that may be as old as 7,000 years B.P. On the basis of this information, this writer has revised the previous age estimate for the initiation of the Stage 4 channel (Autin et al. 1991) backward by several hundred years to about 7,500 years B.P. (Figure 50).

Several Poverty Point Period sites have been well documented along the natural levees of the Joe's Bayou Meander Belt (Webb 1977, Jackson 1982). Those indicate that the Stage 4 channel in that area was abandoned (or largely inactive) by at least 3,000 years B.P. and possibly as early as 3,500 years B.P.

Along the Cross Bayou Meander Belt south of Larto Lake (Plate 10), excavations at two Poverty Point Period sites have yielded eight radiocarbon dates that are especially chronologically significant (Debusschere, Miller, and Ramenofsky 1989). They establish that the meander belt was abandoned by the Arkansas River prior to 4,790 years B.P., and about that time the area started experiencing the effects of sedimentation from the Stage 2 Mississippi River channel.

The Stage 3 channel developed as a result of a diversion at the extreme upper end of the Arkansas Lowland near Little Rock. It developed a meander belt just east of the present course of the river to near Pine Bluff, and thence at the edge of the alluvial valley along the route of Bayou Bartholomew to as far south as Dermott (Plate 8). From there southward, the channel is indicated by the Lighterwood Bayou Meander Belt (A3A, Plate 1) and the Bonne Idee Meander Belt (A3B, Plate 1). As a consequence of the progressive overbank sedimentation that had been proceeding in the Boeuf Basin since the early Holocene, east of Bastrop, Louisiana (Plate 9), the latter meander belt was able to cut directly across the meander belt of the earlier Stage 5 channel. The valley alluviation also allowed the Bonne Idee Meander Belt to develop across the lower two levels of Macon Ridge which at that time was no higher than the floodplain of the Tensas Basin.

Valley alluviation and the resultant very low gradients additionally might have been a factor in the development of the extensive distributary systems that originated from the Bonne Idee Meander Belt (see Chapter 5 and Figure 17). Conceptually those systems should offer an unusually good opportunity for chronostratigraphic studies employing archeological evidence, but the limited effort to date (Kidder, Ring, and Saucier 1986) has not been particularly significant. Several sites have been investigated in the area, but none are older than the Late Archaic period and probably do not date earlier than about 4,000 years B.P.

Between Little Rock and Pine Bluff, the Stage 2 Arkansas River channel developed the Plum Bayou Meander Belt (A2A, Plate 1), and south of that area, the Bayou Bartholomew Meander Belt (A2B, Plate 1). The latter extends for over 100 airline miles to the vicinity of Bastrop (Plate 9) where the channel diverted from the Boeuf Basin through a narrow gap in the uplands (north of the Bastrop Hills) into the Ouachita River Lowland. Assimilating the flow of the Ouachita River, the Arkansas River continued southward along the western margin of the alluvial valley and through the gap west of Sicily Island before becoming tributary to the Mississippi River (Plate 28L). Between Sterlington, Louisiana (Plate 9), and Monroe, there are several short segments of abandoned meander belts, the origins and ages of which have not been established. At least one, now occupied by Bayou de Siard, appears to be definitely of Arkansas River origin. Two others, one occupied by Bayou Petticoat, exhibit smaller cutoffs and may have been formed by the Ouachita River during Stage 3 or earlier.

As indicated in Chapter 1, the advent of new chronological data and the period reinterpretations of the ages of meander belts in all parts of the alluvial valley (of both Mississippi and Arkansas river origin) invariably have meant assigning a greater antiquity to the features than previously. Such has definitely been the case on several occasions with the Stage 2 channel. In the most recent interpretation (Autin et al. 1991), that channel was estimated to have been active between about 4,000 and 1,800 years B.P. However, it is now known that artifacts dating to the Late Archaic period are common on the Plum Bayou Meander Belt (House 1980), meaning that the initiation of that meander belt must predate 4,000 years B.P. by at least several hundred years. A recently acquired, strategically located radiocarbon date[1] from point bar deposits of the Bayou Bartholomew Meander Belt is even more indicative. The date of 3,890 years B.P. indicates that the meander belt was well developed and several cutoffs had already taken place prior to that time. Direct evidence is not available as to the abandonment of the Stage 2 channel, but archeological sites indicate that the Stage 1 or modern channel of the river was active by about 2,500 years B.P., about 500 years earlier than previously estimated. If the time estimate is correct, it is reasonable to move the termination date of the Stage 2 channel from 1,800 years B.P. to 2,200 years B.P. (Figure 50).

Elsewhere, there are several sites with suites of radiocarbon dates that are on or closely associated with the Stage 2 channel, including the famous Toltec Mounds site located about 15 mi southeast of Little Rock. All of the sites, however, were occupied well after the channel had become inactive (e.g., during the Baytown period) and are not geologically meaningful.

As indicated previously, the Stage 1 channel is now estimated to have started developing about 2,500 years B.P., although the archeological evidence is actually not as definitive as in the case of the Stage 2 and 3 channels. The present river diverted from the Stage 2 channel in the vicinity of Pine Bluff and adopted a much shorter (and therefore higher gradient) route to the Mississippi River (Plates 7 and 8).

Allowing for the complicating effects of subsequent alluviation makes it quite apparent from Plates 7 and 8 that the modern meander belt is considerably wider than any of the previous ones. The number of cutoffs is comparable in number to that in several of the older meander belts, but they are noticeably larger and more complex. This difference in meander belt morphology cannot be explained, as it can in other cases, by differences in the nature of the bed and bank materials, or by duration of occupation. The explanation seems to lie in part with the number of historic period artificial cutoffs, but to a larger extent it may be a reflection of the steeper gradient which has imparted some braided-channel tendencies to the meandering regime. On the other hand, a climatically-induced response cannot be dismissed. It is interesting to note that the modern meander belts of both the Arkansas and the Mississippi rivers, which are of comparable age, are considerably larger than the older ones. This could be a reflection of a Holocene-long trend caused by a progressive change in the character of the sediment load, but is it possible that

[1] Personal Communication, 1993, H. Markewich, U.S. Geological Survey, Atlanta, GA.

it is a response to a postulated tendency toward a higher degree of discharge variability due to a more meridional atmospheric circulation pattern during the last 2,800 years?

Red River Meander Belts

Reconstructing the Holocene meander belt chronostratigraphy of the lower Red River valley, i.e., south of Alexandria, Louisiana, (Plates 10 and 11) has proven to be extraordinarily difficult--to the extent that no attempt has been made herein to estimate any absolute ages. That is essentially the same position adopted by Autin et al. (1991) who presented only a tentative, relative meander belt sequence, concluding that the previous chronological model (Saucier 1974) could not be substantiated even by indirect evidence and should be abandoned. This writer fully concurs, and no estimates are presented in Figure 50.

The *locations* of meander belts and relict Red River channels in the area are clearly evident in the surface physiography and have been mapped and described by Fisk (1944), Schultz and Krinitzsky (1950), and most recently by Russ (1975). Saucier and Snead (1989) made slight modifications to Russ's interpretation, and their version is presented herein. They recognize eight meander belt segments representing six channel stages (Plate 1): detailed meander belt configurations are shown in Plates 10 and 11. Russ presented a chronological model for the meander belt sequence, based in large part on relationships to Mississippi River channels. However, necessary changes in the chronology of the latter have pointed out the need for corresponding (but not yet accomplished) changes in the former.

It is important to mention certain observations regarding Red River meander belt ages even though accurate estimates are impossible. The lower Red River valley intersects the Mississippi alluvial valley near the head of the deltaic plain; consequently, it experienced significant floodplain degradation and entrenchment in direct response to the eustatically lowered sea level of the Late Wisconsin Stage. Subsequently, beginning about 12,000 years B.P., conditions reversed and the Red River valley began experiencing appreciable and relatively rapid alluviation and floodplain aggradation. This eventually led to the accumulation of 90 to 100 ft of overbank sediments, mainly in a backswamp environment, in the area southeast of Bunkie, Louisiana (Plate 11). If the sedimentation rates inferred from age/thickness relationships are considered, it is likely that any meander belts older than about 5,000 to 6,000 years will lack any surface expression. This therefore provides a rough estimate of the likely maximum age of the six observable channel stages.

Cross-cutting relationships discerned in detailed mapping (Smith and Russ 1974) indicate that at least the two oldest meander belts predate the final stages of full Mississippi River discharge through the Stage 3 channel in the Teche Meander Belt. However, because the Mississippi River was flowing continuously in that location since Stage 6 and because it has not been possible to ascribe specific features or parts of the meander belt to specific stages, those cross-cutting relationships are imprecise and not particularly helpful. Perhaps of greater potential value are relationships that indicate the younger Red River meander belts either occupy or cut across the abandoned Mississippi River course in the Teche Meander Belt and therefore obviously postdate the last Mississippi River flow in the Stage 3 channel.

Archeological evidence relating to meander belt ages is sparse and thus far has been of very little help. Because of the rapid sedimentation rates in the lower part of the valley, sites older than the Middle Archaic period, if present, are certain to be buried. Indeed, the only sites of that antiquity thus far discovered have been along the edges of terraces overlooking the valley. Sites dating to the last 2,000 years or so (the latter Formative Stage) are similarly sparse, suggesting possible unfavorable environmental conditions for permanent habitation during the late prehistoric period. To date the only known case where an archeological site appears to be chronologically meaningful involves the time of the establishment of the Stage 1 (modern) channel and its diversion through the Moncla Gap (Plate 1). That event was originally estimated at about 500 years B.P. by Saucier (1974) and later at about 600 years B.P. by Russ (1975), but a date of about 1,800 years B.P. or earlier now appears to be more viable (Pearson 1986). Changes of that magnitude are indicative of the high degree of uncertainty that exists in dating any Red River valley Holocene events.

Mississippi River Delta Complexes

Discussions in Chapter 5 cover the formation and characteristics of deltaic environments of deposition, delta lobes, and delta complexes. Since the 1930s, there has been little disagreement over and refinement in the identification of at least the principal distributaries, but several classification schemes have been advanced. Most of the interpretations differ only slightly in their recognition of 16 discrete lobes, but much more so in their assignment of lobes to particular complexes. Considerable changes in nomenclature have taken place, however.

Fisk (1944), elaborating on the work of Russell (1940), recognized six "deltas" (delta complexes) and subdivided those into subdeltas (delta lobes). Kolb and VanLopik (1958) were responsible for the first major revision to Fisk's scheme and proposed a delta framework consisting of seven "deltas" and introduced several new names (Figure 29). In both studies, the estimated total time span involved about 5,000 years. Subsequently, Frazier (1967) used over 160 radiocarbon dates to develop a much more detailed stratification of the deltaic plain (16 lobes but only 5 complexes) and to extend the time span to about 7,200 years. Although slightly modified in subject works, notably those of Penland and Boyd (1985) and Penland, Boyd, and Suter (1988), Frazier's work remains the most definitive to date and is heavily relied upon herein. Considering the paucity of documentation published by Penland, Goodwin et al. (1991) have concluded that most of the differences between the two interpretations are due to simple stratigraphic miscorrelations.

Outer shoal

Prior to about 12,000 years B.P., the rate of postglacial sea level rise was so rapid that it is unlikely that a deltaic plain analogous to that of today was able to form. Instead, fluvial sediments were probably reworked and winnowed in the littoral zone with finer materials carried offshore and the coarser materials deposited as sand sheets or shoals in a nearshore Gulf environment. After 12,000 years B.P. and especially after about 10,000 years B.P., a substantial decline in the rate of sea level rise made the formation of a deltaic plain much more likely or at least significantly increased the volume of fluvial sediments being deposited in shallow-water offshore areas. However, no direct evidence for a deltaic plain greater than 10,000 years B.P. has been detected and, if present, would lie on the continental shelf in water depths greater than 150 ft.

The oldest delta complex recognized by Frazier (1967) was the Maringouin. He dated the complex to between 7,300 and 6,200 years B.P. but indicated it could have begun forming a few thousand years earlier. Penland, Boyd, and Suter (1988) were the first to recognize tangible evidence for a sedimentary sequence older than the Maringouin Complex. Using extensive seismic and vibracore data from offshore central Louisiana, they identified two relict, submerged shorelines which they attributed to temporary sea level still stands, and two ravinement surfaces formed by regional planation during episodes of relatively rapid sea level rise. According to their interpretation, deposits of the Maringouin Complex overlie the deeper (Holocene) ravinement surface (which occurs at a depth of 45 to 75 ft), and deposits of an older (Earlier Holocene) complex underlie the surface. They designated the latter as the Outer Shoal Delta Complex, and Goodwin et al. (1991) have suggested that it may date between 9,200 and 8,200 years B.P. No radiocarbon dates have been reported for this complex, and currently no one has attempted to postulate its areal extent. There is no question, however, that no *in situ* deltaic deposits (e.g., distributary natural levee or interdistributary marsh) have survived the Holocene transgression and sand shoals are its only manifestation.

Plate 28I shows the hypothesized location of the Outer Shoal Complex and the probable configuration of the Gulf shoreline between about 9,000 and 8,000 years B.P. The pattern of deltaic distributaries shown in that figure is purely illustrative, intended only to indicate that multiple distributaries were probably present.

Maringouin

The Maringouin Complex as described herein should not be confused with the delta of the same name described by Fisk (1944). The latter was located entirely in the Atchafalaya Basin rather than offshore from central Louisiana. The presently defined Maringouin Complex was named by Frazier (1967) in order to retain that designation for the oldest recognized delta complex even though it is an entirely different complex in a different area. He did identify clays in the central Atchafalaya Basin which he correlated with the offshore complex, but an actual connection between the two was not documented.

The location and extent of the Maringouin Complex have been inferred largely from the positions of the Tiger, Ship, and Trinity Shoals which are believed to represent the reworked upper portion of the deltaic sedimentary sequence (Autin et al. 1991) (Figure 29). The lower portion of the sequence is not known to contain deltaic distributaries; however, two zones of interdistributary peats, representing two discrete depositional cycles, were encountered in borings along Bayou Sale (Plate 13) and dated to between 7,240 and 6,150 years B.P. (Coleman and Smith 1964, Frazier 1967).

With consideration given to the depth of the peat horizons below present sea level, and correction made for the effects of regional subsidence, it is estimated that sea level was perhaps 25 ft below its present level when the Maringouin Complex began forming and no more than about 15 ft below present when it was abandoned. It is reasonable to assume that a relative still stand in the Holocene transgression was responsible for its initial development, but *not* that a resumption of a relatively rapid rise was responsible for its abandonment. Although most workers consider the Maringouin Complex and the subsequent Teche Complex to be discrete units, Penland et al. (1991b) consider the two to be transitional and associated with a single delta plain lying between two ravinement surfaces. Under their scenario, it would be necessary to assume that distributaries of the Maringouin Complex would have been abandoned as sea level rose, the position of the overall coastline shifted slightly inland, and new distributaries formed as sea level approached to within 10 ft of its present level. Irrespective of those considerations, it is apparent that the trunk channel for the Outer Shoal, Maringouin, and Teche complexes remained in the same meander belt.

The pattern of the Maringouin Complex deltaic distributaries shown in Plate 28J once again is to be considered only diagrammatic. That illustration does show, however, that sea level had risen high enough that deltaic plain development no longer was constrained by the western valley wall and extended several tens of miles farther westward than did the Outer Shelf Complex.

Teche

With a continued slow rise in sea level about 6,000 years B.P., the outer margins of the Maringouin Complex were submerged and eroded, distributaries were abandoned, and the locus of deltaic sedimentation shifted slightly inland. New distributaries began forming from the Teche Meander Belt between about Jeanerette and Morgan City, Louisiana (Plate 13), and deltaic sediments were deposited on top of the surviving Maringouin Complex deposits in what is believed to have been a poorly drained swamp environment (Smith, Dunbar, and Britsch 1986). Based on the work of Coleman (1966b) and VanLopik (1955), the earliest distributaries with surface manifestation formed in the western portion of the complex and included Bayou Cypremort and Bayou Sale (Plate 13). Radiocarbon dates place their development between about 4,700 and 4,200 years B.P. (Coleman and Smith 1964). Somewhat later, distributaries developed farther to the east and include a complex that radiates to the southeast and south of Morgan City south of Bayou Black. Because of greater subsidence than to the west and later sedimentation from the Lafourche Complex, the distributaries are largely obscured and not mapped on Plates 13 and 14. Their locations are shown in Smith, Dunbar, and Britsch (1986) along with radiocarbon dates that place their formation between about 4,500 and 3,500 years B.P.

Subsurface data indicate that the Bayou Cypremort and Bayou Sale distributaries developed over interdistributary marsh surviving from the Maringouin Complex. The distributaries mapped by Smith, Dunbar, and Britsch (1986) near Morgan City apparently developed into shallow open water. From this consideration, it was concluded that the maximum eastern limit of the Teche Complex was located about 10 mi west of Houma (Plate 14). However, Weinstein and Gagliano (1985) cite archeological evidence to include other distributaries in the complex, the effect of which would be to place the eastern margin of the complex about 40 mi farther to the southeast. They argue that some distributaries traditionally considered to be part of the later Lafourche Complex were actually Teche Complex features that were reoccupied (rather than created) in Lafourche times. The latter interpretation appears most viable and, consistent with the most recent delineation of delta complexes (Saucier and Snead 1989), is retained herein.

The paleogeographic "snapshot" shown in Plate 28K, representing the period from about 6,000 to 4,500 years B.P., portrays only the early portion of the total Teche Complex since that part of the deltaic plain was a time-transgressive sequence covering perhaps 3,000 years of geologic time (Figure 50). The later portion is illustrated in Plate 28L which represents the time from about 4,500 to 3,000 years B.P.

Two widely divergent scenarios have been offered concerning sea level between about 6,000 and 5,000 years B.P. The older and more widely accepted scenario, supported by studies such as those of Coleman and Smith (1964) and Saucier (1963), indicates that sea level rose from about 10 to 12 ft below present about 6,000 years B.P. to within 5 ft of its present level by 5,000 years B.P. Those studies (and others) have used radiocarbon dates to calculate and separate the effects of local and regional subsidence from true eustatic sea level rise. More recently, Penland et al. (1991b) have employed a similar approach but argue that sea level was about 18 to 20 ft below present during that time frame. This writer still strongly favors the former scenario because Penland et al. (1991b) have failed to reconcile a considerable volume of archeological evidence that is in direct conflict with their views. That evidence primarily consists of numerous *Rangia* shell middens in the Pontchartrain Basin area which, calculating for subsidence, would not be located where they are if sea level had been more than 5 ft below present. The favored scenario is critical to the paleogeographic reconstruction shown in Plate 28K.

Active deltaic sedimentation was taking place along the western side of the Atchafalaya Basin and offshore from central Louisiana while the eastern side of the basin evidently was a shallow (i.e., less than 10 ft deep) embayment. Estuarine conditions (paludal and shallow-water environments) extended as far inland as Baton Rouge. Evidence for the embayment includes occurrences of radiocarbon-dated, 5,600- to 5,500-year-old *Rangia* shells in paludal sediments in the central basin area, and similarly dated (5,475 years B.P.) shell middens along the edge of the Prairie Complex just south of Baton Rouge (Saucier 1963).

Farther east in the Pontchartrain Basin, the interval between 6,000 and 4,500 years B.P. was the time of the formation of the previously discussed Pine Island Beach Trend. Besides the radiocarbon dates from the beach ridge proper, there are at least six others from bay-sound or nearshore Gulf deposits that verify that open water conditions prevailed from at least 5,400 to 4,800 years B.P. (Saucier 1963).

St. Bernard

Penland, Boyd, and Suter (1988) and Penland et al. (1991b) have recently interjected controversy into the character and timing of the initiation of the St. Bernard complex. This controversy occurred because of their emphasis on a step-like shape to the curve depicting the last several thousand years of sea level rise (still stands followed by short periods of rapid rise) rather than a steady rise at a slowly declining rate. According to their scenario, sea level rose abruptly from 18 to 20 ft below its present level at 4,000 years B.P., attaining its present level about 3,000 years B.P. That rate of rise was so rapid that deltaic sedimentation in the Teche Complex could not keep pace, and the lower portion of the deltaic plain to within 10 mi of Houma (Plate 14) was submerged and destroyed by marine erosion (the Teche ravinement). Only after 3,000 years B.P., under a relatively constant sea level, was the Mississippi River able to start constructing a new deltaic plain, i.e., the St. Bernard Complex.

Apparently, there are at least two major problems with that scenario. First, Weinstein and Kelley (1992) have recently discovered several archeological sites with probable Poverty Point Period components on distributaries well south of Houma. If that cultural affiliation is correct, the distributaries must date to the Teche Complex (rather than the later Lafourche Complex), indicating that the ravinement, if it existed, did not extend as far north as proposed. Otherwise, the sites would have been destroyed. Second, Penland et al. (1991b) gave no consideration to events in the alluvial valley area that were essential, causal factors in the formation of the St. Bernard complex. Specifically, they ignored the ramifications of the dated and documented abandonment of the Stage 3 (Teche) meander belt and the shift of the Mississippi River to the eastern side of the valley. Thus, for those reasons, this writer favors and presents below the "traditional" interpretation of Frazier (1967).

The first definitive chronostratigraphic evidence relating to the initial development of the St. Bernard Complex was discerned 30 years ago when Gagliano and Saucier (1963) discovered a buried Poverty Point site situated on a deltaic distributary that formed shortly after 4,000 years B.P. in the New Orleans area. At first, the distributary was correlated with the Cocodrie delta of Fisk (1944) since the St. Bernard Complex was believed to have originated at least a thousand years later. However, before long the nebulous concept of a Cocodrie delta was abandoned, and a positive link was made between the deposits in the New Orleans area and a meander belt (buried beneath the modern one) between Baton Rouge and New Orleans (Saucier 1963). Shortly thereafter, Frazier (1967) acquired radiocarbon dates and stratigraphic evidence that indicated the meander belt dated to as early as 4,650 years B.P. and that the initial delta lobe had actually extended well east of New Orleans into St. Bernard Parish along Bayou Terre aux Boeufs (Plate 14) by 4,000 years B.P.

It is hypothesized that the Mississippi River rapidly filled the shallow embayment south of Baton Rouge and extended its course as far as 40 mi southeast of New Orleans within a few hundred years after it abandoned its Stage 3 channel and diverted to a course along the eastern valley wall south of Old River about 4,800 years B.P. (Figure 50, Plate 28L). Even though the river was still discharging part of its flow into the younger part of the Teche Complex (and did so for perhaps a thousand years), it was able to extend a new lobe so rapidly and so far because of the extremely shallow water (generally less than 50 ft) into which it was depositing sediments. By about 3,400 years B.P., the Bayou Terre aux Boeufs distributary was abandoned in favor of one to the south of New Orleans along Bayou des Familles. Frazier (1967) estimates the latter remained active for about 1,400 years.

About the time the Mississippi River finally abandoned the Teche Complex about 3,000 years B.P., the full discharge of the river was not directed to the St. Bernard Complex (Figure 50). Rather, it is believed that a distributary formed along the route of Bayou Terrebonne (Plate 14), initiating the development of the Lafourche Complex. No attempt has been made to quantify the amount of flow or sediment load in each system, but it is apparent that enough was still being discharged through the St. Bernard Complex to form several additional lobes. In sequential order, those included the Bayou La Loutre lobe, an unnamed lobe into southern St. Bernard Parish south of the latter lobe, and the Metairie Bayou-Bayou Sauvage lobe, all of which formed between about 3,000 and 1,000 years B.P.

Based on the sizes of the abandoned distributary channels, the amount of lateral migration (as evidenced by point bar deposits), and the areal extent of the lobes, it is evident that the Bayou des Familles and Bayou La Loutre lobes were the largest (but not necessarily active for the longest period of time) and the Metairie Bayou-Bayou Sauvage lobe was the smallest. Cumulatively, the St. Bernard Complex achieved the largest areal extent of any of the Holocene complexes, primarily a reflection of the shallowness of the receiving water body. Analyses of bottom sediments and data from high-resolution acoustic subbottom profiling substantiate that at maximum development, the Bayou La Loutre and adjacent lobes extended distributaries and interdistributary marshes as far as 20 mi east of the Chandeleur Islands (Plate 14 and 28L).

In a comprehensive geoarcheological study of the Pontchartrain Basin (Saucier 1963), it was interpreted that because of the partial blocking effect of the Pine Island Beach Trend, even at maximum extent the Metairie Bayou-Bayou Sauvage lobe did not significantly reduce the size of Lake Pontchartrain. In a recent study emphasizing the effects of neotectonic structural movement on the basin, Lopez (1991) hypothesized that the lake was almost entirely filled with sediment by the delta lobe. That interpretation is not viable in light of the sequence of lake beaches and abundant shell middens spanning the entire Formative Stage which parallel most of the south shore of the lake. The archeological data indicate the southern shoreline of the lake was never more than a mile or so from its present location and had essentially the same configuration for at least the last 2,200 years.

Recently, another aspect of Frazier's (1967) chronological model has been challenged in light of new data from offshore vibracores, high-resolution seismic profiles, and radiocarbon dates. Levin (1991) has introduced evidence indicating that the Bayou des Familles lobe is considerably older than previously believed, dating to as early as 4,600 years B.P. and remaining active for about 1,000 years. That interpretation appears to be compatible (at least not refuted) by archeological evidence. If correct, it would either (a) *significantly* increase estimates of the amount of sediment introduced into the deltaic plain soon after the 4,800-yr-B.P. diversion of the river to the eastern side of the alluvial valley, or (b) necessitate revising backward by several hundred years the date of the diversion of the river to the Stage 2 meander belt. Of the two possibilities, the latter appears to be less likely.

Lafourche

According to Frazier (1967), the first manifestation of the Lafourche Complex appeared about 3,500 years B.P. when the Bayou Terrebonne lobe developed from the Mississippi River trunk channel near Donaldsonville (Plate 11) southward into the Houma area (Plate 14). However, most of the Mississippi River's discharge at that time was being directed into the St. Bernard Complex. About 2,000 years B.P., all of the St. Bernard Complex distributaries except the Metairie Bayou-Bayou Sauvage lobe were abandoned, and the locus of sedimentation shifted to the Lafourche lobe. That rapidly led to the formation of the complex, branching, fan-shaped pattern of distributaries that radiate to the southwest, south, and southeast of the Thibodaux-Houma area (Plate 14) (Smith, Dunbar, and Britsch 1986).

Development of the initial Lafourche Complex lobe had a major impact on the lower portion of the Atchafalaya Basin as well as the area to the southeast. Prior to the formation of the lobe, the mouth of the basin merged with a broad, estuary-like, intradelta depression between Houma to the south and Donaldsonville to the north. No doubt brackish-water conditions extended well inland from that line at least periodically. However, with the development of the Lafourche lobe, the lower end of the basin was blocked by a natural levee ridge that extended from the trunk channel of the river across to the Teche Ridge. That development no doubt was what caused the Lower Atchafalaya River to cut through the Teche Ridge at Morgan City to provide an outlet to the Gulf for the interior drainage of the basin. Henceforth, the basin became an exclusively freshwater environment throughout.

Between about 2,000 and 1,000 years B.P., a wave of deltaic sediments spread across most of the Terrebonne Parish area. Most distributaries developed into marsh and shallow lakes in the subsiding and deteriorating distal portion of the Teche lobe (Plate 28M). Some of the surviving Teche Complex distributaries apparently were reoccupied, and the drowned, distal end of the Teche meander belt between Gibson, Louisiana, and Houma (Plates 13 and 14) was reoccupied, but with flow moving from east to west rather than *vice versa.*

Most workers agree that the prominent Bayou Lafourche distributary *per se*, which trends southward from Donaldsonville through Thibodaux to the Gulf of Mexico just west of Grand Isle, Louisiana (Plate 14), represents the last major active channel of the Lafourche Complex. Radiocarbon dates indicate that a significant meander belt ridge started forming no later than about 800 years B.P. Flow through that system started declining within several hundred years but did not cease entirely until 1904 when it was artificially closed at Donaldsonville.

Prior to the formation of the Lafourche Complex, the location of the Gulf shoreline in the chenier plain area is unknown. Because sea level was lower than at present, and in view of the extremely shallow gradient of the offshore area, the shoreline was definitely south of the southernmost cheniers and probably well south of the present shoreline. Thus, the entire history of chenier formation can be directly linked with the growth and decay of delta lobes in the Lafourche Complex.

There is disagreement between Gould and McFarlan (1959) and Penland, Suter, and McBride (1987) as to how far inland the Gulf shoreline transgressed prior to the formation of the first cheniers. The former believe that about 3,000 years B.P. the shoreline approximated the edge of the Prairie Complex (Plate 13) while the latter believe the oldest chenier trend--the Little Chenier trend--marks the maximum inland extent of the Gulf. Regardless, the ages of the Little Chenier and younger trends have been well established with radiocarbon assays (Gould and McFarlan 1959) (Figure 50), but precise relationships to particular delta lobes have not been attempted. It can only be presumed that mudflat formation along the western Louisiana coast was most active when the more western lobes of the complex were near their maximum extent and, conversely, coastline retreat and chenier formation coincided with progradation of the more eastern lobes. In this regard, there is general but not complete agreement between the delta lobe chronology of Frazier (1967) and the dates of the major chenier trends, suggesting that the relationship may be more complex than generally presumed. Attempts have not yet been made to factor in the possible effects of minor sea level oscillations and/or variations in climate (as manifest by changes in tropical storm frequency).

Plaquemines

When the main portion of the Lafourche Complex was actively prograding between about 2,000 and 1,000 years B.P., formation of the Metairie Bayou-Bayou Sauvage distributary attests that the trunk channel of the Mississippi River between Donaldsonville and New Orleans decreased in size but obviously was not completely abandoned. About the same time that the Bayou Lafourche distributary became active (1,000 to 800 years B.P.), there was an apparent increase in discharge past New Orleans, initiating development of the Plaquemines Complex. The increase in discharge continued in direct proportion to the decrease in the Bayou Lafourche lobe, with a majority of the river's flow being funnelled into the newly forming complex by 600 years B.P.

The Plaquemines Complex is marked by a series of small distributaries that radiate southeastward from a point about 10 mi southeast of New Orleans (Plate 14). Those indicate that the Plaquemines Complex prograded rapidly through the intradelta lowland between the subsiding and deteriorating margins of the Bayou des Familles lobe to the west and the Bayou La Loutre lobe to the east. The progradation continued rapidly until in late prehistoric times (500 to 400 years B.P.?), the delta front began approaching the edge of the continental shelf. That rather abrupt and first encounter with relatively deep water dramatically changed the physiographic character of the active delta lobe as well as its sedimentary framework. Thus began the last phase of the Plaquemines Complex which is also often referred to as the Balize or Modern delta (or subdelta).

Atchafalaya

After more than 10,000 years of being marginal to active sedimentation and meander belt formation under the influence of a slowly rising base level, it was inevitable that the Atchafalaya Basin would be the site of an attempt by the Mississippi River to divert to a new course to achieve a steeper gradient. In what arguably is his second most important contribution, Fisk (1952) recognized and described the beginning and evolution of that process which began in early historic times. That work recognized the pivotal role played by 19th century raft removal activities, artificial channel changes, and navigation improvements (in the Old River area at the junction of the Red and Mississippi rivers) in assuring a diversion and was perhaps the most important argument in justifying the need for an elaborate artificial control structure. Completed in 1963, the Old River Control Structure was designed and has been operated on the premise that because of the extent to which a diversion had progressed, it was economically and technically feasible to prevent its further progression but not to reverse the process. To that extent, nature has had its way. The Atchafalaya River will remain a major distributary rather than become the trunk channel of the river.

Development of the Atchafalaya delta lobe by the combined discharge of the Mississippi and Red rivers has been the only geomorphic event of that scale to occur in modern times in the Mississippi Valley area. Consequently, it has become a virtual laboratory for the study of deltaic processes (Roberts, Adams, and Cunningham 1980; Van Heerden et al. 1991). Actually, the first manifestation of the lobe has been well inland in the basin area. Prior to the early 1930s, the Lake Fausse Point-Grand Lake-Six Mile Lake-Flat Lake complex in the lower basin area (Plates 11 and 13) was an open freshwater lake environment. However, as well documented by several series of aerial photography and special sedimentological studies (see Chapter 6), the development of a lacustrine delta between the 1930s and the 1970s has reduced the open water to a tiny fraction of its previous extent (Gagliano and Van Beek 1975). Elsewhere in the basin area, heavy overbank sedimentation accompanying the increased sediment load is threatening to make significant ecological changes in the inland swamp environment. A meander belt ridge with well-developed natural levees has not been allowed to form, however, since the river is confined by artificial flood control levees between Old River and approximately Interstate Highway I-10.

Once the natural sediment trap provided by the lakes in the Atchafalaya Basin was essentially filled, the locus of active delta lobe formation shifted to the coast. Surveys revealed that by the early 1950s, a subaqueous delta began forming in Atchafalaya Bay at the mouths of the Lower Atchafalaya River and Wax Lake Outlet, an artificial channel about 10 mi to the west (Shlemon 1975). As a consequence of a major flood on the Mississippi River, a subaerial lobe began forming in 1973 and expanded rapidly due to several consecutive major annual floods. The pattern of development has been extremely well documented, and its future growth has been the subject of several predictive models.

With rapid delta deterioration due to both natural and anthropogenic causes now characterizing essentially the entire deltaic plain, the Atchafalaya Bay and environs are the only areas experiencing an expansion of intratidal marshes. The rate of expansion experienced during the last two decades will not persist indefinitely for several reasons, however. Delta lobe construction will take place in progressively deeper water, and there likely will be a decline in the sediment load entering the system. To date, a significant portion of the sediment has been derived from the widening and deepening of the Atchafalaya River channel, but that contribution has now ended as the discharge has been stabilized. Further, there has been an appreciable reduction in the sediment yield of the Red River as a result of recently completed navigation improvements on that system. It can only be expected that similar reductions in sediment yield will characterize both the Mississippi and Red rivers in future years as erosion and soil loss in the drainage basins are reduced by improved land management practices and resource conservation.

8 Tectonics and Neotectonics

New Madrid Seismic Zone

Parkfield! Loma Prieta! Northridge! Periodically a major seismic event will instantly focus national attention on a small and little publicized community or area. Such was the case in late 1811 and early 1812 when the tiny settlement of New Madrid, Missouri, became notorious for being near the epicenter of a series of earthquakes that rank among the largest in historical times in eastern North America. Three and possibly four major seismic events in the series had estimated magnitudes of 7 m_{bLg} (Street and Nuttli 1984) and were felt as far away as Boston, Washington, DC, and Charleston.

For more than a century following the events, descriptions focused on surface effects such as doming, subsidence (sunk-land formation), widespread liquefaction features (sand blows), fissures, and landslides. With the limited knowledge of geologic structure and tectonics of the middle and late 19th century, little progress was thus made in understanding the causal mechanism responsible for the seismicity. However, beginning about 1970, there was sudden upsurge in interest in applying the results of modern geophysical techniques and concepts of continental-scale seismotectonics to a study of the New Madrid Seismic Zone, prompted in large part by concern over the seismic design of nuclear power facilities (Stearns and Wilson 1972). Shortly thereafter, the U.S. Geological Survey sponsored a series of comprehensive, multidisciplinary investigations, several of which focused on crustal structure, and began interpreting the seismic zone in light of modern concepts of plate tectonics (McKeown and Pakiser 1982). Within the last 18 years, that same agency has tremendously advanced the state of knowledge through numerous studies (many still ongoing) funded under its National Earthquake Hazards Reduction Program. Many of those can be classed under the heading of paleoseismology, which is the study of the age, frequency, and size of prehistoric earthquakes (Crone and Omdahl 1987). As those authors point out, Quaternary geology has a critical role in paleoseismology because most of the evidence of major prehistoric earthquakes is preserved in Quaternary deposits or interpreted from geomorphic features.

Geologic structure

The New Madrid Seismic Zone is situated in extreme northeastern Arkansas, southeastern Missouri, and northwestern Tennessee along and just west of the axis of the Mississippi Embayment (Figure 6). Knowledge of and concepts regarding the deep crustal structure of the area in relation to the history of the seismic zone are evolving quite rapidly with new articles appearing each year.

In overview, it is important to recognize that the seismic activity in the zone is not attributable to a single, well-defined, strike-slip, or thrust fault such as would be the case for shallow faults breaking the surface in a continental margin area like California. Rather, the zone is located within an ancient, failed, intraplate rift in which faulting is present but is extremely complex and largely (but possibly not exclusively) without surface expression.

The northeast-southwest trending Reelfoot Rift (Figure 52), an elongated graben, originated in late Precambrian time (Figure 3) and was accompanied by the upwelling and intrusion of magma into the crust (Ervin and McGinnis 1975). In the early Paleozoic, a basin or trough approximately coincident with the modern embayment formed because of isostatic subsidence into which several miles of sediment were deposited. Widespread uplift occurred in the midcontinent area in the middle to late Paleozoic, forming the Ozark Uplift to the west and the Nashville Dome to the east. An uplift of lesser magnitude bridged between the two across the rift and elevated the sedimentary sequence, causing the Pascola Arch. A period of rift reactivation and intrusion occurred in the late Mesozoic, prompting renewed isostatic subsidence in the embayment and forming the elongate depositional trough observed today. The igneous intrusion is manifest by a series of plutons which flank or lie within the rift zone. Subsequently, subsidence has continued, forming the present Mississippi Embayment and promoting the deposition of the thick Cretaceous and Tertiary sedimentary sequence. Along the rift margins, the Paleozoic rocks are vertically offset by about 3,000 ft.

Figure 52 illustrates that most of the recorded earthquakes in the New Madrid Seismic Zone are centrally located within the Reelfoot Rift along a feature called the Blytheville Arch or associated with the Pascola Arch (which may be a single feature). Although the pattern of epicenters shown in that figure is actually based on microearthquakes (Luzietti et al. 1992), their distribution is essentially the same as that of historic period earthquakes recorded by strong-motion instrumentation. Several mechanisms have been proposed for the formation of the Blytheville Arch, including reverse faulting and igneous intrusion, but McKeown et al. (1990) believe it was formed in the late Paleozoic by diapirism along with the Pascola Arch. Those authors point out that most earthquake hypocenters in the seismic zone occur within the weak, deformed, and fractured sedimentary rocks of the arches above the crystalline basement rocks and are probably associated with fault zones in that area.

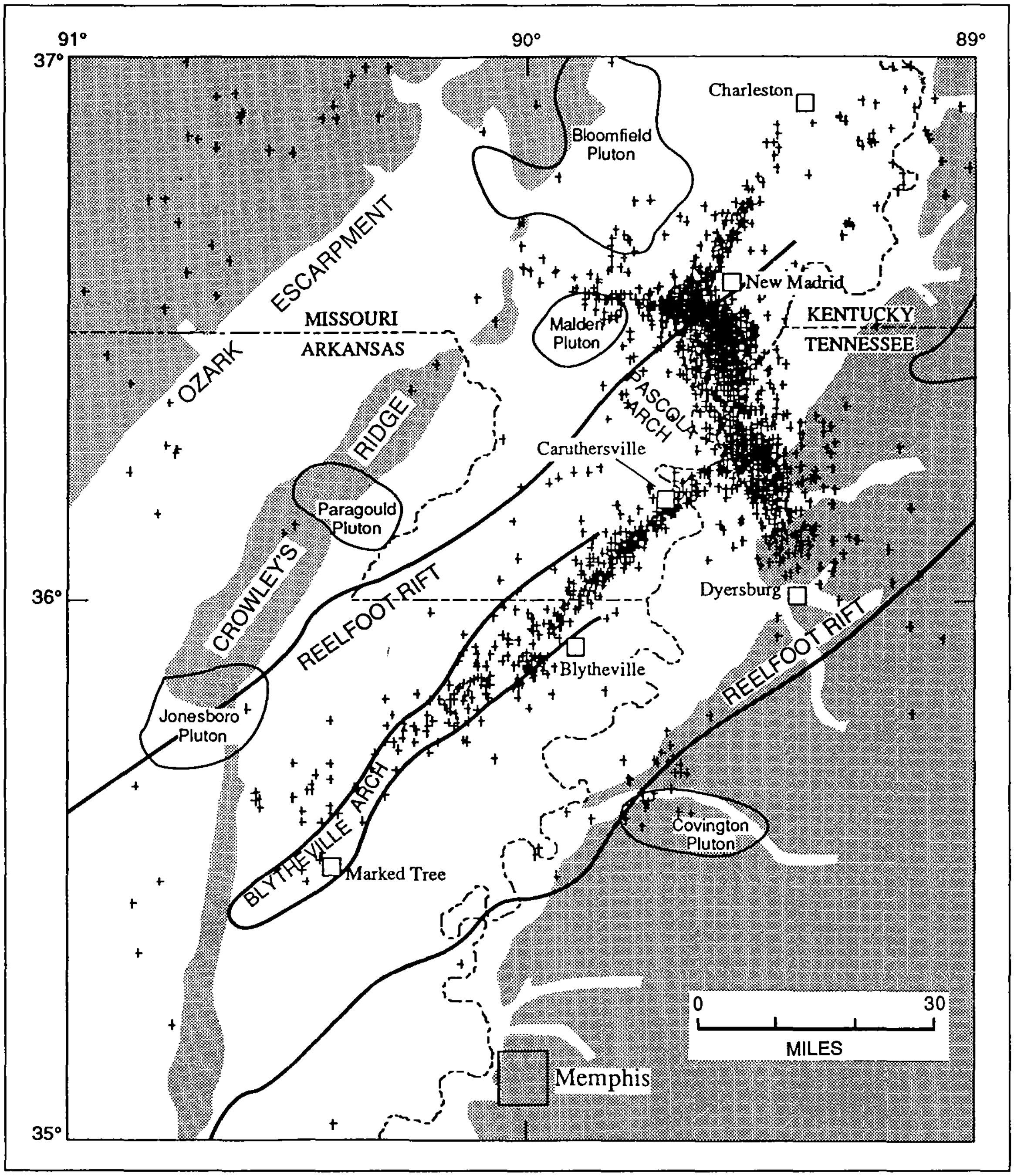

Figure 52. Major structural features of the New Madrid Seismic Zone along with epicenters of microearthquakes (modified from Luzietti et al. 1992)

Seismicity

The infamous 1811-1812 earthquake series, known as *the* New Madrid Earthquake, traditionally has been considered as involving three major events. The first occurred on December 16, 1811 (M_s=8.6, m_b=7.2, Modified Mercalli Intensity or MMI=XI) and is estimated to have been located along the Blytheville Arch between Blytheville and Marked Tree (Hopper, Algermissen, and Dobrovolny 1983) (Figure 52). The second, of slightly less magnitude (M_s=8.4, m_b=7.1, MMI=X-XI), occurred on January 23, 1812, and was probably located near the town of Caruthersville. The third, on February 7, 1812, was the strongest (M_s=8.7, m_b=7.3, MMI=XI-XII) and was located over the Pascola Arch in the vicinity of New Madrid. Recently, Street and Nuttli (1984) concluded from a reanalysis of contemporary accounts that a second shock of comparable magnitude and intensity (m_b=7.0, MMI=XI) occurred only six hours after the first one; hence, the New Madrid series consisted of four major events.

In addition to the four major events, Fuller (1912) documented 203 damaging aftershocks during a three-month period following the initial shock. Hopper, Algermissen, and Dobrovolny (1983) mention that 18 of the shocks (including the four major ones) were strong enough to be felt in Washington, DC, and at least 1,874 were felt during the same period 200 mi away in Louisville, Kentucky.

Since the aftershocks have subsided, there has been continuing intense microactivity in the zone. Between 1812 and 1974, the area has experienced 488 earthquakes with a magnitude (m_b) of 3.0 or greater. Twenty earthquakes are classed as damaging (m_b=5.0 to 6.2, MMI=VI-IX) (Nuttli 1982), the events of 1843, 1895, and 1968 being considered as large earthquakes. As another measure of the nature of the continuing activity, 731 events were recorded in a four-year period in the mid-1970s by a recently installed microearthquake network (Stauder et al. 1976).

The keen interest that has been generated in the last several decades in documenting the history of seismic activity in the zone is obviously driven by the great concern over "when will it happen again," or, borrowing terminology from California, when will the "Big One" occur. Fueled by public opinion aroused by recent spurious forecasts or predictions of an imminent major earthquake, the interest has been manifest by a flurry of paleoseismological studies which are just now beginning to bear fruit. Paleoliquefaction features offer the greatest promise of reconstructing the prehistoric sequence of major earthquakes and are discussed in the next section.

Fissuring and liquefaction features

Occurrence and characteristics. Perhaps not more than a few hundred persons resided in the St. Francis Basin area in 1811 and 1812; consequently, eyewitness accounts were rare and only a few of those were ever recorded in newspapers or science journals (Fuller 1912, Krinitzsky 1950, Nuttli 1972). Although somewhat fanciful and exaggerated, the accounts unquestionably document the occurrence of such phenomena as widespread bank caving along the Mississippi River, reversal of riverflow, landslides, earth waves, and forest destruction. However, by far the most widespread and dramatic geomorphic responses were the occurrence of land fissuring and sand blows. The latter, pseudovolcanic features that ejected liquefied sand and water into the air and caused extensive flooding, understandably caused the greatest consternation and comment.

Liquefaction occurs in a mass of saturated, cohesionless sediment when there is a sudden total loss of shear strength resulting from excess pore water pressure caused by earthquake shaking. When the liquefied deposits are confined by a thin, cohesive topstratum and the pressure exceeds the overburden pressure, sands will penetrate into the topstratum and/or violently extrude to the surface through linear fissures or discrete vents. The result is a complex pattern of subsurface sand-filled vents, fissures, dikes, sills, and surficial, conical- or ridge-shaped masses of extruded sand. Figure 53 contains sketches of typical liquefaction features and associated ground disturbance as exposed in canal banks near Manila, Arkansas (Plates 5 and 6) (Wesnousky and Leffler 1992a,b).

Both the point bar deposits of Mississippi River meander belts and the glacial outwash deposits of the Early and Late Wisconsin Stage valley trains are highly susceptible to liquefaction (Obermeier et al. 1990), as noted in 1811 and 1812. Fuller (1912) was the first to map the surface distribution of sand blows and fissures in the New Madrid Seismic Zone, and Saucier (1977a) provided a more detailed delineation and classification based on soils maps and an extensive, comparative aerial photo interpretation (Figure 54). A slightly different and less detailed interpretation, based on a different classification of the amount of extruded sand, has been subsequently presented by Obermeier (1989). In both the latter studies, the total extent of the zone of liquefaction is essentially the same and involves an area of about 4,000 sq mi.

A comparison of Figures 52 and 53 reveals only a rough correlation between the pattern of seismicity as revealed by earthquake epicenters and the distribution of sand blows and fissures. The areas with the greatest amount of extruded sand (50 to 100 percent of the land surface) lie west of the Blytheville and Pascola arches and west of the epicentral positions of the major 1811 and 1812 events. There is a greater correlation between liquefaction phenomena and the age of the deposits, with the largest amount of liquefaction corresponding to the area of the Late Wisconsin glacial outwash. The most positive correlation, however, exists with topography and groundwater occurrence. The greatest concentrations of sand blows and fissures approximate the areas with the lowest elevations in which water tables are highest. It is believed that liquefaction phenomena are relatively poorly developed on the Early Wisconsin valley trains such as Sikeston's Ridge because those areas generally lack a cohesive topstratum and have relatively low water tables.

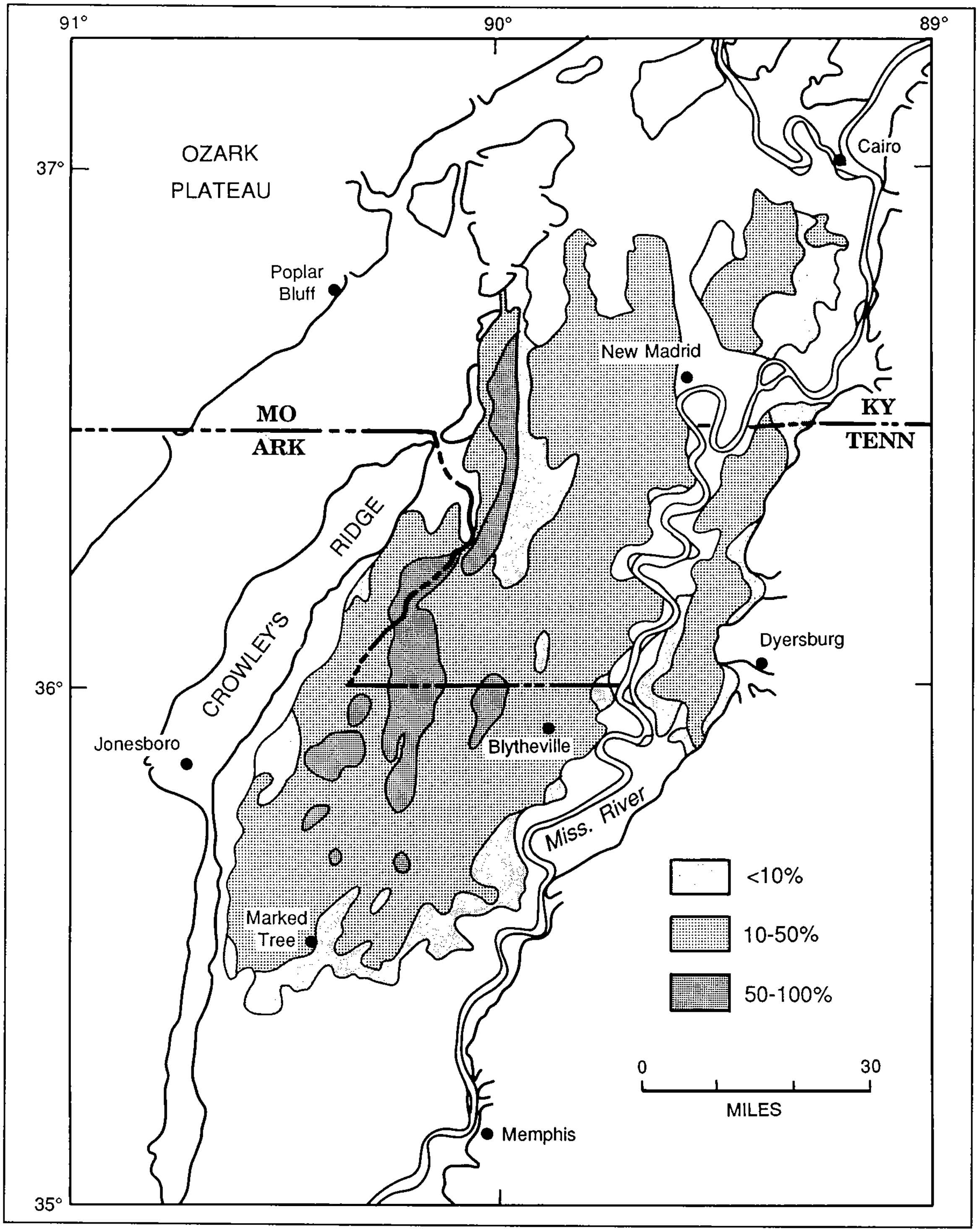

Figure 53. Distribution of liquefaction features (sand blows and fissures) in the New Madrid Seismic Zone (density classification based on estimated percentage of ground surface covered by extruded sand prior to widespread disturbance by agriculture)

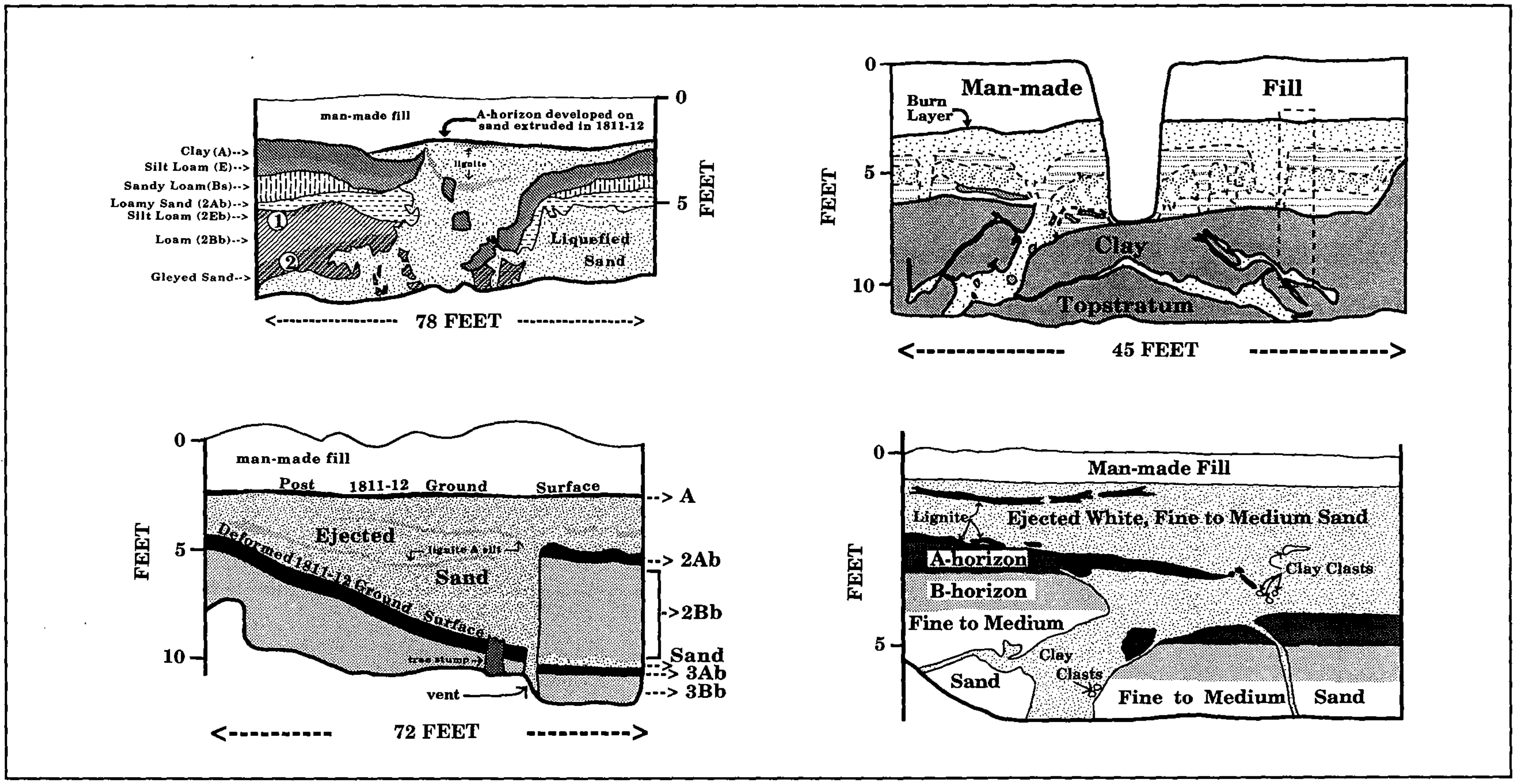

Figure 54. Typical characteristics of liquefaction features in the New Madrid Seismic Zone as revealed along banks of drainage ditches (modified from Wesnousky and Leffler 1992)

At a larger scale, it is apparent that subtle variations in topstratum thickness and surface topography are important in determining the detailed pattern of sand blow and fissure development. For example, the features are less numerous and smaller beneath the relict braided channels than they are beneath interfluve areas on a valley train where the topstratum may be only a few feet thinner. Similarly, abandoned channels in the Mississippi River meander belt generally have topstratums too thick and too cohesive to have allowed sand to extrude to the surface. Fissures are more numerous and more continuous along the flanks of small drainage features like streams and sloughs where the topstratum was capable of spreading laterally as it cracked and "floated" on the underlying liquefied sands.

Paleoliquefaction features and implications. This writer is credited with pioneering the use of liquefaction features to unravel the paleoearthquake history of a major seismic zone (Saucier 1977a). He was also the first to look to archeological sites and cultural deposits as a means of estimating the age of such features. After having examined various forms of ground disturbance at numerous sites throughout the New Madrid Seismic Zone, as recently as the late 1970s he had failed to find evidence of any features that could not be attributed to the 1811-1812 series. This presented a dilemma because extrapolation of magnitude-frequency statistics from post-1811 earthquakes implies a 550- to 1,100-year repeat time for major seismic events (m_b=7.0, M_s=8.0), and glacial outwash deposits throughout the zone are greater than 10,000 years old (Wesnousky and Leffler 1992). Evidence of multiple episodes of liquefaction should be preserved if the extrapolations are correct.

The first reported geologic evidence for paleoearthquakes was reported by Russ (1978) from a trench across a suspected fault just west of Reelfoot Lake (Plate 5). He interpreted that three earthquakes of sufficient size to produce liquefaction occurred during the last 2,000 years, suggesting a recurrence interval of 600 years or less. More recently, this writer has found evidence for multiple episodes of liquefaction at two archeological sites in the seismic zone. In one, located near Trumann, Arkansas (Plate 6), three distinct episodes were discerned; however, it appears that all three are associated with different events in the 1811-1812 series (Saucier 1989). In the other, cultural deposits and radiocarbon dates from the Towosahgy State Archaeological Site located south of Charleston, Missouri, affirm the presence of two series of true paleoliquefaction features dating to within the period from about 1,500 to 950 years B.P. (Saucier 1991a). Those suggest a recurrence interval of about 468 years. More recently, Kelson et al. (1992) have found new evidence west of Reelfoot Lake that allows an interpretation of two large earthquakes during the last 650 years. Additionally, Vaughn (1991) has reported evidence of three earthquakes preceding the 1811-1812 period based on liquefaction features that have disturbed Holocene and late Pleistocene deposits in the Western Lowlands area east and southeast of Poplar Bluff, Missouri. However, the latter area is beyond what most workers would consider the limits of the New Madrid Seismic Zone; thus, their tectonic association is in question. Despite the evidence for paleoliquefaction features that has just begun to emerge, Wesnousky and Leffler (1992b) made an extensive and definitive study of exposures along tens of miles of drainage ditches in the Blytheville area (well within the seismic zone), finding no unequivocal evidence of pre-1811 liquefaction during the last 10,000 years.

An explanation of these seemingly conflicting results is beginning to emerge. It appears that the New Madrid Seismic Zone experiences every several hundred years a shock of sufficient magnitude (m_b=6.2) to cause localized liquefaction (i.e., to a maximum distance of several tens of miles from the epicenter), but nothing comparable to the 1811-1812 series has occurred during the Holocene. What still remains to be determined is to what extent the seismicity in the zone is episodic in geological terms rather than relatively uniformly distributed either spatially or temporally.

Land doming and sinking

When Fuller (1912) prepared his classic monograph on the effects of the New Madrid Earthquake(s), the tendency was to assume a tectonic origin for any significant positive or negative topographic anomaly. For example, he postulated uplift as being responsible for certain areas of Early Wisconsin Stage outwash such as Sikeston's Ridge. However, that feature has since been more satisfactorily explained by differential erosion and shifting patterns of sedimentation. Similarly, he designated as the Blytheville Dome an unusually large area of relatively high ground near that town. In consideration of a lack of evidence for uplift on the underlying suballuvial surface (Saucier 1964) and other factors, that area is now attributed to the unusual width and large number of abandoned channels (and corresponding natural levees) along the Mississippi River meander belt in that area rather than a seismic origin.

There can be no question, however, over the seismic origin of a complex uplift east of the Mississippi River in extreme northwestern Tennessee. Recognized more than 150 years ago, that area includes the Tiptonville Dome and Ridgely Ridge, two features near towns of the same names that are now considered to be part of a larger feature known as the Lake County uplift (Russ 1982). That uplift, described as an uneven and asymmetric, 30-mi-long, topographic bulge, extends northward to include the town of New Madrid and westward into Missouri to near Portageville (Plate 5). Its eastern boundary is marked by the Reelfoot fault scarp that forms the western edge of Reelfoot Lake. Topographic profiles and seismic reflection profiles indicate that both Wisconsin Stage valley train areas and Holocene meander belt deposits have been differentially uplifted by as much as 30 ft. Based on data from an exploratory trench across the scarp (Russ, Stearns, and Herd 1978), Russ (1982) concluded that the deformation responsible for the uplift occurred in spatially irregular pulses over a period of at least several thousand years. He concluded that little fault movement occurred along the scarp during the 1811-1812 earthquakes, but he acknowledges the historic accounts that indicate that uplift did take place and may also have been responsible for the rapids and reversal of riverflow that was observed just upstream from New Madrid.

From the viewpoint of assessing the role of neotectonics on river engineering, Fischer and Schumm (1993) have carefully analyzed patterns in channel planform, dimensions, and bed and water surface profiles along the present Mississippi River. They found considerable evidence to substantiate that the Lake County uplift strongly influenced the morphology and behavior of the modern river in that area. The data also revealed anomalies in the vicinity of Caruthersville (Plate 5) and Osceola, Arkansas (Plate 6), which suggest geologically recent deformation, but its relationship to the New Madrid Seismic Zone has not been established.

With regard to land sinking or tectonically-induced subsidence, the St. Francis sunklands (including Big Lake) are features widely attributed to the 1811-1812 series. As discussed in Chapter 7, there are opposing views regarding the origin of the features (Saucier 1970), with the presently most viable explanation being that they were accentuated (i.e., deepened) but not created by the seismic events. Elsewhere within and peripheral to the seismic zone, Fuller (1912) identified areas he referred to as having been either depressed or submerged during the New Madrid earthquake series. All of those are now regarded as being only areas of formerly swamp environment that occur within relict braided channels on the Wisconsin Stage valley trains. There is no need to evoke tectonics as a causal factor.

Reelfoot Lake is the feature most frequently cited (and popularized) as indicative of the geomorphic effects of the 1811-1812 earthquake series, although aspects of its specific origin remain in doubt. The lake is actually a series of irregularly shaped, interconnected water bodies, most of which occupy parts of Mississippi River abandoned channels (Russ 1982; Blythe, McCutchen, and Stearns 1975). Since oxbow lakes existed in the area prior to the earthquake series, it is inappropriate to say that the seismic events caused the lake; however, narrative accounts and a recent, detailed tree-ring analysis leave no doubt that the existing lakes were deepened and enlarged during the earthquake series (Stahle, VanArsdale, and Cleaveland 1992; Van Arsdale, Stahle, and Cleaveland 1991). Nevertheless, it is not known whether the submergence was due to the blocking of Reelfoot Creek (a natural drainage outlet for the basin) by uplift of the Tiptonville dome portion of the Lake County uplift, or the area occupied by the lake subsided, or both (Fischer and Schumm 1993). If subsidence was involved, it is not known if it represents compaction of Holocene deposits, tectonic downwarping, or faulting. In any event, about 20 ft of relative vertical displacement is indicated.

Landsliding and bank caving

Narrative accounts of the 1811-1812 earthquake series make repeated mention of landslides along the bluffs forming the eastern side of the alluvial valley, but because of the inaccessibility (due to heavy vegetation) and low population density of the area, no attempt was made to locate them and map their distribution until Fuller (1912) investigated the region. He described the principal concentration as being between approximately the Kentucky-Tennessee state line and the mouth of the Forked Deer River (Plate 5). In that area, the bluffs formed in the loess-capped upland average about 120 ft high and have a typical slope of 15 to 25 deg.

The first comprehensive investigation of landslides in that area was not accomplished until the 1980s when Jibson et al. (1988) not only mapped more than 200 individual features but also developed a classification based on the failure mechanism involved. They recognized old coherent slides (of both translational and rotational nature), earth flows, and young rotational slumps extending over a distance of about 160 mi from the mouth of the Ohio River south to approximately the Tennessee-Mississippi state line. It was concluded that only the latter type, comprising about 11 percent of the total, was due primarily to undercutting of the bluffs by river channel migration and *not* primarily due to the 1811-1812 earthquakes. Thus, seismic triggering of landslides in an inherently unstable setting was the cause of most of the failures. In a subsequent statistical analysis (Jibson and Keefer 1993), they concluded that coherent slides can occur suddenly at the epicenter of an earthquake with a magnitude of $m_b = 5.8$ while earthflows can occur in an earthquake as small as $m_b = 5.4$.

Of greater concern to river engineering than landslides is the matter of seismically induced bank caving or bank failures along the Mississippi River. Narrative accounts of the 1811-1812 series, many of which came from persons travelling on the river in flatboats, contain vivid descriptions of masses of land up to acres in extent suddenly falling (or sliding) into the river along with thousands of standing trees, creating huge waves and strong currents and sinking numerous boats. It is quite likely that cutbanks along the river failed almost continuously over a distance of tens of miles from above New Madrid to the mouth of the St. Francis River near Helena. Other narrative accounts relate to whole islands suddenly disappearing or sand bars suddenly emerging in deepwater areas as far south as Vicksburg. Within a few years following the earthquakes, newspapers contained mention of the continuing difficulties that boatsmen were having with tree- or sand-choked channels and chutes.

Most of the more reliable accounts of significant channel changes pertain to locations north of Memphis, but there is no way to establish the areal extent over which bank caving occurred. This writer firmly believes, however, that north of Memphis the amount of sediment suddenly introduced into the river channel can be measured in terms of cubic miles. It is inconceivable that that amount of sediment did not result in permanent changes in channel morphology (Walters 1975), but the evidence has thus far been undetectable in the geologic record. For example, there are no observed anomalies in point bar sequences suggesting sudden changes in accretion rates or unusual numbers of cutoffs.

Interesting questions remain to be answered as to how long it took the river to assimilate the inferred sediment load and how far downstream the effects were felt. The New Madrid Earthquakes apparently did not immediately cause radical river changes such as sudden avulsions, but the long-term effects could be consequential. Is it even possible that some of today's channel instability is still a partial response to those events? It has been speculated that prehistoric precursors of the earthquake series may have been influential in triggering distributary formation and possibly even new delta lobes.[1] If a delicate process balance existed in a geomorphic situation, could such a perturbation have been significant? That, of course, predisposes that there have been other great earthquakes in the area in Holocene times--a possibility that remains unverified.

Faults and lineaments

Terminology concerning faulting in the New Madrid Seismic Zone has evolved during the past several decades but is still confusing and inconsistent. Historically, the terms New Madrid fault or New Madrid fault zone have been widely used, but with the realization that the New Madrid Earthquake(s) were not associated with a single crustal fracture, the term New Madrid Seismic Zone became more widely used and accepted. The term New Madrid Fault Zone (and sometimes New Madrid Fault System) is still occasionally used but is defined in broader context than originally--referring to a wide zone of multiple faults trending southwestward from southern Illinois and western Kentucky.

Prior to the 1970s, faults that had offset Quaternary alluvial deposits and/or that had surface expression had been documented only from the Reelfoot Lake area (Krinitzsky 1950). Major faults were postulated on the basis of the trends of epicenters (especially microearthquakes), but confirmation from cores and geophysical surveys was lacking. During the 1970s and 1980s, extensive gravity, magnetic, and seismic surveys revealed that the seismic zone contains a complex pattern of both strike-slip and thrust faults (as interpreted from disrupted and discontinuous reflectors) that have affected the Paleozoic, Cretaceous, and Tertiary deposits of the area. Most of the faults appeared to be associated with either the rift margins or probable igneous intrusions, but the seismic data were not of sufficient resolution to determine if Quaternary deposits had been affected.

Beginning in the early 1980s, under sponsorship of the U.S. Geological Survey, several investigators began acquiring hundreds of miles of high-resolution, continuous, seismic reflection profiles using the Mini-Sosie technique. For the first time, these data permitted the recognition of faults and disturbed zones in the 150- to 2,500-ft depth range. To date, several discrete faults or fault zones have been mapped which have affected the uppermost Tertiary or Quaternary deposits. These are indicated in Figure 33 (numbers 1 through 4 plus undesignated ones) along with probable faults interpreted from unusually strong lineaments detected by aerial photography and remote

[1] Personal Communication, 1990, B. R. Winkley, Civil Engineer (Retired), U.S. Army Engineer Division, Lower Mississippi Valley, Vicksburg, MS.

sensing. There is a high probability that all of the faults have been active during the Quaternary and could experience displacement during a major earthquake.

The Reelfoot fault is the most extensively investigated fault in the seismic zone, having been explored by borings (Krinitzsky 1950), geophysical surveys, and trenching (Russ, Stearns, and Herd 1978). It has been interpreted by most workers as a normal fault bordered by a 7-mi-long scarp, the only one in the zone to have prominent topographic expression. About 8 ft of displacement is indicated at the surface, but Eocene formations at a depth of about 180 ft have been offset by about 40 ft. Rather than a single shear plane, the fault is manifest in the shallow subsurface by a several-hundred-foot-wide zone containing numerous small offsets, fractures, folds, convolute deformation, and sand-filled dikes caused by liquefaction. Based on more recent trenching (Kelson et al. 1992) and geophysical surveys (Sexton and Jones 1986), the fault has been reinterpreted as being related to extension in the crest of a monocline (Tiptonville Dome), and the monocline represents deformation above a west-dipping "blind" reverse fault that does not reach the surface. That new data can result in such strikingly different interpretations illustrates the complexity of the zone and the nondefinitive nature of the evidence.

The Crittenden County fault zone, one of the more recently discovered zones, has been described in detail by Crone (1992) and is indicative of structural conditions in the area. The rift boundary in which the zone is located is 2.5 to 5.0 mi wide and is manifest by disrupted reflectors and distinct faults that extend from the Precambrian crystalline basement through the Paleozoic formations to (on some seismic lines) the Upper Cretaceous and Tertiary rocks. The fault zone itself is interpreted as a northwest-dipping, high-angle reverse fault with an up-to-the-northwest throw. It is at least 20 mi long and has an offset of about 200 ft in the Cretaceous sequence.

Schweig and Marple (1991) recently discovered an interesting feature of probable tectonic origin called the Bootheel lineament which trends for about 85 mi through the center of the seismic zone from just east of Marked Tree to west of New Madrid (Figure 55). The feature is distinctive because of a contrast in sand-blow density on opposite sides; occasional, shallow, linear depressions; continuous or discontinuous linear bodies of sand; and the apparent truncation of some fluvial features. Trenches excavated across the lineament revealed dikes filled with liquified sand, vertically displaced blocks of the cohesive topstratum deposits, and shear zones. Lacking conclusive proof but weighing the evidence pro and con, the authors believe that the lineament is the surface expression of a major fault (strike-slip?) rather than simply an alignment of fissures due to land spreading. Seismic profiles across the lineament have subsequently revealed that it lies in a zone of complex deformation and fractured rock that is at least as young as the base of the Quaternary sequence.

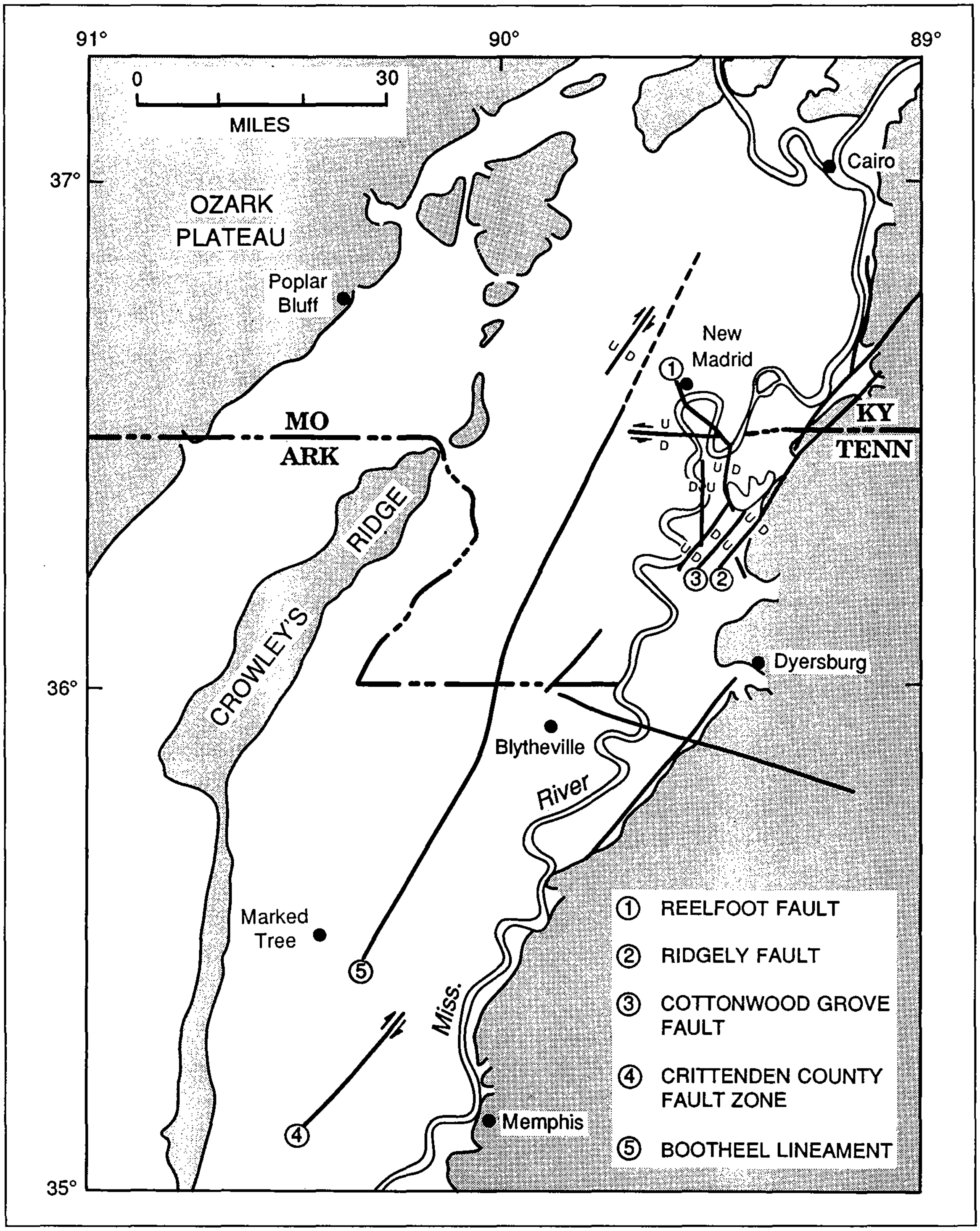

Figure 55. Known and inferred faults and major lineaments in the New Madrid Seismic Zone (compiled from Crone 1992; Heyl and McKeown 1988; Schweig, Marple, and Li 1992; and Zoback et al. 1980)

Noting that the trend of the Bootheel lineament departs significantly from that of known faults and the trends of recorded microearthquakes, Schweig (1992) has postulated that it is a relatively new tectonic element that has formed through a relatively weak zone to alleviate a restraining step in the regional stress field. If correct, that could mean that it is a fault that could be as young as a few tens of thousands to hundreds of thousands of years old.

Faulting and Regional Fracturing

Mississippi Valley area

Numerous faults have been recognized and mapped since before the turn of the century in the uplands bordering the alluvial valley, especially in the Paleozoic and Mesozoic formations of Arkansas and Missouri and in the older Cenozoic formations of southern Illinois. The more significant faults are delineated on the state geologic maps and in special reports. However, Fisk (1944) was the first to comprehensively address the matter of faulting in Quaternary deposits and did so in a characteristically highly graphic manner.

Reflecting the prevailing geologic concepts of his time, and taking advantage of the aerial photo coverage that was still a relatively new tool, he interpreted the Lower Mississippi Valley area (both alluvial valley and adjacent uplands) to be criss-crossed by a dense rectilinear pattern of northeast-southwest and northwest-southeast trending fractures. As Krinitzsky (1950) explained later, geologists of that era believed the earth's crust contained a world-wide grid pattern of faults caused by planetary-scale influences. Using classic geomorphologic criteria such as changes in soil types, topographic breaks, oriented drainage, linear lake margins, and sharp river bends, Fisk mapped hundreds of fractures that conformed to the trends of the assumed crustal pattern. Within the context of the regional fracture pattern, Fisk placed special emphasis on major northwest or northeast trending physiographic boundaries and elements such as alluvial valley margins and tributary valleys. On that basis, he designated eight major fault zones, each 200 or more miles in length and containing 10 or more individual faults. Those fault zones conformed with physiographic breaks like the Ozark Escarpment and the valleys of the Arkansas, Ouachita, and Red rivers.

Fisk assumed that the zones involved relatively old but active faults in the underlying pre-Quaternary deposits and formations (the "bedrock") which had manifestations in the Quaternary deposits, but he did not present confirming geologic evidence for their presence in the alluvium. A few years later, Krinitzsky (1950) used physiographic evidence to map the distribution of several hundred fault traces in what at that time was believed to be Recent (i.e., Holocene) deposits of the alluvial valley and deltaic plain. However, he was able to verify the presence of faults in the suballuvial surface (uppermost Tertiary deposits) by closely spaced borings in only three cases; along Big Creek southwest of Helena (Plate 7), near Old Town Lake west of Friars Point, Mississippi

(Plate 7), and west of Reelfoot Lake south of New Madrid (Plate 5). In those cases, a surface manifestation of faulting such as an escarpment was confirmed.

Since 1950, the detailed, systematic quadrangle-scale mapping of alluvial deposits of the Lower Mississippi Valley and numerous site-specific engineering geologic investigations have failed to substantiate the presence of the vast majority of the faults and fault zones indicated by Fisk and Krinitzsky. There can be no doubt that they are lineaments, but their origin cannot be attributed to fault displacements. Conversely, where high-resolution seismic surveys have been conducted, both land-based and waterborne, the results invariable indicate a much larger number of faults in the shallow Tertiary and Cretaceous formations than was suspected. However, there is seldom any surface indications of their presence, including lineaments. The lack of surface expression is a reasonable but not an unequivocal indication that the faults have not been active during the Quaternary.

Perhaps the most conspicuous lineament in the entire alluvial valley area occurs near the western side of Macon Ridge in northeastern Louisiana and southeastern Arkansas. Consisting of three slightly offset segments in an *en echelon* pattern, the lineament extends in a north-northeasterly direction for about 90 mi between the community of Riverton, Louisiana, on the Ouachita River, through the community of Start, Louisiana, along and slightly west of the Boeuf River, and through the community of Chicot, Arkansas (Plates 8 and 9). Despite sharp soil tonal changes, aligned drainage, and topographic breaks, there is no subsurface evidence to substantiate the presence of a fault.

If a generalization can be made, it is that there is very little substantiation of faults previously mapped on the basis of physiographic evidence while at the same time, there are likely surface manifestations of faults that have previously not been identified or predicted. In the first instance, the only exceptions may be where physiographic evidence (e.g., drainage-basin asymmetry) coincides with geological evidence such as seismicity patterns (Cox 1994). With regard to Crowley's Ridge, Fisk (1944) believed that the general configuration of the ridge suggested the presence of faulting, but it was not until the high-resolution seismic reflection profiling of the last several years (VanArsdale 1992; VanArsdale et al. 1990, 1992) that it has become evident that the configuration of the ridge may be dominantly fault controlled. Those geophysical data indicate that faults are present to within 60 ft of the surface (the upper limits of the technique) on both sides of the ridge from at least 12 mi north to 12 mi south of Jonesboro (Plate 6). At the time of the writing of this synthesis, geomorphic studies are in progress to determine if the faults are actually present at the surface. Fischer and Schumm (1993) have recently gone even further by interpreting ridge drainage patterns and topography as indicating the presence of separate fault blocks within the ridge, but that has not been verified by subsurface data. In none of the studies has there been any evidence discerned of fault movements as recent as the Holocene.

Consideration was given to providing in this synthesis a map of verified faults affecting the Quaternary deposits, but excluding the New Madrid Seismic Zone (Figure 55) and the deltaic plain area, there are too few to justify that approach. It is sufficient to conclude that outside of the areas mentioned, faults affecting Quaternary deposits are rare, but not so rare that the possibility can be discounted in any site-specific investigation for an engineering project. Physiographic evidence, especially the mapping of lineaments, has been shown to be nondefinitive at best.

Gulf Coast area

In the consideration of the comprehensive scope of Fisk (1944) and in view of the number of geologists that were involved in its preparation, it is remarkable that there are so few internal inconsistencies and contradictions. However, a few have been observed over the years, perhaps the most important being with regard to the conflict between the highly emphasized rectilinear regional fracture pattern and the Baton Rouge Fault Zone. As discussed in Chapter 6, that fault zone is the northernmost of a series of 10 composed of deep-seated, down-to-the-coast, normal faults that collectively constitute the South Louisiana Growth Faults. Trending essentially east-west, the well-documented growth fault zones preclude the possibility that the same area exhibits a pattern trending at approximately 45 deg.

Of all of the growth fault zones, the Baton Rouge Fault Zone is one of the best known both in the subsurface and at the surface (Durham and Peeples 1956). It has been extensively investigated because of its importance in petroleum production, groundwater occurrence and movement, and urban and engineering geology. Next to the Reelfoot Fault, it is the next most significant in the Lower Mississippi Valley area in terms of the extent and recentness of Quaternary displacements.

As traditionally mapped (Snead and McCulloh 1984), the Baton Rouge Fault Zone consists of several discrete faults that trend eastward from the southern part of the city of Baton Rouge, across the Amite River, through the Ponchatoula, Louisiana, area, to near Madisonville, Louisiana, and thence south-southeastward to near Slidell, Louisiana (Plates 11 and 12). Near Baton Rouge, the fault zone has been traced to a depth of at least 20,000 ft and has as much as 450 ft of displacement. It has offset buried Late Pleistocene strata by as much as 100 ft and deposits of similar age as much as 35 ft at the surface. Several authors have commented on its geomorphic effects which include scarps, offset drainage lines, oriented drainage, soils changes, and swampy depressions.

In contrast to movements along the Reelfoot Fault which have been episodic and triggered by seismic events, movement along the Baton Rouge Fault Zone has been in the form of progressive, nonseismic creep. Vertical offsets in roads, bridges, and drainage ditches and damage to structures attest that movement has continued from the Pleistocene (and earlier) into the Holocene and into late historic times.

Recently, Lopez (1991) has presented compelling geophysical and other evidence that the surface manifestation of the portion of the fault zone east of about Ponchatoula has been mistaken for depositional features (beach ridges?) and that the fault zone lies beneath rather than north of Lake Pontchartrain. There is evidence from borings that the Wisconsin Stage deposits beneath the lake have been offset along the fault zone (Kolb, Smith, and Silva 1975), and Lopez (1991) indicates that the Lake Pontchartrain Causeway across the center of the lake and the 62-year-old U.S. Highway 11 bridge across the eastern end of the lake have been offset by as much as 5 to 6 in. Kolb, Smith, and Silva (1975) have also pointed out that high-resolution, acoustic, subbottom profiling has indicated fault displacements in the Wisconsin Stage deposits at other locations beneath the lake.

West of Baton Rouge, there is perhaps slight indication of the effects of the Baton Rouge Fault Zone in Holocene backswamp drainage patterns,[1] but it is not manifest in the suballuvial surface. Perhaps this is only because there are extremely few borings in the area (Plates 15 and 24). However, as discussed in Chapter 5, the contouring of the suballuvial surface does indicate a general downvalley steepening of the main Mississippi River entrenchment about where the South Louisiana Growth Fault zone begins. Specifically, there are conspicuous east-west trending depressions, in one case over 400 ft deep, that coincide with the northernmost growth faults that trend through the towns of Baker and Zachary, Louisiana (Plate 11).

There is no convincing evidence to indicate that Holocene and historic period displacements comparable to those described above have not occurred along the growth fault zones located farther to the south. However, probably because of the masking effect of the considerably greater thickness of poorly consolidated Holocene deltaic deposits closer to the coast (which warp rather than shear), surface evidence has not been discerned.

The geologic literature contains only scattered mention of other possible fault displacements in shallow deltaic deposits. For example, Fisk (1944) interpreted a line of widely spaced borings to indicate offset of both Pleistocene and Holocene deposits along his postulated Lake Borgne Fault Zone east of New Orleans. However, more recent studies with vastly larger amounts of data (Saucier et al. 1984, Saucier 1991b) have failed to verify the presence of faulting and instead attribute the observed stratigraphic evidence to normal deposition processes. Another investigation of possible faulting focused on a slight earth tremor and associated mile-long surface displacement and cracking observed in 1943 near Vacherie, Louisiana (Plate 12) (Fisk 1943, 1944). Borings at the site revealed evidence in both Pleistocene and Holocene

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[1] Personal Communication, 1993, R. McCulloh, Geologist, Louisiana Geological Survey, Baton Rouge.

deposits for recurrent movement along a fault; however, there was probably not one in the Red River Fault Zone as Fisk postulated. Rather, the fault is probably one of local extent associated with the diapiric movement of the Hester salt dome. This writer believes that many undetected salt dome-related faults exist elsewhere in the deltaic plain area and are capable of causing surface displacement.

9 Special Engineering Considerations

Groundwater Occurrence

The alluvial aquifer of the Lower Mississippi Valley is one of the largest and most exploited sources of shallow, fresh water in the United States. In essence, it is the largely uninterrupted mass of coarse-grained substratum deposits that overlies the eroded suballuvial surface and extends from valley wall to valley wall (Plate 16). It includes the lower, coarse-grained portions of the Holocene point bar environment, glacial outwash underlying the Early and Late Wisconsin Stage valley trains (the Pleistocene substratum as described in Chapter 6), the substratum deposits underlying the Prairie Complex, and probable remnants of earlier Quaternary valley-fill sequences. Longitudinally, the aquifer extends from the head of the alluvial valley downstream to and beyond the deltaic plain where it includes the deposits that fill the entrenched valley. However, below the latitude of Baton Rouge, water in the aquifer becomes brackish or saline and is of limited utilization.

Having a nominal thickness of about 125 ft, the aquifer has an estimated volume of about 790 cu mi. About 30 percent of that amount consists of fine to medium sands (the upper part) while the remainder consists of coarse sands and gravels. Almost everywhere the sequence coarsens downward but not in a uniform manner. Between the head of the valley and the latitude of Vicksburg, the amount of fresh water stored in the aquifer has been estimated at slightly more than 120 trillion gal (Boswell, Cushing, and Hosman 1968).

The aquifer is confined where the alluvial topstratum is thick and continuous, such as in backswamp areas, but it is essentially an open hydrologic system with relative rapid recharge and discharge, both natural and artificial. As indicated in the only comprehensive geological treatment of the entire aquifer to be published (Krinitzsky and Wire 1964) and in certain regional geohydrological overviews such as Sumner and Wasson (1990), recharge occurs primarily through infiltration in point bar and valley train areas where the topstratum is thin, and along major streams during high stages; and secondarily from underflow from suballuvial aquifers of Tertiary and Cretaceous age. Discharge occurs mainly as a result of contributions to stream base flow during low stages, and withdrawal by extensive pumpage. Except near well fields, water levels in the aquifer are generally less than 30 ft below the land surface, and seasonal fluctuations typically are about 20 ft.

Water from the alluvial aquifer is extensively and locally intensively used for urban and industrial purposes and, contrary to what most would expect in a humid climate, for irrigation. Some of the more severe aquifer depletion historically has occurred in areas of rice cultivation and more recently where catfish farming is being practiced. Irrigation wells, numbering in the thousands, typically have yields of several hundred gallons per minute (gpm) each, and high capacity wells in urban areas sometimes exceed 7,000 gpm. Wells average about 150 ft deep and generally are set to produce from just above the suballuvial surface where the deposits are coarsest. In a recent study in northeastern Louisiana (Whitfield 1975), which is typical of much of the alluvial valley area, aquifer tests indicated transmissivity values ranging from 13,000 to 45,000 sq ft per day, with the hydraulic conductivity ranging from 130 to 530 ft per day. In other parts of the alluvial valley area, transmissivity values as high as 65,000 sq ft per day have been observed (Broom and Lyford 1982). As long ago as 1965 (the latest date for which such estimates are available), it was estimated that withdrawals from the alluvial aquifer (north of Vicksburg) averaged about 1,430 million gal per day, or 1,600,000 acre-feet (Boswell, Cushing, and Hosman 1968). No doubt present withdrawal rates are considerably higher.

Because of the economic importance of groundwater and concerns about its use and depletion, there have been extensive studies of each of the major basin areas by the U.S. Geological Survey in cooperation with state agencies. These have been supplemented with frequent maps and compilations of water-level trends, and data on piezometric surfaces in monitored wells are readily available from several sources.

Nearly all groundwater in the alluvial aquifer is hard to very hard with a high concentration of iron and manganese; calcium, magnesium, and bicarbonate ions; and dissolved solids. Specific amounts vary widely because of the character of the surface soils, the influent source, and the character of the water in the suballuvial aquifers. In virtually all areas, however, the concentrations are high enough that treatment is necessary for the water to be satisfactory for domestic, municipal, and many industrial uses.

Historically, contamination of the alluvial aquifer has not been a significant problem and has mainly involved the localized presence of salty water. High chloride concentrations are typically encountered near the base of the aquifer. They are believed to originate mainly from discharge or underflow from certain Tertiary suballuvial aquifers, but some are known to be attributable to influent chloride-rich water from certain streams like the Arkansas River. In other instances, natural causes are not viable explanations, and chloride concentrations are believed to originate from petroleum well blowouts, saltwater disposal pits, and especially leaky abandoned wells. Contamination from such sources tends to form sharply bounded plumes that follow topographic lows on the suballuvial surface, braided paleochannels, or regional gradients created by pumpage.

Within the past two decades, attention has been directed to contamination of the aquifer by anthropogenic chemicals. High levels of pollution apparently are restricted to localized areas near industrial complexes, but agriculture is adversely affecting water quality in other areas at an increasing rate. Herbicides and pesticides generally are adsorbed by fine-grained deposits of the topstratum and do not reach the aquifer by infiltration, but such is not the case with nitrates. On the other hand, moderate levels of herbicides and pesticides are present in many surface streams, including the Mississippi River; thus, entry into the aquifer by direct recharge during high stream stages is a definite possibility.

Water quality of the alluvial aquifer is a matter of engineering as well as environmental concern, particularly with regard to sodium and potassium salts, sulfate and chloride ions, organic acids, pH, and iron. As discussed by Krinitzsky and Wire (1964), particular considerations in the use of groundwater include the setting and curing of cement; aggregate reaction; cement staining; the corrosion of metal such as pilings, steel reinforcing rods in concrete (rebars), and relief well screens; and the clogging of relief and recharge wells.

The state of the art of Quaternary geology thus far has not allowed the correlation of regional variations in aquifer yield and water quality with the ages and specific modes of origin of the deposits, but the possibility exists and should be pursued. With time and as use of the alluvial aquifer increases, it will be necessary to further explore the nature, extent, and hydraulic significance of lithofacies in general and specifically features such as paleosols and relict channels.

Most interest in the Lower Mississippi Valley area has been on groundwater availability and use, but there is one situation in which excess groundwater (in the form of high water levels) has been a chronic problem. That is the matter of levee underseepage. Although not unique to the area, underseepage has been of special engineering concern because artificial levees increased the flood stages and most levees overlie point bar deposits of the modern (Stage 1) meander belts of the Mississippi and Arkansas rivers. No levee failures or crevasses directly attributable to underseepage have occurred since 1913, but this process definitely poses a threat to the integrity of the area's flood control system if not controlled.

The occurrence of subsurface piping and the formation of sand boils as a result of excess hydrostatic pressure and seepage through deep pervious strata during flood stages have been extensively investigated (Mansur 1956). It has been found that seepage flow can be estimated on the basis of the distribution of environments of deposition (i.e., the geological context) in conjunction with seepage formulas and/or piezometric data. Control of underseepage has been achieved at problem sites by various combinations of the careful placement of ditches and borrow pits and the construction of impervious riverside blankets, relief wells, and landside seepage berms.

Mass Movements

Bank caving

Underseepage can be regarded as the nemesis of flood control levees, but bank caving plays that role in regard to revetments and bank stabilization efforts. Fisk (1947), in what some regard as his "finest hour," eloquently characterized the effect of fine-grained alluvial deposits on Mississippi River activity, with special attention to the causes and patterns of bank caving. He recognized that bank caving is an integral component of the process of river meandering and is controlled by variations in the relative thicknesses and composition of the topstratum and substratum components of the alluvial sequence. He explained that the primary cause of bank caving is scouring in the thalweg of the river channel at the toe of the concave bank, but that the actual manifestation of failures is dependent on the nature of the subaqueous profile of the sedimentary sequence. For example, failures in thick backswamp or abandoned channel clays are significantly different from those in noncohesive point bar sequences.

Fisk identified several types of bank failure mechanisms, but Turnbull, Krinitzsky, and Weaver (1966) were the first to describe and illustrate the mechanisms and relate them to the thickness of the fine-grained topstratum in various alluvial and deltaic environments of deposition (Figure 56). In each case, a scour pool forms and the upper bank deposits are oversteepened and destabilized. Where noncohesive deposits are present (Figure 56A through 56D), subaqueous failures of various sizes by either flow or shear mechanisms result. The former involve either the partial or complete liquefaction and flowage of sandy deposits. In turn, the subaqueous failures trigger upper bank failures again by either flow or shear. As illustrated in Figure 56, the upper bank failures can vary widely in type, size, and frequency within short distances depending on the nature of the deposits. Where cohesive deposits extend to the maximum extent of channel scour, such as in lower portion of the deltaic plain (Figure 56E), only shear failures are possible, but they can be of significant size (Stanley, Krinitzsky, and Compton 1966).

Being largely dependent on the formation of scour pools, bank caving therefore occurs dominantly during seasonal high river stages. Under natural conditions, the result was a rather constant retrogression of bank lines by raveling or the coalescence of innumerable small slumps. It has been estimated that cumulatively, the process involved about 1,000,000 cu yd of material per year per mile of river (Winkley 1970). However, there is documentation of numerous failures along the Mississippi River involving as much as 10 to 20 acres of land wherein banks receded hundreds of feet within a matter of hours or a few days (Carey 1969).

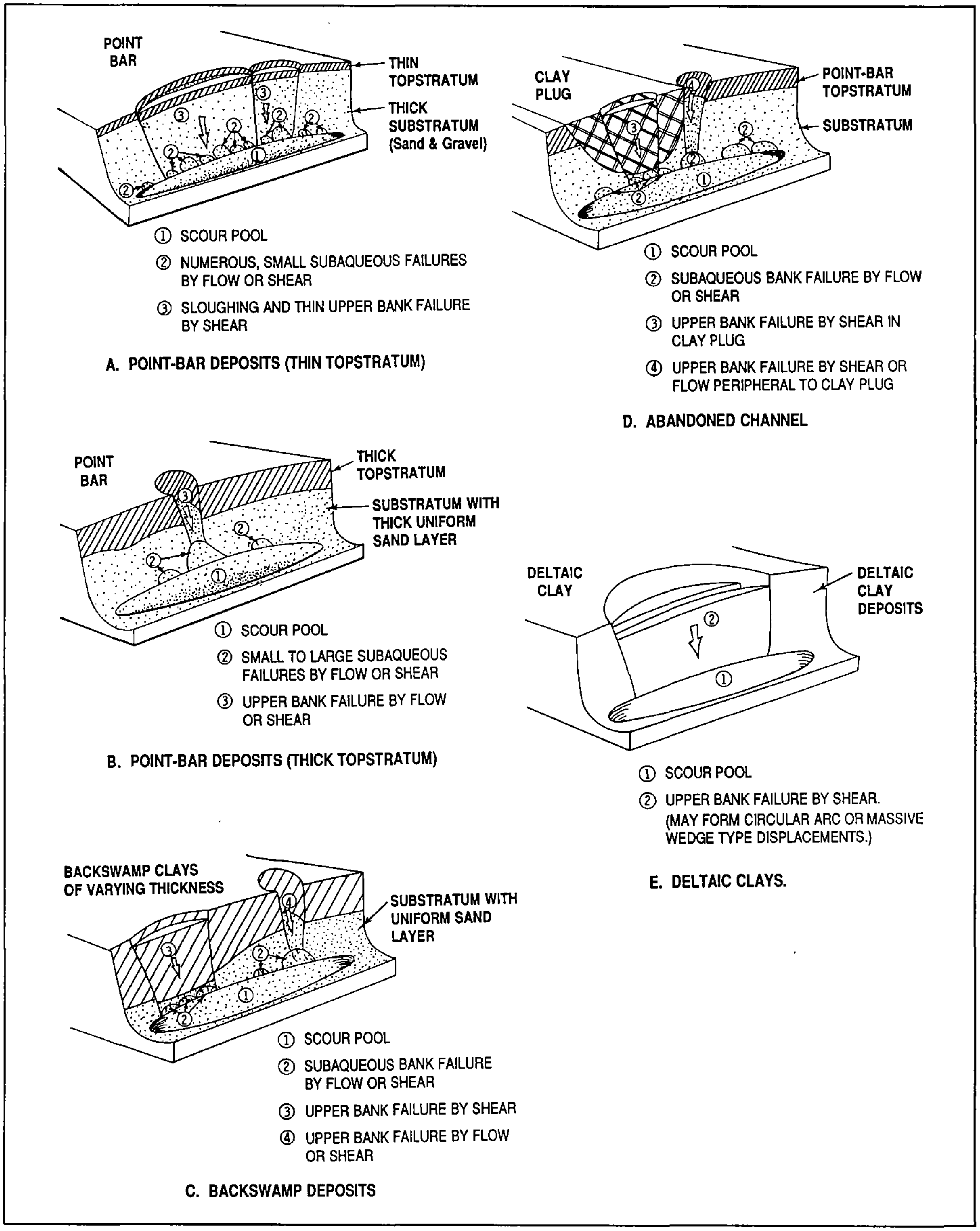

Figure 56. Typical bank failure mechanisms in various alluvial and deltaic environments of deposition (from Turnbull, Krinitzsky, and Weaver 1966)

Artificial bank stabilization measures have become widespread along the Mississippi River below Cairo since their introduction in the late 1870s, and they have advanced in sophistication from crude willow mattresses to mass-produced articulated concrete revetments used in conjunction with upper bank paving. Cumulatively, almost precisely 1,000 mi of revetment have been constructed along the 953 river miles between Cairo and Head of Passes (Plate 14). Some of the revetments are inoperable, and others have been stranded by river channel changes. The latter technique has dramatically reduced the incidence of bank failures but has not prevented them, including some of extraordinary size that cannot be prevented by conventional revetments. These failures include major liquefaction failures in point bar deposits. Because of a series of rather spectacular failures of that type in the 1940s, WES conducted extensive field exploration and laboratory experimentation during the 1950s. Those efforts led to the development and eventual modification and verification (by the 1980s) of empirical criteria for predicting susceptibility of deposits to flow failure (Torrey, Dunbar, and Peterson 1988). It has been found that the susceptibility of a given bank to large flow failures is dependent on the relative thickness of the point bar topstratum and the zone of fine sand (designated Zone A) which characterizes the upper part of the point bar substratum. Flow failures have never been known to extend into the lower part of the point bar substratum which consists of coarse sands and gravels. In contrast, a ratio of topstratum thickness to Zone A thickness of 0.85 or less and a Zone A thickness of 20 ft or more indicates an unstable situation susceptible to failure.

Historically, concern over large flow failures has focused on the alluvial valley between Cairo and Baton Rouge because that area is where most failures have been observed and investigated. However, examination of borings and their geological context (Kolb 1962, 1963) revealed that conditions conducive to flow failures also existed below Baton Rouge, causing a shift in attention to that area. That concern was justified when during the severe flood of 1973, four large flow failures occurred in that area, requiring levee setbacks to protect the integrity of the flood control system.

In a geological context, direct evidence of bank caving is sometimes preserved in the stratigraphic record to a greater extent than one might imagine. Noncohesive sediments, such as Zone A sands, are not preserved since the deposits are rapidly entrained in the riverflow and have become part of the bed load. However, blocks of cohesive sediment, which fall vertically or translate laterally into the channel when failures occur, can be preserved in those cases where a change in channel planform causes an abrupt interruption in meandering (such as a cutoff), or where a course is abandoned. For example, examination in the late 1960s of the deep and extensive excavation for the Jonesville Lock & Dam on the Black River (Plate 10) revealed the presence of numerous displaced blocks of backswamp clays in the relict Stage 2 channel in the Walnut Bayou Meander Belt. The blocks, measuring tens of feet in width, lay near the base of the western side of the channel in a chaotic pattern and were rotated up to 30 deg from their original orientation. Rotated slump blocks of similar material have also been interpreted in closely spaced undisturbed borings in Prairie Complex deposits on the basis of inclined strata. Where such deposits are disintegrated by riverflow, they are the principal source of the large numbers of clay balls that are found in basal portions of abandoned channel and course fillings and in point bar deposits.

Landsliding

Failures of the bluffs along the eastern side of the alluvial valley have been a frequent and continuing geomorphic process from nearly as far south as Baton Rouge northward virtually without interruption to the head of the valley at Cairo. As discussed in Chapter 8, the New Madrid earthquake series was a major triggering mechanism for unstable bluffs north of Memphis. However, the underlying cause of the instability has been the undercutting and over-steepening of the bluffs by the lateral migration of the Mississippi River and other large streams, both during the Holocene while in a meandering regime and during the Pleistocene while in a braided regime. Although the Mississippi River has impinged against its valley walls at one time or another during the Quaternary in most areas, it is only along the eastern bluff line and possibly along Crowley's Ridge that the geological conditions are conducive to widespread failures.

The areas of highest instability occur where the river is actively cutting into the base of the bluffs. Large landslides were noted as early as the mid 19th century by Lyell (1847), and Russell (1934) described a large failure at Fort Adams, Mississippi (Plate 10), that affected about 97 acres of the uplands. Others involving tens of acres of land each have been observed at Natchez, Vicksburg, Yazoo City, Mississippi, Memphis, and in other populated areas where such mass movements impose hazards to property and structures.

Individual landslides in those areas have received attention in geotechnical investigations of such structures as highways and bridges, but the only extensive process-oriented geomorphological and geological investigation has been conducted at Natchez. There, extensive field mapping and subsurface studies were made as part of a project aimed at identifying bluff stabilization measures that could be employed to reduce recurring serious damage and destruction of historically significant buildings and other resources (Church and Hunt 1984).

At Natchez, the geological sequence, from the top of the bluff to the elevation of the bed of the river, is comparable to that all along the eastern valley wall, i.e., about 90 ft of loess overlying about 150 ft of mostly coarse-grained Upland Complex deposits (locally designated the Natchez and Citronelle formations) and thence hard clays or shales of the Hattiesburg formation (Miocene). It was determined that seven separate mass wasting processes have affected all or part of the sedimentary sequence, including rotational slumps, soil slides, soil fall, soil creep, solifluction, mudflow, and surface wash. The seven are interrelated and actually part of a continuum of processes, and any individual failure may be characterized by a combination of several processes.

Each occurs dependent on the steepness of the slope and the amount of water available, but the seven often are gradational from one to another.

In a typical scenario, a large-scale slump occurs because of undercutting of the bluffs by river scour. Medium-scale slumping occurs as a secondary consequence of the large-scale slump, which may in turn cause small-scale slumping. Accumulation of slide debris on lower slump masses causes reactivation of old slumps. Soil falls and soil slides become active on steep slump scarp slopes, and mudflows and solifluction then gradually transport the slide debris farther downslope. Surface wash and soil creep will occur throughout the lower portions of the landslide mass.

In areas where large rivers are no longer actively undercutting the bluffs, a different suite of processes will be active. Slope failure processes triggered by large-scale slumps will be diminished in their importance or will be inactive. Erosion by surface wash and soil creep will become dominant erosional processes, but these may result in gully development which, in turn, will promote soil fall and soil slides and possible small-scale slumps in the loess.

Based on the studies at Natchez, it was postulated that time scales for the recurrence of large-scale slumps can probably be measured in terms of hundreds or thousands of years, whereas medium-scale slumps can probably be measured in terms of tens or hundreds of years. Small-scale slumps, soil falls, soil slides, solifluction, and mudflows are relatively frequent phenomena, especially where the bluffs have been disturbed and the vegetative cover removed.

Practical experience with landslide phenomena in the Vicksburg area has demonstrated that they are ubiquitous within at least several hundred feet of the bluff line, often becoming discernible for the first time only in road cuts or excavations. This experience suggests that landsliding in all parts of the eastern valley wall is a much more consequential and widespread engineering concern than previously suspected. Instability may remain latent for thousands of years after active river undercutting ceases and large-scale slumps have temporarily stabilized. Any activity, such as removal of material from the base of a slope or alteration of the groundwater regime, may cause a renewal of slide movement, often along deep-seated failure planes. Moreover, soil falls and slides and solifluction masses typically represent disturbed loess deposits that are highly susceptible to erosion when vegetation is removed.

Experience has also shown that the presence of competent deposits of Tertiary age (such as hard clays or even limestones) near the base of the bluffs is not necessarily an indication of stability. Deposits of that type have been known to have been undercut or destabilized and involved in large-scale slumps in several locations other than at Natchez. Materials of that type have been displaced tens of feet vertically and horizontally as a consequence of slumping.

Avoidance of landsliding is best achieved by preventing the migration of a stream into the bluff line and the consequent oversteeping of slopes.

Stabilization of existing landslides is obviously dependent on the process or processes involved, but it almost always includes reducing or eliminating the infiltration of water (from both natural and artificial sources) and controlling runoff. In turn, the careful management of vegetation as well as possible artificial drainage channels, berms, and slope protection is usually involved. Under some circumstances, earth fills and relief wells may be required.

River Meandering and Long-Term Stability

During the last several decades and as part of the extensive efforts to provide effective flood control and navigation improvements in the Lower Mississippi Valley, the state of the art has advanced from the engineering of individual projects to engineering of the river in a holistic sense and has witnessed the advent of potamology (defined as the scientific study of rivers). For more than two decades, Winkley (1977) has been an outspoken leader in both disciplines, arguing that in addition to a knowledge of hydrologic, hydraulic, and geometric variables, an understanding of the river *system* can be achieved only with a knowledge and appreciation of *time* and *geologic* variables. To the fullest extent, he believes that a study of the past is critical to an understanding of the present and the future. He has argued that the success or failure of man's changes in the river regime can only be measured against the *natural* trend of the system over geologic and historic time frames, since it is constantly changing rather than being static. Moreover, he has also stressed that man-induced changes often set in motion responses that propagate long distances and last for long periods of time.

Winkley (1970) proposed a hierarchical classification of river variables which, from most independent to the most dependent, includes time, geology, climate, vegetation, relief, valley dimensions, discharge and sediment, channel morphology, and flow characteristics. The interaction of the variables has produced a quasi-state of equilibrium in which the river may be regarded as graded or "poised." Winkley believes that the poised condition has developed only during the last 400 to 500 years during which time most of the variables have been independent and steady. Prior to that time, most were changing (unsteady) because of the transition from glacial to postglacial conditions. Subsequently, during the past 30 years (from a publication date of 1970), manmade structures have upset the natural equilibrium, possibly causing certain variables to change.

Many workers question whether any major river system in the world has ever been truly graded or poised in the classic sense at any time during the Quaternary (Knox 1975), with or without the influences of man; hence, the geomorphic concept of dynamic metastable equilibrium developed. This writer believes that a graded or poised condition existed in the late Holocene, but strictly only in a *relative* sense. As discussed in the following paragraphs, changes definitely occurred in several variables, but unfortunately it has not been possible to identify the responses or even to determine if one change has been compensated by another. Some of the changes are well known and have been debated at length, while others have come to light only during the compilation of this synthesis.

Of the more than 30 individual geologic, hydrologic, and hydraulic variables identified by Winkley and Robbins (1970), they considered that certain ones could only be considered as constants during the time frame of the Holocene, not that changes have not occurred, but rather because data were too limited to allow meaningful discussion. Included in this category are total river discharge, temperature, vegetal cover, longitudinal profile, channel cross-sectional shape, bed roughness, turbulence, groundwater interaction, and others. However, at least brief subjective discussion is in order for others like valley slope, annual hydrograph, bed and bank materials (including the suballuvial formations), total sediment load, and bed-load characteristics.

Excluding regional neotectonic influences, overall valley slope has not remained constant even during the late Holocene (i.e., approximately the last 4,000 years) even though sea level is regarded as having been essentially stationary. Just a few feet of possible rise during that time frame has been significant in the lower part of the valley, but of equal importance, the occurrence of diversions and the formation of new meander belts have resulted in a net vertical aggradation of the valley. However, the impact on the river profile has not been uniform in space or constant in time. The low water profile of the present river is not necessarily optimal or "in grade"; rather it reflects only what has been possible for the river to achieve considering the overall physiography of the valley as determined by the cumulative pattern of meander belts. Variations in the profile must be evaluated first in the context of the total sedimentary history before other variables like geologic structure are considered.

In geological time frames, it is apparent that the rate of vertical aggradation in the alluvial valley that has occurred because of overbank sedimentation during the Holocene cannot continue indefinitely (assuming no *major* changes in sea level). It is interesting to note that the total vertical aggradation that took place during earlier interglacial stages, as evidenced by the thickness of the Prairie and Intermediate complex sedimentary sequences, is quite comparable to that of the Holocene sequence despite much longer periods of time involved in the former. Some self-regulating mechanism must be present.

That mechanism is believed to involve the ability of valley (in terms of size) to provide an opportunity for meander belt development in new locations, i.e., the concept of accommodation space as described by Shanley and McCabe (1994). During the Holocene, the alluvial valley has experienced a major diversion affecting a significant part of its area on average about once every 1,700 years. In consideration of the distribution of ridges of Wisconsin-stage outwash and older meander belts, there are not enough lowlands available (that could provide the gradient advantage necessary for a diversion) to allow more than a few more to form. Instead, either the frequency of diversions will decrease, or there will be an increasing tendency over a period of several thousand years for the river to reoccupy abandoned meander belts rather than create new ones. Gradually, this will affect the overall character of the sedimentary sequence of the valley as explained in the following paragraph.

As discussed in Chapter 5, the river in a relatively new and enlarging meander belt is characterized by a relatively large percentage of cohesive sediments (e.g., backswamp deposits) along its banks, and that percentage decreases as meandering progresses. This affects the amount of sand available for point bar development and the ability of the river to transport that load progressively downstream to the Gulf. It is true that as a meander belt matures, the number of clay plugs (abandoned channels) increases and these act to restrict free meandering but not nearly to the same extent as bordering backswamp deposits. Further, with continued meandering within a given meander belt or upon reoccupation of an older meander belt, "high spots" on the suballuvial surface are progressively removed by lateral planation or vertical scouring and become less and less a factor in inhibiting the meandering process. These factors have led this writer to believe that with increasing duration of an interglacial stage, progressively more sediment is transported through the system rather than stored within the system, thereby leading to a decline in the rate of overall vertical aggradation. Obviously, the onset of glacial or stadial conditions causes a major interruption.

The evolution of a meander belt through a life cycle and that of an alluvial sequence through an interglacial stage involve thousands of years, but it is not an irrelevant consideration in terms of engineering projects. It is important to recognize the status of the historic period in a *geological context* in order to understand and explain the regime trends that are presently taking place. Kesel, Yodis, and McCraw (1992) have recently underscored the significance of sediment storage within the lower Mississippi River by calculating the probable budget for the early historic period prior to most human modifications. They concluded that only about 25 percent of the total sediment load went into relatively long-term overbank storage during the 31-year period of study, but the longer-term *natural trend* needed to establish a meaningful baseline for evaluating anthropogenic changes has not been established.

With shorter-term variables (centuries rather than millennia), it is important to recognize that the annual hydrograph of the modern regulated river is considerably different from that of early historic and prehistoric times. Due to extensive land clearing for agriculture and timber production, drainage improvement, and general urbanization in all parts of the drainage basin, runoff is much faster, leading to briefer but higher flood stages. When the drainage basin and the alluvial plain were basically forested and undeveloped, runoff was significantly retarded and there was extensive temporary overbank storage of flood flows. Although a net effect of river channel stabilization and navigation improvements (e.g., revetments, dikes, and dredging) has been an increase in stream velocity, the increased discharge capability of the channel has not been adequate to compensate for the increases in runoff rates. Tributary reservoirs are capable of mitigating but not negating the changed runoff regime.

In the matter of relatively long-term changes in the total sediment load of the river, it has been concluded that there are no generally accepted quantitative relations describing the response of river regimes to changes in sediment loads (Mueller and Dardeau 1990). Nevertheless, it has been firmly established that reservoirs on major tributaries (mainly in the Missouri and Arkansas river basins) and the stabilization of caving banks have caused a significant recent downward trend in the measured suspended sediment load of the Mississippi River (Dardeau and Causey 1990, Tuttle and Pinner 1982). Kesel, Yodis, and McCraw (1992) believe that there has been almost a 70 percent decrease in the suspended load of the river since 1850 and that there has probably been a corresponding decrease in the quantity of the bed load.

Changes in total sediment load of that magnitude are dramatic and certainly have had geomorphic consequences such as the accelerated loss of wetlands (interdistributary marsh) in coastal Louisiana. However, irrespective of how well documented the sediment load reductions have been for the period prior to major river regulation, they do not represent *natural* conditions or necessarily a geological trend. It is well established that extensive initial land clearing for agriculture in the Lower Mississippi Valley area and environs, especially in the first several decades of the 19th century, caused extensive and intensive gullying and erosion. Consequently, it is highly probable that the 1850 baseline against which sediment loads have been calculated represented an anomalous situation in a long-term perspective. The difference between the present sediment load of the river and that of prehistoric times (essentially natural conditions) may be considerably less than is implied: the real difference may only approximate that fraction directly due to reduced bank caving.

As has been discussed in Chapter 5, the modern (Stage 1) Mississippi River meander belt maintains a rather constant width below Vicksburg to just north of Baton Rouge at which point it begins to progressively narrow. The narrowing has been attributed primarily to the more erosion-resistant nature of the bed and bank materials (cohesive deltaic deposits), but it is apparent that time is also an important variable. Under natural conditions, progressively more sand would be transported downstream by the process of "trading" (Tuttle and Pinner 1982). Point bars would develop more fully, and cohesive bank deposits would eventually fail at a faster rate. Over a period of perhaps a few thousand years, and barring a major diversion, the meander belt would have a relatively more constant width all the way to the Gulf. It is well known that hydraulically the river has the capacity to transport into the Gulf a greater quantity of bed load of coarser particle size than has thus far been furnished from upstream.

Natural conditions no longer prevail, and the amount of sediment (both suspended load and bed load) available to be transported downstream has diminished because of the factors described earlier. While the total load has diminished, there have also been changes in its character. Because of revetments and dikes and through the action of hydraulic sorting, the relatively finer grained sediments are being progressively winnowed from the bed material, and the coarser sizes are being moved farther and farther downstream (Winkley 1970). This would suggest an eventual trend toward greater channel instability and an enhanced meandering tendency along the lower river (in human rather than geological time frames), but a compensating process may be occurring. That process is sand and gravel mining in which the largest grain-size fraction of the bed load is selectively removed by hydraulic dredging for use in the construction industry. No estimates are available of quantities involved, but they have probably reached significant levels in terms of river behavior and are increasing.

In overview and in terms of river engineering, it is important to consider that changes in most variables that occurred under natural conditions did so in the context of a whole meander belt, and eventually the entire floodplain and alluvial and deltaic sequences. Hence, there was an appreciable buffering effect. However, with artificial river regulation, the river is effectively isolated from the floodplain and alluvial/deltaic sequence, and all changes must be accommodated in a narrow, several-mile-wide corridor. Anthropogenic changes in the variables have not introduced any new geomorphic processes, but they have dramatically increased the magnitude of the processes and the rates at which they have operated. Moreover, the changes are affecting a landscape relatively more sensitive to change and one in which the assimilative capacity of the system has been reduced.

10 Summary and Future Research Needs

Summary

Introduction

This summary is a selective treatment of certain subject matter in the synthesis with emphasis on major findings, conclusions, and limitations in knowledge. It focuses on geomorphic processes and controls and chronostratigraphy and does not attempt a summary of the systematic descriptions of the lithology, soils, and geotechnical characteristics of various deposits.

The impetus for the present synthesis has been the 50 years of advancements in geosciences knowledge that have occurred since publication in 1944 of Harold Fisk's classic monograph, *Geological Investigation of the Alluvial Valley of the Lower Mississippi River.* Some of Fisk's contributions, such as the influence of alluvial deposits on fluvial processes and the dynamics of stream diversions, have required little modification. However, other aspects of his work, such as concepts of terrace formation and valley entrenchment, are outdated. Major revisions have been made and still others are needed. In particular, his chronology of valley events and stream channel changes is largely invalid, and a tentative new model has been proposed.

The present two-volume synthesis includes the results of a comprehensive evaluation of the literature but is based heavily on (and contains a compilation of) the more than 30 years of large-scale mapping of alluvial deposits and interpretations of environments of deposition accomplished by staff of WES. No new field investigations were conducted, and many of the interpretations, especially those related to valley history, are based on work and concepts developed by the writer.

In contrast to the 1944 publication which was written exclusively for engineers, the present synthesis is aimed at a broad, multidisciplinary audience of geotechnical and environmental engineers and earth scientists, including archeologists, ecologists, geologists, and sedimentologists. Its scope includes the origin and characteristics of the deposits and landforms of the entire alluvial valley, deltaic plain, and eastern chenier plain and adjacent uplands and terraces of Quaternary age.

The synthesis incorporates the results of investigations conducted using technologies such as high-resolution geophysics, remote sensing, and radiometric dating methods that have only been developed during the past 50 years. Despite the new technology base, which includes more than 1,500 radiocarbon dates, many of the chronological reconstructions are necessarily based largely on indirect archeological evidence. Fortunately, the Lower Mississippi Valley records many aspects of 12,000 years of human activity in thousands of archeological sites, and there are well-established interpretations of cultural sequences. These data allow development of a general chronological framework but not a detailed reconstruction of relict channels as Fisk attempted to do 50 years ago. The synthesis discusses the effects of climate on valley processes and landforms and attempts to place in proper perspective the roles of base level changes and faulting which heretofore have been overemphasized.

Geologic processes and controls

In terms of volume and areal extent, most of the deposits of the Lower Mississippi Valley are of Wisconsin and Holocene age (Late Pleistocene) and are dominantly associated with constructional geomorphic processes. Sangamon and older Quaternary deposits mostly form terrace complexes and upland veneers and are presently associated with erosional (destructional) geomorphic processes. Physiographically, the Late Pleistocene alluvial valley is characterized by six, broad, shallow basins which are topographically and hydrologically distinct units. They are separated by either older upland or meander belt ridges. The deltaic plain also is characterized by broad, flat basins which are separated mostly by low distributary ridges.

Quaternary sediments represent deposition in fluvial, lacustrine, eolian, deltaic, and deltaic-marine environments, but those laid down in fluvial environments are far more abundant. Climate, eustatic variations, tectonics and diapirism, and subsidence are processes that have broadly affected the deposits, but the most important process has been continental glaciation.

On multiple occasions during an estimated 2.8 million years, the Lower Mississippi Valley has functioned as a huge sluiceway for glacial meltwater and outwash. The simplistic glacial model based on five glacial cycles is known to be obsolete, but evidence from the alluvial valley area does not allow correlation of deposits/landforms with a more complex model. A conceptual process model is presented indicating the nature of the response of the area to a typical glacial cycle. The model indicates that various parts of the area responded in different ways to base level controls and, in general, that sea level variations have been overemphasized as an impetus for entrenchment except in the coastal zone. It is now known that all alluvial terrace sequences do not represent aggradation during waning glaciation as traditionally believed.

With the other processes considered, loess and sand dunes are the most obvious indications of climate change. The Deweyville Complex is interpreted as a response of the area to increased glacial-stage runoff. With the possible exception of the alluvial fan of the Current River, the area contains no direct evidence of the geomorphic effects of the presumed greater aridity of the Holocene Altithermal episode.

An attempt has been made to link aspects of valley history to eustatic variations only during the Late Pleistocene. A model developed largely from oceanographic evidence is employed which includes five stadial/interstadial high and low sea level stands since the Sangamon stage. There is no direct evidence from the Lower Mississippi Valley area, however, that allows verification of the assumed timing and magnitude of the variations.

All parts of the area have been tectonically influenced by differential downwarping and uplift beginning with continental rifting in the Late Triassic or Early Jurassic and continuing into the Quaternary. The most important tectonic influence has been broad, seaward tilting which has allowed the cyclical deposition of tens of thousands of feet of sediments. Along the Gulf Coast, movements along coast-parallel growth faults and subsidence are active neotectonic processes that have affected nearly all aspects of Mississippi Delta topography and hydrography. With the inclusion of the compaction of recent sediments and other local processes, subsidence rates as high as 2.0 ft per century have been observed.

No part of the Lower Mississippi Valley is completely aseismic, but the preeminent area of active seismicity is the New Madrid Seismic Zone located mostly in northeastern Arkansas and southeastern Missouri. In 1811-1812, four of the largest earthquakes in historic times in eastern North America occurred in the zone. The seismic events caused widespread bank caving, reversal of riverflow, landslides, earth waves, forest destruction, and land sinking; however, the most dramatic geomorphic effects were land fissuring and sand blows caused by liquefaction. These liquefaction features, of which there are millions over an area of about 4,000 sq mi, have been invaluable in unraveling the seismic history of the zone, especially as they are preserved in archeological sites. Results of recent studies indicate that earthquakes large enough to cause localized liquefaction have occurred every several hundred years, but none of the magnitude of the 1811-1812 series have occurred in the last 10,000 years. Besides liquefaction, the most noticeable effect of the 1811-1812 series was the creation of Reelfoot Lake in western Tennessee and the possible creation (but more likely the accentuation) of the St. Francis Sunk Lands of Arkansas.

The New Madrid Seismic Zone is located in an ancient, failed, intraplate rift with complex deep-seated faulting, but very few faults have surface or near-surface expression. Only four faults have been mapped in the zone, and there is one questionable major lineament. Elsewhere in the alluvial valley area, faults with surface expression are rare, and there is no evidence to substantiate the presence of a valleywide rectilinear system of either faults or lineaments as postulated by Fisk. The most active faulting in the region occurs in the deltaic plain area where there have been appreciable Quaternary movements in the form of progressive nonseismic creep along the coast-parallel growth faults. The most widely studied has been the Baton Rouge Fault Zone where there is strong geomorphic manifestation and ample evidence of vertical offsets in highways and bridges.

An aspect of geologic history is an ancestral Mississippi River *per se* (drainage into the Gulf of Mexico) that probably dates back to the Jurassic, but the effective beginning of the system is much more recent. It is believed that the Missouri and Ohio rivers were not part of the drainage system until their courses and directions of flow were changed with the onset of the first continental glaciation in the early Quaternary.

Thirty-nine formations ranging in age from Ordovician (Paleozoic) to Miocene (Tertiary) compose the uplands flanking the alluvial valley and underlie the thick Quaternary alluvial sequence. Cenozoic formations consist mostly of unlithified clays, silts, and sands of terrestrial to marine origin. Mesozoic and Paleozoic formations are mostly limestones, sandstones, and shales. The uplands have been predominantly an erosional landscape during the Quaternary, but upland erosion has increased by orders of magnitude during historic times because of land use practices.

Mississippi River tributary fluvial systems have experienced episodes of entrenchment caused by base level changes in the alluvial valley area. Entrenchment of the southern part of the main valley has been caused by sea level variations while climatically-induced regime changes have been largely responsible elsewhere. That entrenchment process did not result in an incised dendritic pattern on the suballuvial surface as popularized by Fisk. Multiple episodes of planation and scouring by both braided and meandering rivers resulted in a chaotic suballuvial landscape. Computer contouring of the buried surface in the alluvial valley area, presented herein for the first time, portrays a surface characterized by numerous irregular ridges, knolls, depressions, and enclosed basins.

Landforms and geomorphic processes

Depositional landscapes prevail in the area and include the modern (Holocene) floodplain and Pleistocene alluvial terraces of both fluvial and lacustrine origin. It is now recognized that terraces such as the well-known and widespread Prairie terrace represent multiple glacial cycles; whereas most units in the alluvial valley area contain a coarse-grained substratum and a fine-grained topstratum, several distinct environments of deposition are represented. Consequently, following recent precedent, terraces are designated as complexes, and it is appropriate to apply the concept of allostratigraphy, i.e., mappable sedimentary units definable on the basis of bounding discontinuities. Holocene depositional landscapes are classified and mapped in detail according to environments of deposition of which more than 25 are represented.

Widespread upland graveliferous deposits (Upland Complex) flanking the alluvial valley apparently represent a dissected, broad, nonglacial alluvial apron that originated in late Tertiary times in the Appalachians or continental interior. Some glacial outwash may be locally incorporated near the edge of the alluvial valley. Discrete regional terrace levels present in the deposits currently are interpreted as adjustments (primarily erosional) to cyclical Pleistocene base level changes rather than true alluvial terraces.

Next to the Holocene floodplain, valley trains of the Mississippi and Ohio rivers (braided-stream surfaces underlain by glacial outwash) are the dominant landscape of the alluvial valley area. The valley trains are classified according to age (Early and Late Wisconsin stages) and have distinctive morphologies, but all exhibit patterns of relict braid bars and gathering channels. All valley trains also exhibit multiple terrace levels which are interpreted as evidence of cyclic downcutting during waning glaciation. Some of the younger terrace level channel patterns suggest formation during catastrophic outburst flood events.

Uplands east of the alluvial valley, upland ridges within the alluvial valley, and the older valley trains are capped with thick deposits of loess. These eolian deposits represent the glacial-stage deflation of silt from active outwash plains and are closely related to fields of sand dunes which occur mainly on the Early Wisconsin valley trains. The latest and most widespread loess sheet, the Peoria (Late Wisconsin stage), is underlain by remnants of four older loess sheets, but evidence suggests none are older than the Illinoian stage.

The Holocene floodplain surfaces of the alluvial valley are mainly characterized by the active and abandoned meander belts of the Mississippi and Arkansas rivers and intervening backswamp (flood basin) areas. Each meander belt is several miles wide and up to hundreds of miles long and is characterized by natural levees, crevasse splays, distributaries, point bar accretion, abandoned channels, and abandoned courses. Six Mississippi River and eight Arkansas River meander belts or belt segments are recognized in the current interpretation of valley history.

Natural levees are the most widespread and conspicuous landforms of meander belts and the most significant in terms of both prehistoric and historic settlement patterns because of their higher elevations and more competent soils. Major distributaries also exhibit natural levee ridges and other characteristics of meander belts and are perhaps one of the least studied and most underrated alluvial valley landforms in many respects.

The point bar is the most widely studied and best understood meander belt environment, especially in terms of sedimentary characteristics. Point bar deposits constitute the bulk of the Holocene sedimentary sequence but decline in extent in the transition from the alluvial valley into the deltaic plain. The extensive areas of noncohesive, permeable point bar deposits are frequently interrupted by narrow, arcuate masses of cohesive, impermeable clays that fill abandoned channels. Created by neck or chute cutoffs and sometimes characterized by oxbow lakes, abandoned channels of Mississippi River origin number in excess of 600. Attention is directed to the significance to humans, in both prehistoric and historic times, of the typical life cycle that abandoned channels experience between time of cutoff and eventual complete filling.

In addition to composition and internal geometric relationships, the overall valleywide pattern of meander belts and their gross configurations and interrelationships present both problems and important clues in valley history. Significant differences in the sizes of meander belts and the size and abundance of cutoffs are strong indicators of less than full discharge conditions by the river at various times in the Holocene. It is now known with certainty that there is no validity to Fisk's interpretation that such differences are due to separate occupations by the Mississippi and Ohio rivers, and climate does not appear to be a viable explanation. The most viable current hypothesis invokes the belief that the relatively smaller meander belts were never more than major distributaries and that most of the Holocene was characterized by the Mississippi River dividing its flow (at a varying ratio) between two or more major channels. The currently favored model of valley history also includes the observation that stream diversions (avulsions) and distributary and occasionally new meander belt formation was not a random, circumstantial process. Rather, there are definable areas in which probably for multiple reasons, crevassing was relatively abundant and frequent.

The backswamp environment of deposition is well represented only in the southern part of the alluvial valley and reaches its maximum extent in the Atchafalaya Basin area in the transition between the alluvial valley and the deltaic plain. Sedimentary sequences of backswamp deposits exceed 100 ft in thickness in that area and indicate deposition in both deep and shallow swamps. True lacustrine and lacustrine deltaic environments are also represented.

The Holocene Mississippi River deltaic plain is subdivided into five complexes and 16 lobes in the most widely accepted interpretation. Each lobe is attributable to a single set of deltaic distributaries and includes seven distinct environments of deposition. Over time frames measured in centuries, each lobe undergoes a characteristic cycle involving constructional (progradational) and destructional (transgressive) phases as discharge through the system waxes and wanes. Because delta lobes have formed under the ubiquitous presence of active subsidence, the deltaic plain constitutes a composite of multiple, definable, overlapping sedimentary sequences.

Narrow and linear but topographically prominent deltaic distributaries form the skeletal framework of the deltaic plain and are separated by broad expanses of intratidal marshes in more coastal areas and swamps in more inland areas. As in the alluvial valley area, natural levees have dictated the location and configuration of human settlement patterns and communication/transportation routes in both prehistoric and historic times. Classification of marsh and swamp types is based on vegetation which is primarily a function of surface water salinity.

The total sedimentary framework of the deltaic plain also includes materials deposited in the subaqueous delta front (intradelta) and prodelta environments. The former form at the seaward margin of delta lobes while the latter form peripheral to a lobe and are overridden by the coarser materials of the lobe itself. Deltaic-marine environments, related to the destructional phase of a lobe, include bay-sound, beaches and barriers, reefs, and cheniers. The latter developed in southwestern Louisiana marginal to the deltaic plain and represent periods of decreased downdrift transport of fine sediments and shoreline erosion and recession between episodes of mudflat formation.

Nearshore Gulf deposits also represent deposition in a deltaic-marine environment. They occur as a thin veneer directly overlying the eroded and entrenched Late Pleistocene surface beneath the deltaic plain and represent deposition during the Holocene transgression.

Quaternary stratigraphy and chronology

The chronostratigraphy of the area has received considerable attention in this synthesis with discussions centering around 13 paleogeographic reconstructions representing key intervals or periods.

Between about 2,200,000 and 1,300,000 years B.P. (the Early Pleistocene), the Lower Mississippi Valley was directly influenced by events of at least two major glacial cycles. During the first, the narrow, incipient valley must have been widened (to several tens of miles) and deepened significantly to accommodate a load of meltwater and outwash. The overall floodplain was significantly higher than at present, and masses of outwash sometimes were deposited stratigraphically adjacent to the Upland Complex. Throughout the period, the Mississippi River must have flowed through the ancestral Western Lowlands while the Ohio River flowed through the Eastern Lowlands (the St. Francis Basin), but no evidence remains of their exact locations. The two streams probably joined somewhere in western Mississippi.

A cyclical falling base level during that 900,000-year-long period resulted in at least two discrete "terrace" levels within the pre-Pleistocene Upland Complex deposits as a consequence of landscape adjustment: Pleistocene-age depositional components are poorly represented. During each interglacial stage, sea level probably attained a level higher than at present, but no coastal landforms have been identified in the stratigraphic record. Offshore, several thousand feet of sediments accumulated in at least two recognizable sedimentary sequences.

By deductive reasoning, the Intermediate Complex is attributed to events that took place subsequently between about 1,300,000 and 800,000 years B.P. Outcrops of the complex are quite limited in the area, and in general, the unit is one of the least understood in terms of origin. However, a full glacial cycle is indicated, and the surficial deposits constitute the first (oldest) widespread interglacial-stage valley-fill sequence represented by a depositional terrace.

The floodplain level of the Mississippi River, although higher than at present, was well below the average elevation of the Upland Complex.

During the early part of the Illinoian glacial cycle of the Middle Pleistocene (about 800,000 to perhaps 150,000 years B.P.), stream erosion and entrenchment in the coastal zone were extensive, widening the alluvial valley further and removing much of the Intermediate Complex. Subsequently, during the waning glaciation phase of the cycle, a new wave of outwash entered the area as evidenced by a few remnants of valley trains adjacent to Crowley's Ridge and a thick buried sedimentary sequence beneath the deltaic plain and offshore.

A date of about 130,000 years B.P. marks the approximate beginning of the Sangamon stage which was a period of prolonged relative stability in the uplands and the development of meander belts in the alluvial valley. Locations of the latter have been determined largely by inference, and it is believed that the Mississippi and Ohio rivers did not merge until south of the latitude of Vicksburg. Components (alloformations) of the Prairie Complex topstratum dating to this stage predominantly consist of backswamp deposits.

At least twice during the Sangamon stage between about 130,000 and 120,000 years B.P., sea level attained an elevation more than 10 ft above that of the present. As a consequence, the coastline was located north of a Lafayette-New Orleans line and was characterized by extensive beach/barrier island systems, including the Ingleside Barrier Trend.

There is considerable confusion and uncertainty concerning the timing of events that occurred and the Prairie Complex deposits that formed during the subsequent 100,000-year period. Several major eustatic cycles apparently characterized the period now referred to as the "Eowisconsin" stage (120,000 to 70,000 years B.P.) and the following Late and Middle Wisconsin stages (70,000 to 30,000 years B.P.). However, the precise response of the Lower Mississippi Valley area to each cycle is only speculative, and the meager evidence that exists is sometimes conflicting.

A new concept is offered herein that tentatively correlates small beach features and marine deposits south of the Ingleside Trend with an early high sea level stand of the "Eowisconsin" stage. This was followed, in turn, by formation of the Red River deltaic plain of southwestern Louisiana and two Mississippi River meander belts that are evidenced by the Avoyelles Prairie, the Lafayette Meander Belt, and the Mt. Pleasant Bluff locality. The three units were separated by episodes of entrenchment and base level change. A Mississippi River deltaic plain probably extended south of the present coastline in central Louisiana, and elsewhere no major beach features are known. Essentially, no evidence has been recognized indicating the nature of the landscape in the alluvial valley area during this stage.

The deltaic plains and meander belts apparently were abandoned, and major stream entrenchment occurred coincident with onset of the Early Wisconsin glaciation. In the alluvial valley area, streams degraded in response to a climatically-induced change in regime. Large amounts but not all of the glacial outwash deposited during earlier cycles were reworked and flushed from the valley. During the accompanying eustatic low stand, a weathered erosional surface formed beneath the present deltaic plain. The buried horizon has distinctive physical and geotechnical properties and has been detected in borings and geophysical surveys.

Waning of the Early Wisconsin glaciation about 70,000 years B.P. resulted in the deposition of large volumes of outwash in the alluvial valley and the formation of extensive valley trains. The two largest surviving remnants constitute most of the present landscape of the Western Lowlands and Macon Ridge. The valley trains are characterized by multiple terrace levels which reflect episodic outwash deposition, and the higher (easternmost) levels are veneered with Peoria loess.

Deposition of the Early Wisconsin outwash was so rapid and widespread that the mouths of many alluvial valley tributaries were effectively blocked by alluvial drowning. This led to the formation of extensive lakes as evidenced by lacustrine plain terraces and lake beaches that are younger than the adjacent Prairie Complex surfaces. The most extensively investigated and described lake was Lake Monroe which formed in the Ouachita River valley, but others formed in tributaries in western Kentucky, Tennessee, and Mississippi.

Evidence from the Lower Mississippi Valley supports an interpretation (albeit argumentative) that the Early and Late Wisconsin glaciations were separated by a brief but significant interstadial (Middle Wisconsin stage) during which sea level rose but did not quite reach its present level. There is no direct manifestation of events of this stage in the alluvial valley proper, but it is apparent that the Mississippi and Ohio rivers did not change from a braided to a meandering regime. In the coastal area, an embayment may have formed in the Atchafalaya Basin area, and there is evidence for a now-buried beach trend and a seaward thickening wedge of marine sediments overlying an erosional surface in the Lake Pontchartrain area.

Onset of the Late Wisconsin glaciation (formation of the Laurentide ice sheet) about 30,000 years B.P. triggered another cycle of major valley degradation and entrenchment in the coastal zone. Another weathered erosional surface formed (this time on Middle Wisconsin-stage deposits), and thousands of borings, especially in the New Orleans area, have allowed delineation and contouring of narrow, dendritic entrenched stream systems in exceptional detail. The erosional surface, known simply as the "top of the Pleistocene," has major engineering significance since it is the closest approximation of "bedrock" from a foundation viewpoint.

Between about 25,000 and 14,000 years B.P., which includes the time of maximum Late Wisconsin glaciation (about 18,000 years B.P.), many Coastal Plain streams adapted their regimes to accommodate significantly higher stream discharges while being graded to lower-than-present base levels. The Deweyville Complex, with its distinctive "oversized" meanders and abandoned channels, is well developed on the Ouachita River and has been identified on the Arkansas and Red rivers but is absent in the Mississippi alluvial valley *per se*. Full-glacial conditions in the latter area were marked by a cool, moist climate, a forest characterized by boreal species such as spruce and fir, and the deflation of large quantities of silt during seasonal loess-forming episodes.

Relatively little outwash from the waning of the Laurentide ice sheet entered the Western Lowlands since the Mississippi River diverted through a gap in Crowley's Ridge into the St. Francis Basin about 16,000 years B.P. With the flow of the Mississippi and Ohio rivers combined, a valley train rapidly developed southward through the alluvial valley at a floodplain level lower than that of the Early Wisconsin outwash. Relict braided channel patterns are still remarkably well preserved on the higher terrace levels.

Glacial runoff through the area is believed to have reached its peak about 12,000 years B.P. and abruptly declined thereafter. It was a time of rapid and significant amelioration of climate, an effective end of loess and sand dune formation, and a shift in vegetation to a deciduous hardwood forest. It was not, however, the actual end of all glacial outwash deposition. Due to events in the Great Lakes area, there was a final brief (500-year-long?) pulse of runoff that occurred prior to about 9,500 years B.P. This pulse, possibly related to a lake outburst event, is believed to have caused diversion of the Mississippi River through Thebes Gap into the St. Francis Basin and deposition of the Charleston Fan in southeastern Missouri. Within no more than 500 years thereafter, the Mississippi River shifted to a basically meandering regime throughout most of its valley. For the first time in valley history, archeological evidence directly assists in chronostratigraphic reconstructions.

For the last 9,500 years, the Mississippi and Ohio rivers have flowed in a combined channel south of Cairo, and several major diversions have led to the formation of six meander belts or meander belt segments. A short meander belt segment in the northeastern Yazoo Basin, designated the South Lake Meander Belt, represents the probable earliest manifestation (Stage 6) of the Mississippi River. It is obviously less than a full-flow channel, but the location of other channels of this age (10,000 to 9,000 years B.P.) are not known.

The Stage 5 channel is known only from the Coldwater Meander Belt of the Yazoo Basin wherein the Mississippi River apparently achieved full-flow conditions for a short period between about 9,000 and 8,000 years B.P. The Stage 5 channel north of Memphis must have been near the present meander belt, but specific cutoffs of this age have not been identified. South of Vicksburg, the channel must have flowed near the western valley wall, but a well-developed meander belt may not have been present because of the influence of sea level rise.

The Big Creek Meander Belt marks the location of the Stage 4 channel north of Memphis, but for more than 150 mi south of that location, the river apparently divided its flow rather equally between two channels. The Little Mound Bayou and Bear Creek meander belts through the eastern and central parts of the Yazoo Basin mark the locations of the channels between about 7,000 and 6,000 years B.P. Through east-central Louisiana, the river is believed to have followed the Tensas Meander Belt and thence along the western valley wall to the Gulf.

A single full-flow channel is indicated for Stage 3 the entire distance between Cairo and the Gulf. Identified segments include the St. Francis, Sunflower, and Cocodrie meander belts which are interpreted to have been active between about 6,000 and 4,500 years B.P.

During Stage 2 (4,500 to 3,000 years B.P.), the Mississippi River apparently once again occupied two nearly equal meander belts through the Yazoo Basin area. Because of a major diversion in the lower St. Francis Basin, the river formed the Yazoo Meander Belt along the eastern side of the Yazoo Basin and another one along the route of the present river. The two channels merged near Vicksburg and flowed southward in the Walnut Bayou Meander Belt to near the mouth of the Red River. At this point, some of the river discharge flowed to the Gulf through the Teche Meander Belt, but a major distributary formed, and for the first time during the Holocene a significant amount of the discharge moved along the eastern valley wall past Baton Rouge.

By about 2,000 years B.P. (Stage 1), essentially modern conditions had developed in the alluvial valley, and the river everywhere was occupying a single channel and meander belt.

Throughout the Holocene, the Arkansas River occupied at least eight separate channels in the alluvial valley area, but the resultant sequence of 14 meander belt segments is too complex to allow concise summation. These meander belts radiate southeastward from the Little Rock area and occupy positions both east and west of Macon Ridge. On more than one occasion, the Arkansas River flowed as far south as the mouth of the Red River before becoming tributary to the Mississippi River.

In brief overview, it can be stated that with decreasing age of meander belts, archeological evidence becomes progressively more valuable. During the past 2,500 years, such evidence is quite definitive and abundant and has allowed occasional highly detailed reconstructions of channel shifts and their accompanying landscape (and environmental) changes. Nevertheless, in general it has only been possible to estimate the ages of whole meander belts or segments and seldom with a resolution finer than 500 years. Archeological evidence, which has proven to be much more reliable than radiocarbon dating because of problems with sample contamination, has allowed the accurate dating of probably not over one percent of the specific cutoffs. Archeological evidence further substantiates that meander belt reconstructions are exceedingly complex because of complications arising from the occupation of abandoned meander belts by tributary streams. Most meander belts appear to have had multiple episodes of active fluvial deposition and overbank sedimentation.

While the Mississippi River was flowing in a meandering regime to at least as far south as Greenwood by 9,000 years ago and physical evidence of meander belts are visible, sea level was still tens of feet below its present level and coastal landforms of that age are deeply buried. During the following 5,000 years, sea level rose rapidly to essentially its present level, causing the Gulf shoreline to rapidly transgress inland across previously exposed Middle Wisconsin and older deposits. About 6,000 years ago when sea level was only 10 ft or so below present, a large barrier spit known as the Pine Island Beach Trend formed in the greater New Orleans area, and Gulf waters spread inland to form a shallow embayment in the lower end of the alluvial valley to the approximate latitude of Baton Rouge. Located south of the Baton Rouge Fault Zone in an area of active subsidence, the beach trend is almost completely buried by Holocene deltaic sediments but is in an extraordinary state of preservation as revealed by thousands of borings.

Offshore from central Louisiana and west of the Pine Island Beach Trend, the Mississippi River is believed to have begun forming discrete delta complexes by about 9,200 years B.P. The oldest recognized sequence is designated the Outer Shoal Complex and is marked only by submerged shoals and shorelines representing the destructional phase.

The Maringouin Complex is the oldest delta complex with buried constructional phase deltaic deposits. These deposits have been dated between 7,200 and 6,000 years B.P. and formed when sea level was between 25 and 15 ft below present. As sea level continued to rise, the Maringouin Complex was abandoned and the locus of deltaic sedimentation shifted slightly inland. This resulted in formation of the Teche Complex which is manifested by well preserved distributary ridges and interdistributary deposits. Most of the Teche Complex deltaic deposits date to between 6,000 and 4,500 years B.P, but there is evidence that some of the easternmost lobes were active until nearly 3,000 years B.P.

The final rise in sea level to its present stand is believed to have caused a widespread ravinement and destruction of a significant (but not well established) amount of the distal part of the Teche Complex. About 4,500 years B.P., occupation of the Stage 2 channel in the alluvial valley area triggered the development of the St. Bernard Complex which forms the framework for much of the eastern part of the deltaic plain. Six discrete lobes of the complex have been recognized, each with a well-preserved distributary network.

Some lobes of the St. Bernard Complex apparently remained active until only about 1,000 years B.P., but initial sedimentation in the Lafourche Complex probably started about 3,500 years B.P. This deltaic sequence rapidly filled a shallow embayment between surviving lobes of the Teche and St. Bernard complexes. It also resulted in blockage of the southern end of the Atchafalaya Basin, terminating brackish-water conditions and converting it into a totally freshwater environment.

About the same time that the last lobe of the Lafourche Complex became active (about 1,000 years B.P.), a diversion of flow caused an increase in flow past New Orleans and formation of the Plaquemines Complex. This event marks the establishment of the present channel of the river and eventually the formation of the modern or "Balize" delta. The latter has been marked by a distinctive change in the physiographic expression of the active delta lobe because for the first time during the Holocene, the river is discharging its sediment directly into deep water at the edge of the continental shelf.

Also for the first time during the Holocene, the Mississippi River is diverting a significant amount of its discharge through a well-defined channel through the center of the Atchafalaya Basin. Triggered by artificial channel changes in historic times to improve navigation, the diversion has caused the effective end of lacustrine conditions in the lower Atchafalaya Basin and the beginning of a new delta lobe in central, coastal Louisiana. That small lobe is the only part of the deltaic plain in an active constructional phase: the rest is undergoing destruction as a result of both natural and anthropogenic factors.

Special engineering considerations

Information in this synthesis regarding geomorphic and geologic processes, landforms, and sedimentary sequences are directly relevant to the planning, design, construction, operation, and regulation of water resources projects, both from a traditional foundation engineering viewpoint and a natural geomorphic systems viewpoint. There are also brief discussions of several other regional geomorphic and geologic factors that have special engineering significance.

The coarse-grained substratum (glacial outwash) and the contiguous point bar sequences of the alluvial valley constitute the alluvial aquifer, one of the largest and most exploited sources of fresh water in the United States. Although water in the aquifer typically is hard to very hard with high concentrations of iron, manganese, calcium, magnesium, and dissolved solids, it is extensively used for urban and industrial purposes and especially for irrigation. As a consequence, portions of the aquifer are experiencing series depletion, and there is growing concern about contamination and water quality considerations. The character and configuration of the topstratum as described herein are therefore of considerable importance since they strongly influence multiple aspects of aquifer recharge.

It is well recognized that bank caving is an integral component of the process of river meandering and is directly related to the nature of the fine-grained topstratum of the alluvial sequence. As part of the extensive program of artificial bank stabilization along the lower Mississippi River, considerable knowledge has been developed regarding bank caving processes. Attention has been especially focused on flow failures because of liquefaction in point bar deposits since these have often led to sudden and spectacular failures involving tens of acres of bank.

Landsliding is a frequent and continuing geomorphic process affecting much of the bluffs along the eastern side of the alluvial valley. Although earthquakes can be a triggering mechanism, the underlying cause is slope instability due to the undercutting and oversteepening of the bluffs by lateral stream migration. Detailed studies have been made at locations such as Natchez where considerable damage to valuable historic properties is a recurrent problem. It has been determined that the instability involves seven mass wasting processes, including rotational slumps, soil slides, soil fall, soil creep, solifluction, mudflow, and surface wash. The seven are interrelated and part of a continuum of processes, and any individual failure may be characterized by a combination of several processes.

During the past several decades, engineers concerned with river regulation have turned more and more to the field of potamology for answers to questions regarding river meandering and long-term stability. More than 30 individual geologic, hydrologic, and hydraulic variables have been identified that have influenced the degree of equilibrium that has been achieved by the Mississippi River at least during the last few centuries. During the past 30 years, man-made structures have upset the natural balance, however, and observations are made in the synthesis regarding changes in such variables as valley slope, annual hydrograph, bed and bank materials, total sediment load, and bed-load characteristics.

Future Research Needs

It has become apparent that the preparation of an effective synthesis is seriously constrained by dozens of unanswered major questions regarding almost all aspects of the geomorphology and geologic history of the Lower Mississippi Valley. Some are of such magnitude and complexity that only the advent of new technologies or an overall advancement in the state of knowledge will provide answers. However, there are other questions that are amenable to research and investigation at scales that are currently feasible. Some of these include:

a. What are the locations and extent of probable inliers of glacial outwash within the predominantly nonglacial Upland Complex?

b. What environments of deposition are represented and what processes were involved in formation of the Intermediate Complex?

c. Why are there no recognized loess sheets older than the Illinoian stage?

d. Can a correlation be made between alloformations tentatively assigned to the "Eowisconsin" stage and specific eustatic cycles or events?

e. How old (what glacial stage) are the areas of questionable Prairie Complex glacial outwash located just west of Crowley's Ridge?

f. Why is only one loess sheet present on the Lafayette Meander Belt and in the Avoyelles Hills area if the units are as old as postulated herein?

g. What evidence would confirm that the Lafayette and Mt. Pleasant Bluff Meander Belts and the Red River Deltaic Plain are of the same eustatic cycle?

h. Is there evidence from within the Lower Mississippi Valley to establish the probable sea level position during the Middle Wisconsin stage?

i. What are the radiometric ages of the two and possibly three buried Pleistocene weathered horizons beneath the deltaic plain?

j. What processes were responsible for the fluvial response that is manifest by large meander scars in the Deweyville Complex?

k. Is there a way to differentiate the thick substratum sand and gravel sequence at least according to major glacial stage?

l. Why did multiple terrace levels form and why were braided channel networks apparently abandoned so rapidly on the Early and Late Wisconsin valley trains?

m. What chronostratigraphic evidence for base level changes or major climate changes in the alluvial valley are preserved within alluvial fans?

n. What chronostratigraphic evidence exists in backswamp areas that will allow a refined model of meander belt formation?

o. Does the absence of well-developed meander belts south of Natchez indicate an anastomosing regime, and if so, what is the causal connection with delta complex formation?

p. Is climate a relevant variable in the observed major differences in the frequency and distribution of cutoffs between meander belts?

q. Is there a correlation between episodes of increased or decreased flooding frequency on the Mississippi River and avulsions or the formation of delta lobes?

r. Have large earthquakes comparable to the 1811-1812 series occurred in the New Madrid Seismic Zone during the last 10,000 years?

s. How can the traditional classification schemes (e.g., environments of deposition) used in the mapping of alluvial deposits be modified/refined to provide even greater utility in the types of problems encountered in current engineering practice?

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Appendix A Compilation of Geotechnical Property Test Data

Between the 1950s and the 1970s, several publications were prepared by and for the Corps of Engineers which contain compilations of the results of soils laboratory tests on samples from borings and their analysis in terms of the typical physical properties of certain environments of deposition (e.g., Kolb 1962, Kolb and Kaufman 1967, Kolb and Van Lopik 1958, Montgomery 1974, Sherman and Hadjidakis 1962).[1] Even though each of those publications include graphical presentations of statistical data on hundreds of soils tests (e.g., histograms showing ranges of values for a given physical parameter), the original data were not preserved in raw form or databases. During the past two decades, several attempts have been made to expand such compilations both in terms of the amount of test results included and the number of environments considered (McBride 1991, Sacre 1987, Whitworth 1988). Fortunately, in the newer compilations, the raw data were entered into, analyzed by, and preserved in digital relational databases.

In the present synthesis, the three databases cited above, although varying considerably in format and software, were successfully manipulated (after appreciable effort) and converted to form a new master data file. The final master file consists of nearly 60,000 records (measurable physical parameters) from 22,749 individual tests representing 1,484 borings. These were separated into subfiles according to the environment of deposition (as identified by the authors cited above) and transferred into a spreadsheet program for statistical analysis.

Inclusively, the master file contains results on 34 different physical parameters representing 16 different fluvial, deltaic, and marine environments, mostly of Holocene age. However, when all data sets of less than 100 records (total count) were eliminated, there were sufficient data to analyze only nine parameters (water content, liquid limit, plastic limit, plasticity index, dry density, wet density, cohesive strength, blow count, and D10 grain size) for a maximum of

[1] References cited in this appendix are located at the end of the main text.

14 environments. The results are presented by environment in Figures A1 through A12 and by parameter in Figures A13 to A21. With those criteria, there are data on all nine parameters only for the point bar environment and, at the other extreme, on only four parameters for the marsh environment. Regardless of shortcomings, the total amount of analyzed and graphically presented data, representing 53,977 records, represents the most comprehensive file of geotechnical data ever assembled for the Lower Mississippi Valley area.

The selected method of portrayal, simple histogram plots showing the percentages of the total counts that fall within certain ranges of parameter values, is intended to provide the reader with a graphical indication of the frequency distribution of the data. Because of the high similarity among many of the environments, it is important to be able to readily discern the total ranges of the values for a given parameter and the degree to which the ranges overlap.

The data in this appendix are intended to provide the reader with an indication of the *typical* geotechnical characteristics of samples from given environments. Despite the appreciable size of the database, it is insufficient to allow a determination of either spatial or temporal trends within the area. Individual studies such as Kolb (1962) have established that there are regional variations in the lithology and physical properties of natural levee and point bar deposits, but quantification of the range of values for particular parameters throughout the entire area is not possible. In other cases, such as with the deltaic and marine environments, the geographic extent is so limited that spatial variations, if present, are too subtle to discern. Variations in parameters within a given environment as a function of age, such as in the cohesive strength of interdistributary deposits in different delta lobes, are suspected but are beyond the scope of this synthesis and are discussed in detail in McBride (1991).

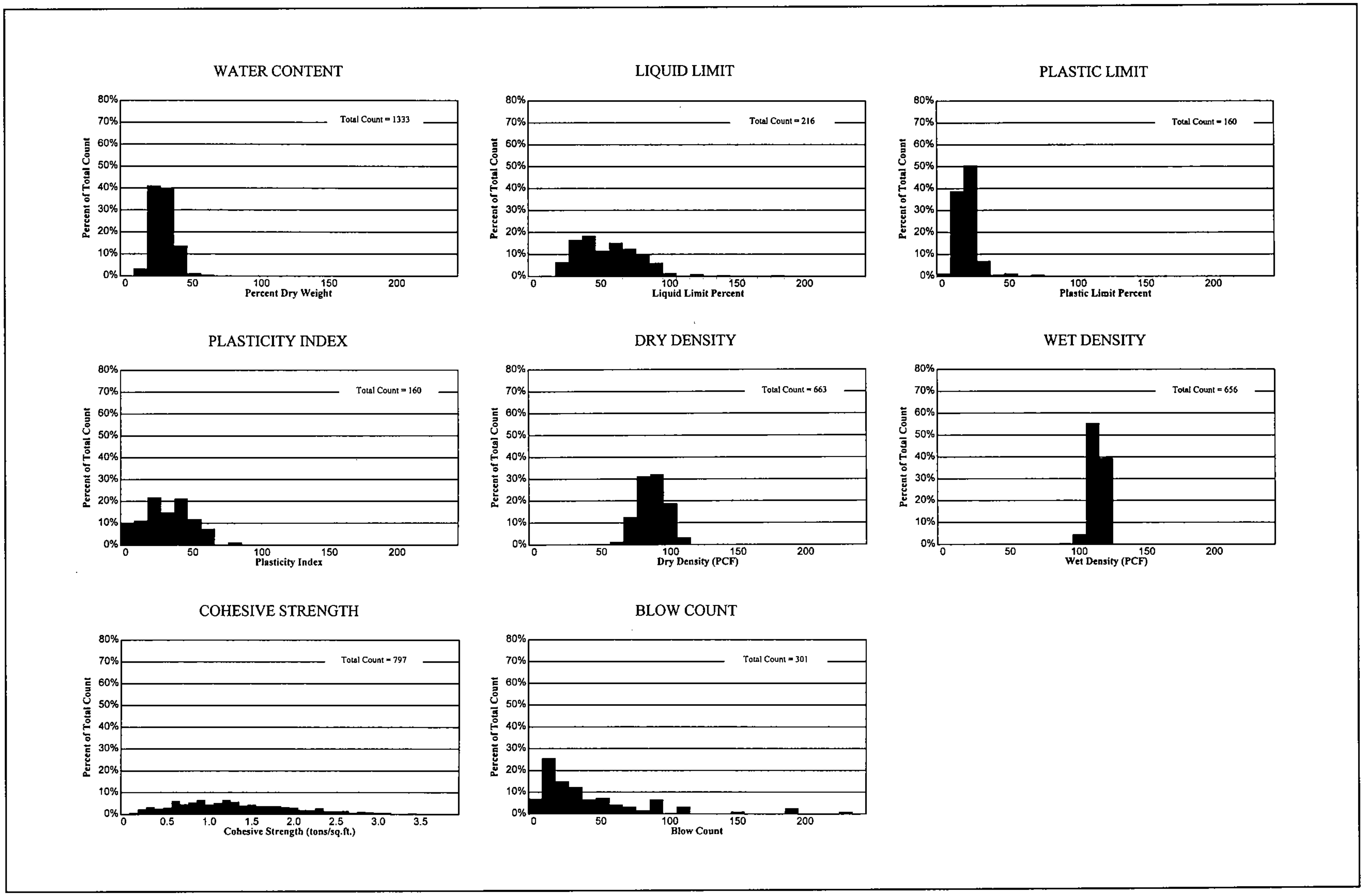

Figure A1. Geotechnical characteristics of shallow Pleistocene deposits from beneath the deltaic plain

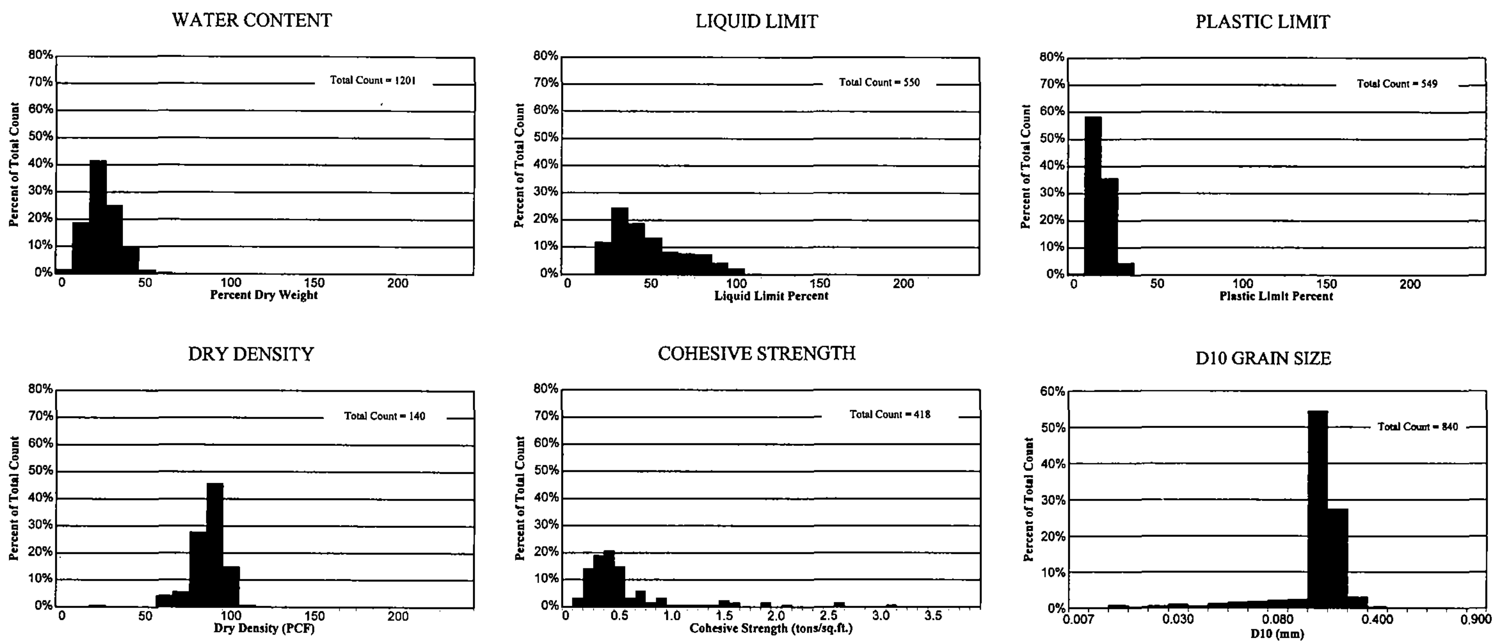

Figure A2. Geotechnical characteristics of Wisconsin-Stage braided stream (valley train) interfluve deposits

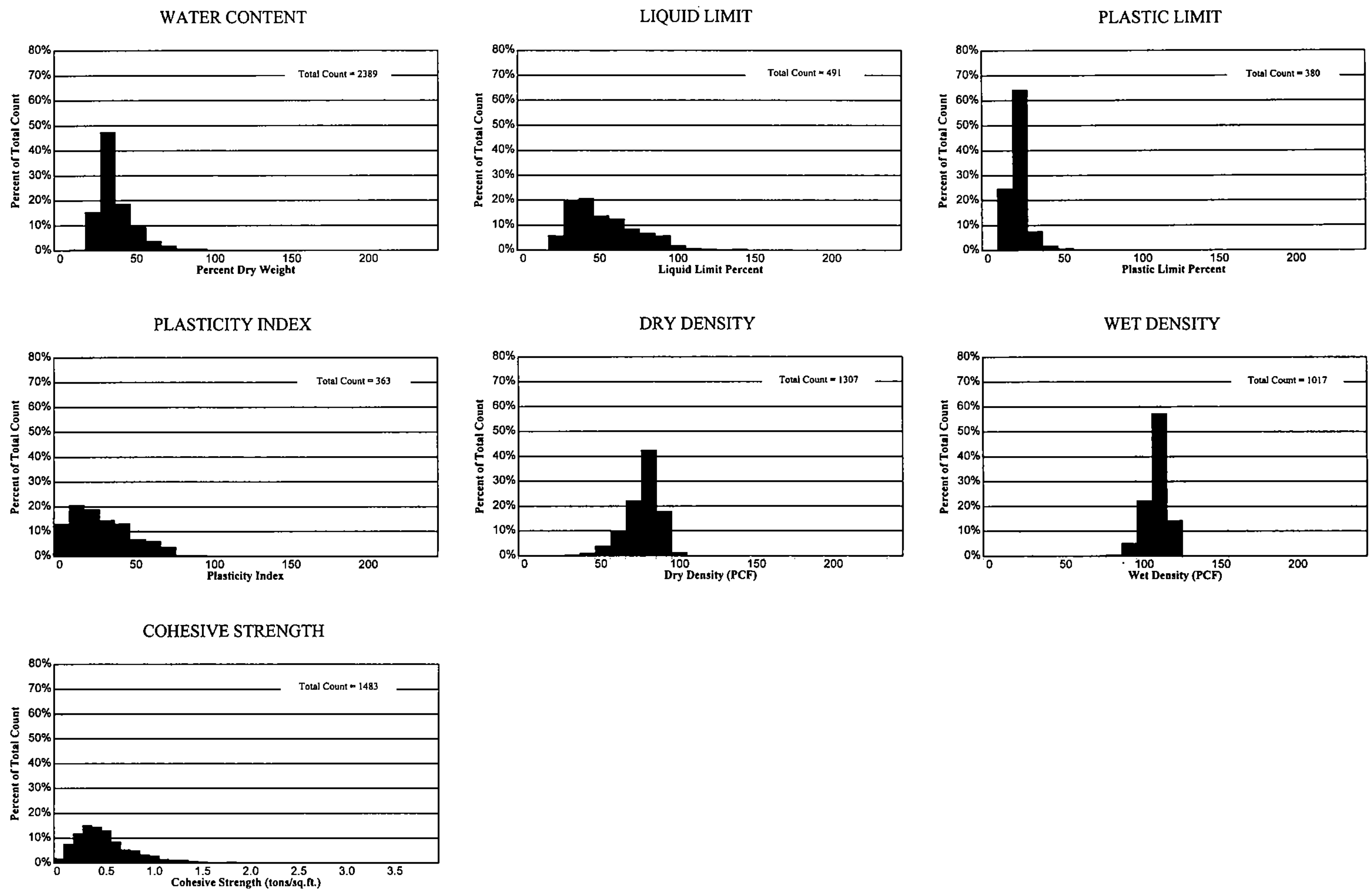

Figure A3. Geotechnical characteristics of Holocene natural levee deposits of the alluvial valley

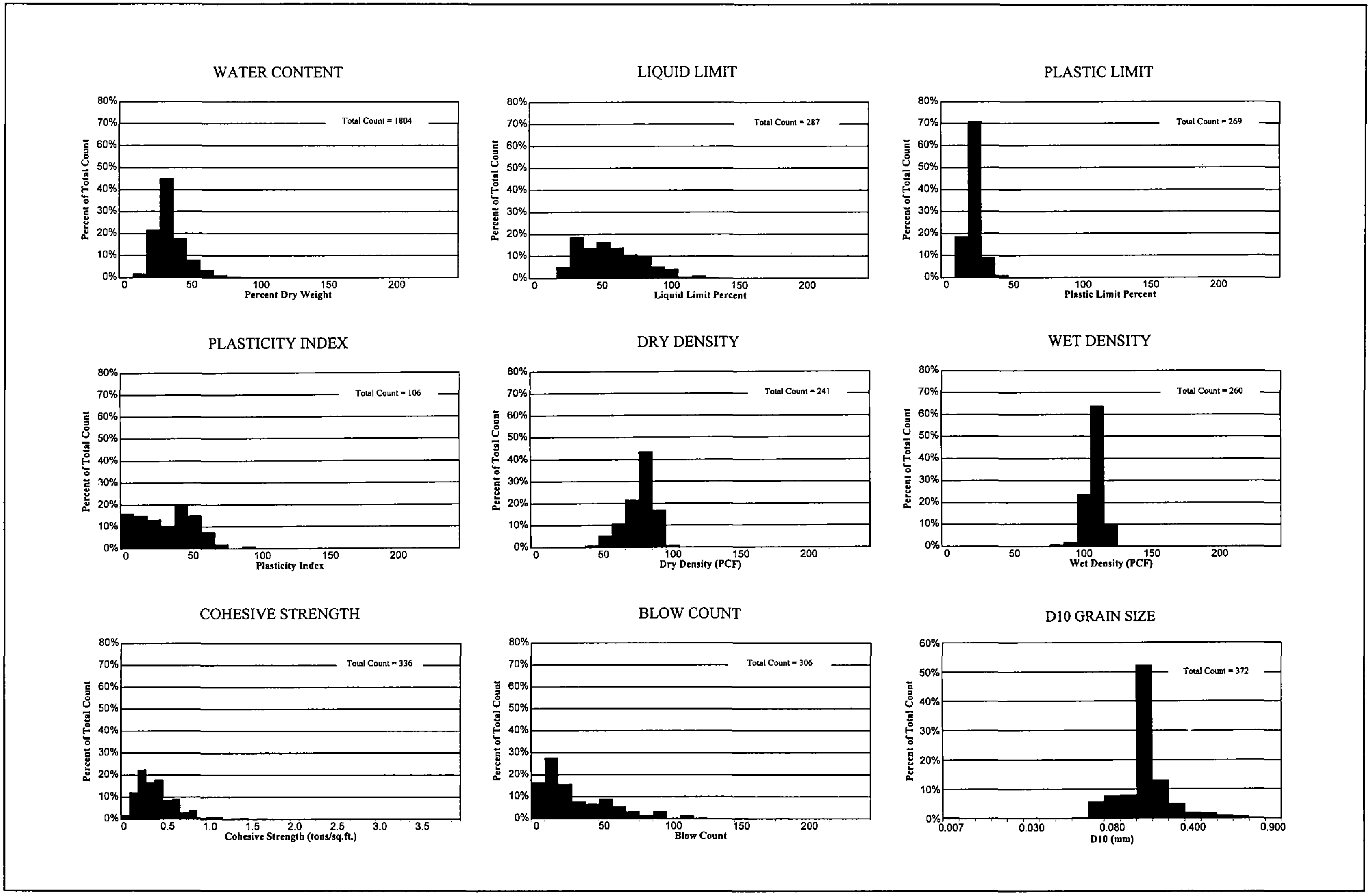

Figure A4. Geotechnical characteristics of Holocene point bar deposits of the alluvial valley

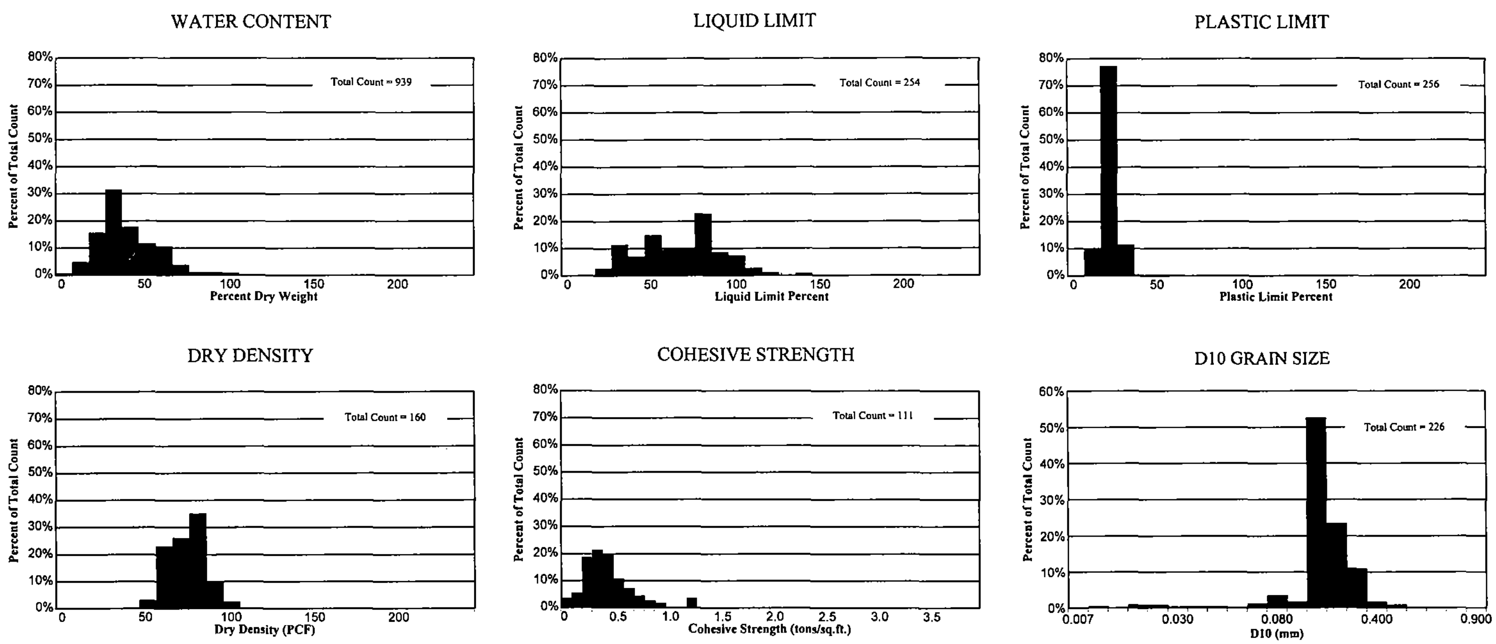

Figure A5. Geotechnical characteristics of Holocene abandoned channel deposits of the alluvial valley

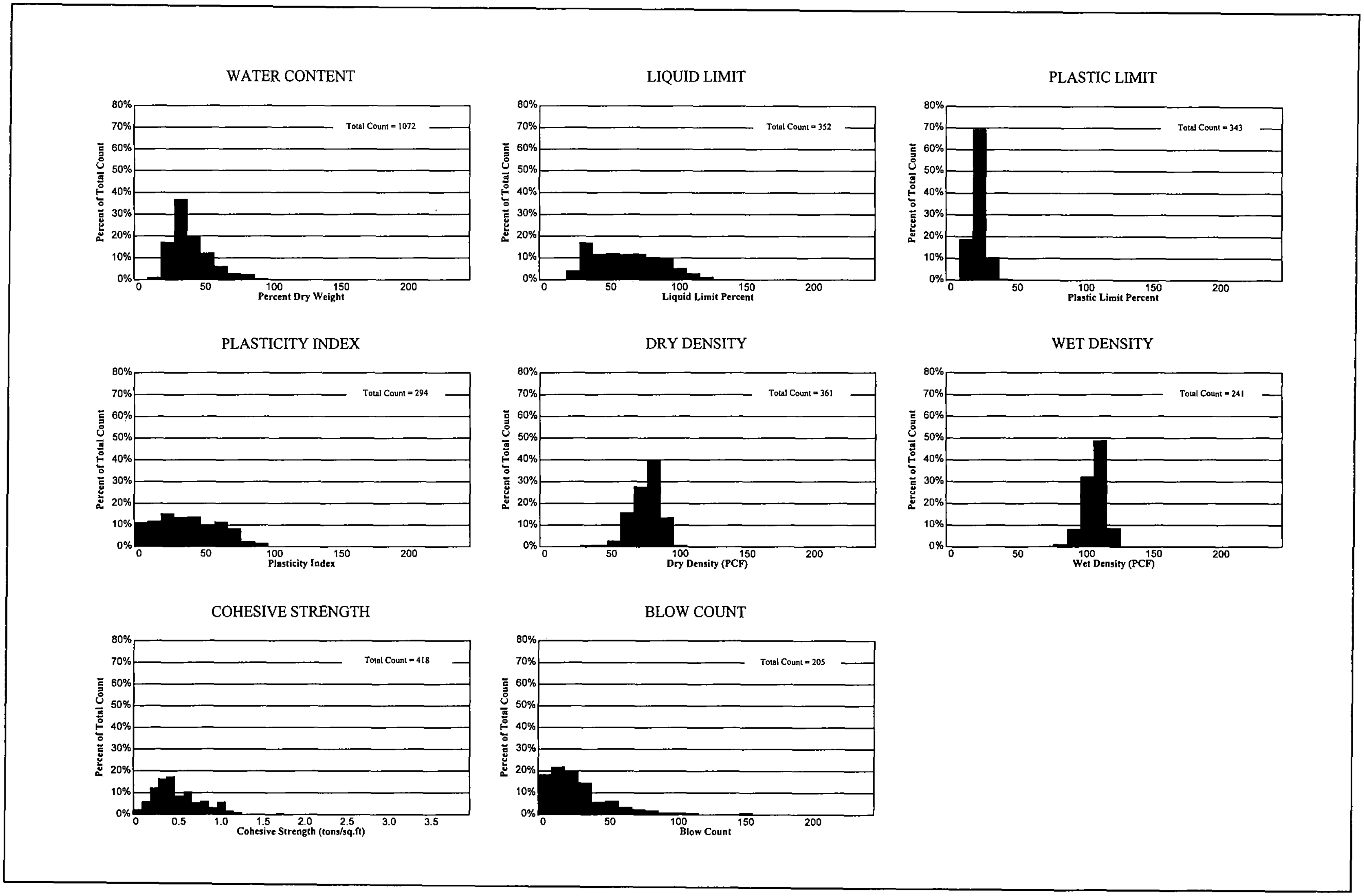

Figure A6. Geotechnical characteristics of Holocene abandoned course deposits of the alluvial valley

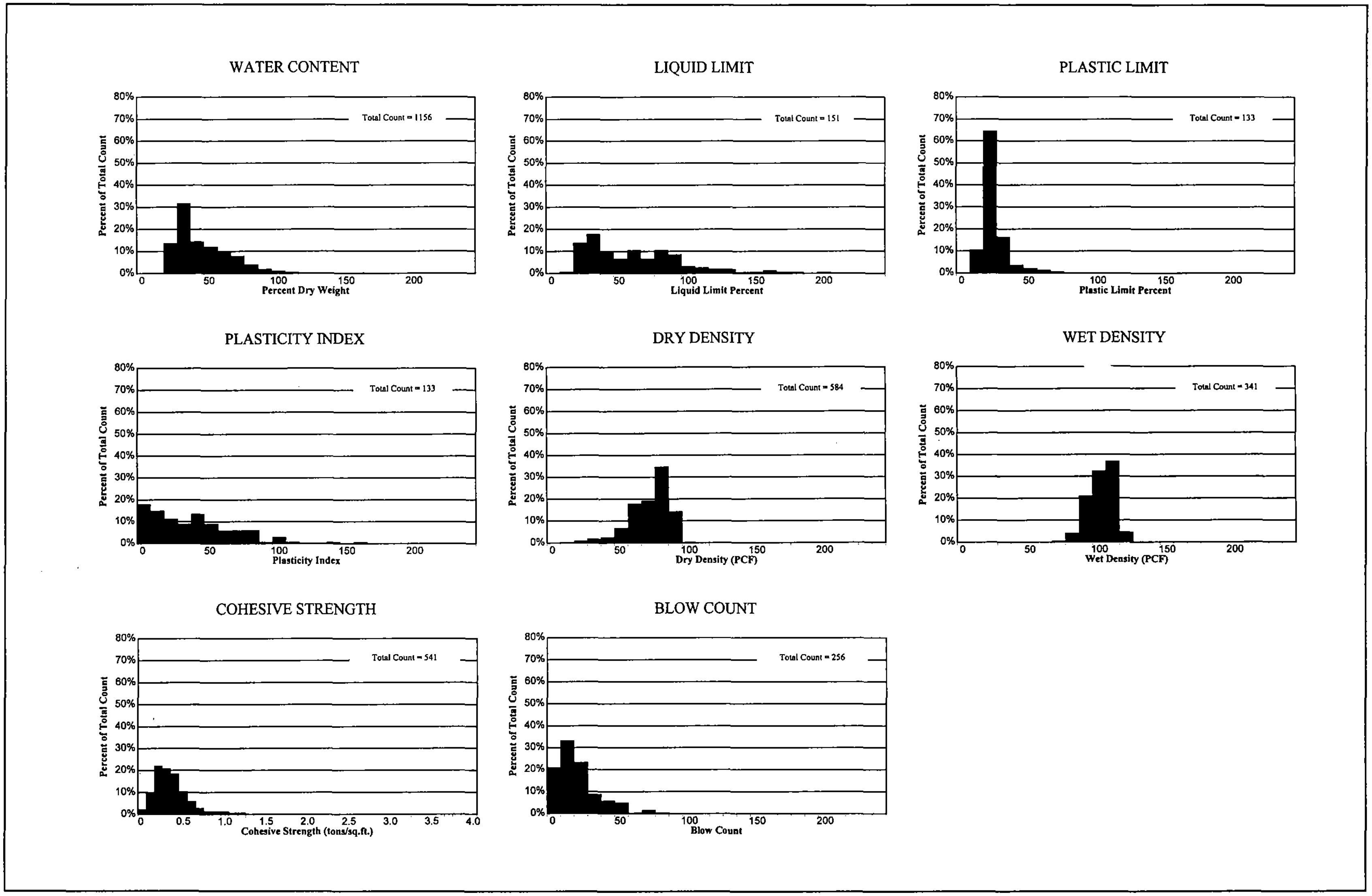

Figure A7. Geotechnical characteristics of Holocene abandoned distributary deposits of the deltaic plain

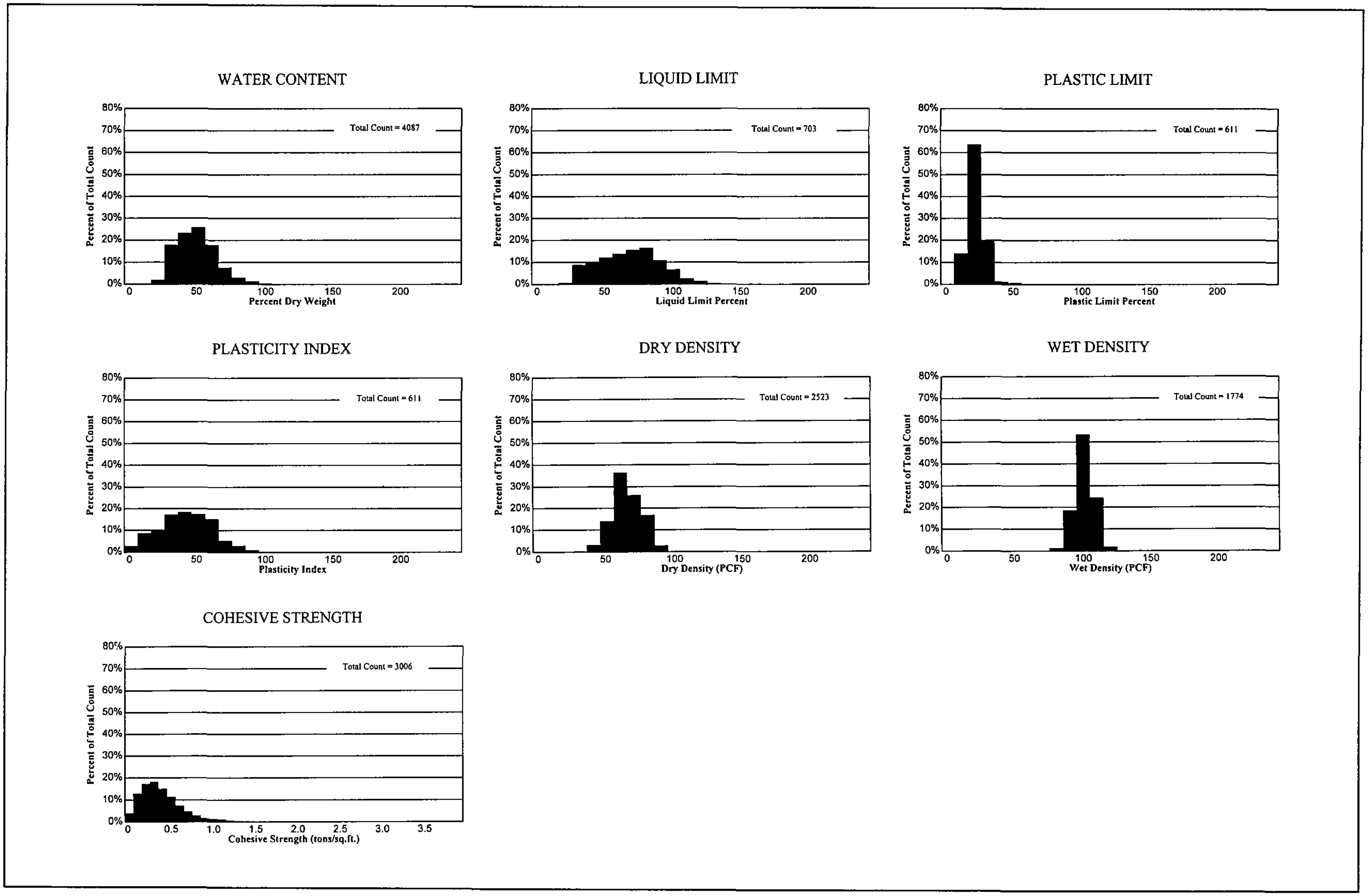

Figure A8. Geotechnical characteristics of Holocene interdistributary deposits of the deltaic plain

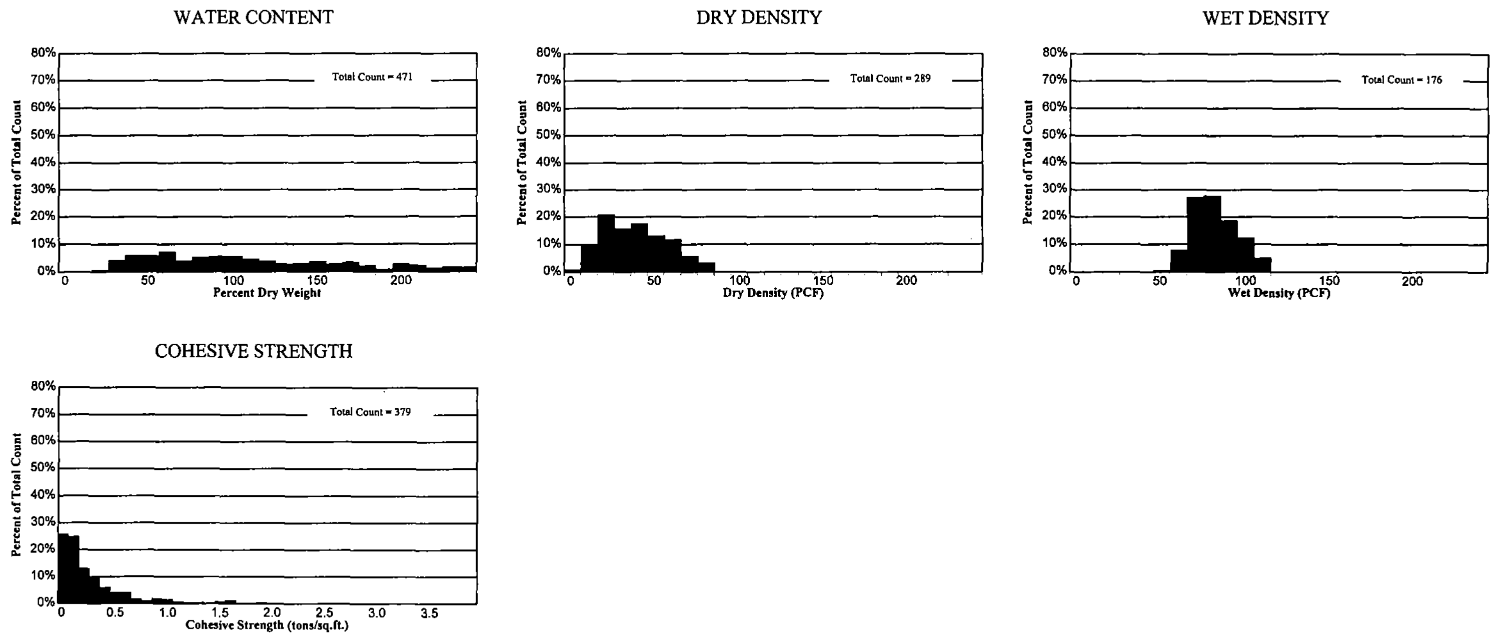

Figure A9. Geotechnical characteristics of Holocene marsh deposits of the deltaic plain

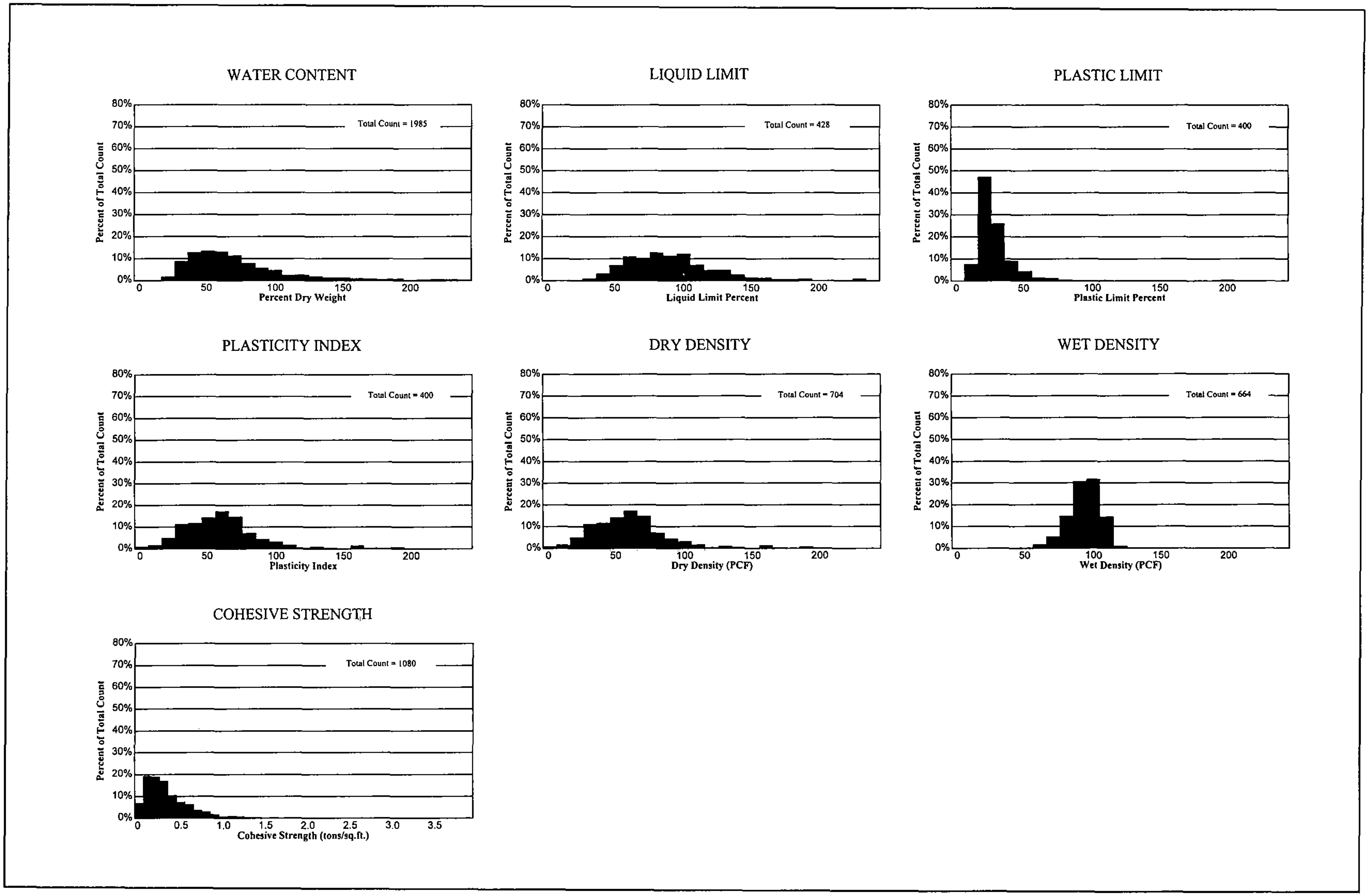

Figure A10. Geotechnical characteristics of Holocene swamp deposits (backswamp and inland swamp environments) of the deltaic plain

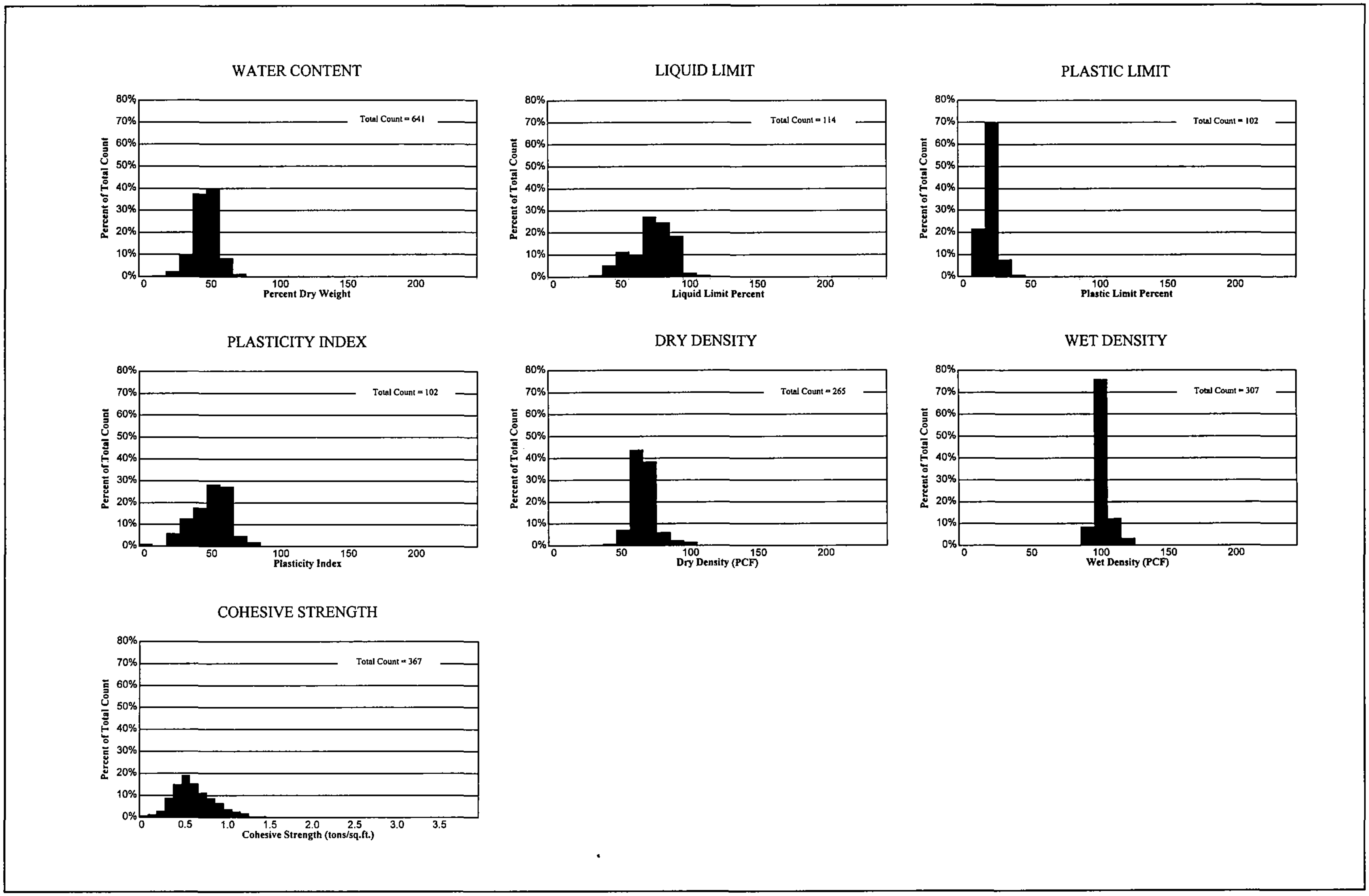

Figure A11. Geotechnical characteristics of Holocene prodelta deposits of the deltaic plain

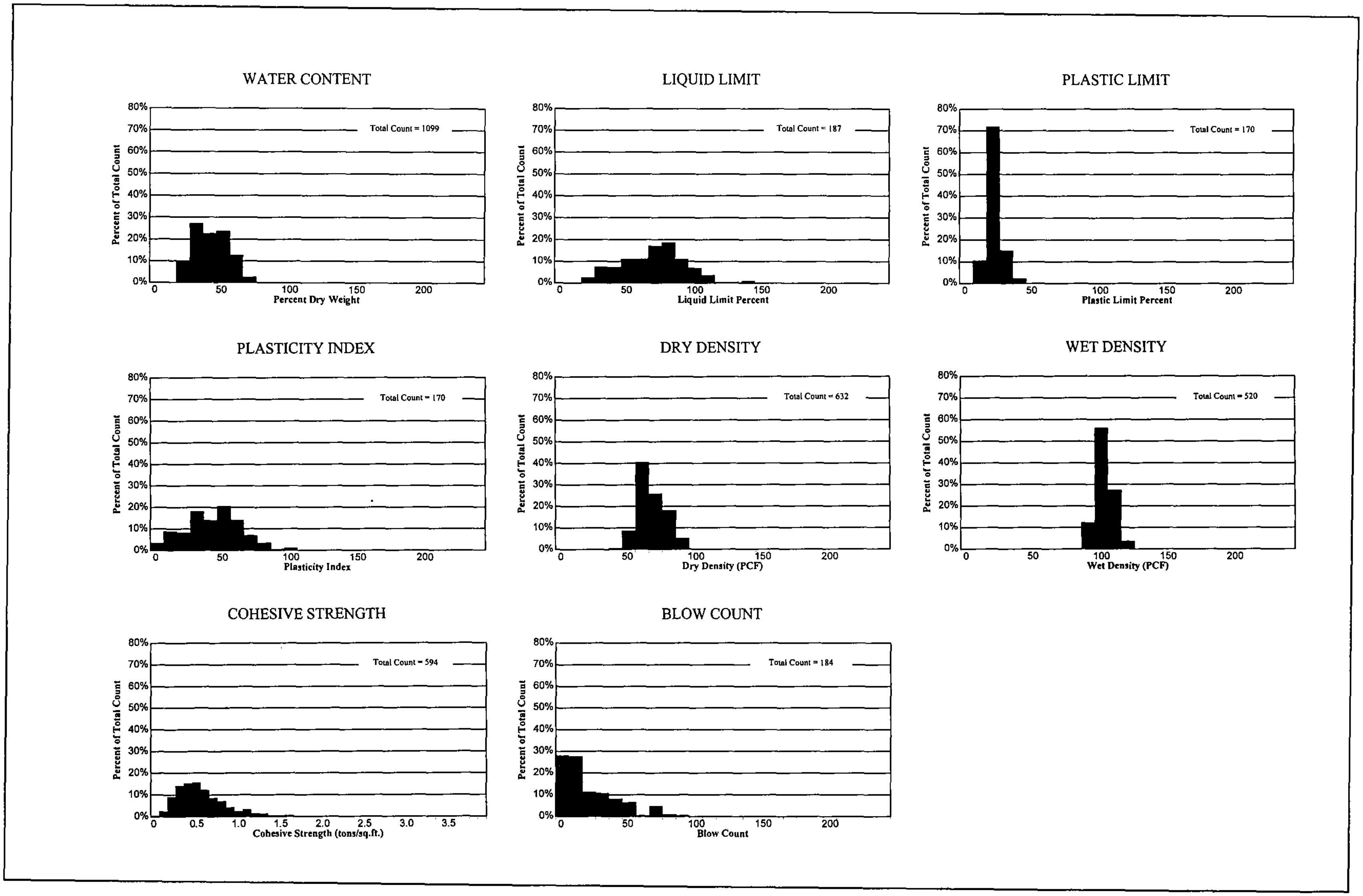

Figure A12. Geotechnical characteristics of Holocene intradelta deposits of the deltaic plain

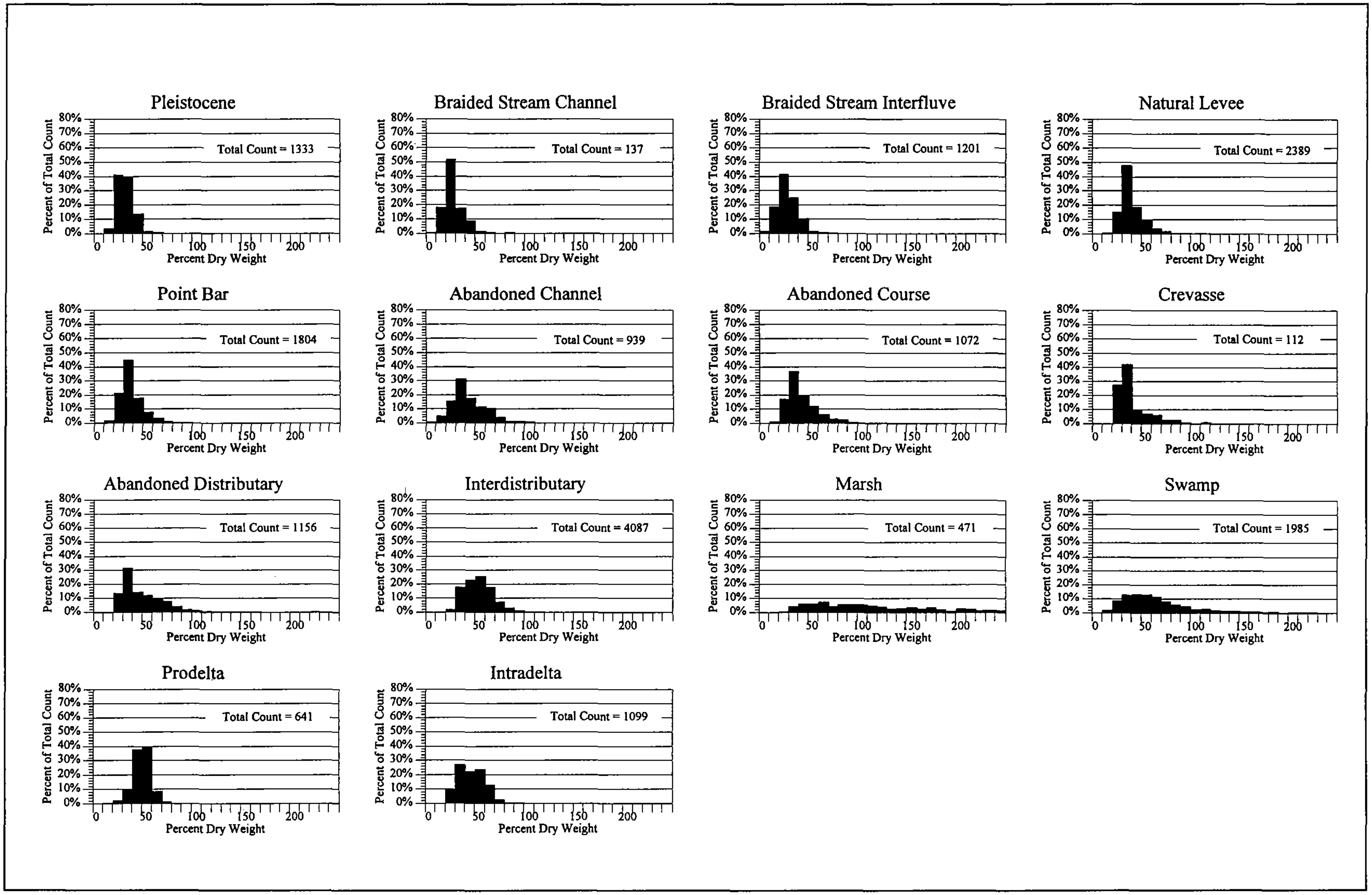

Figure A13. Comparison of water content test data from various Pleistocene and Holocene fluvial and deltaic environments

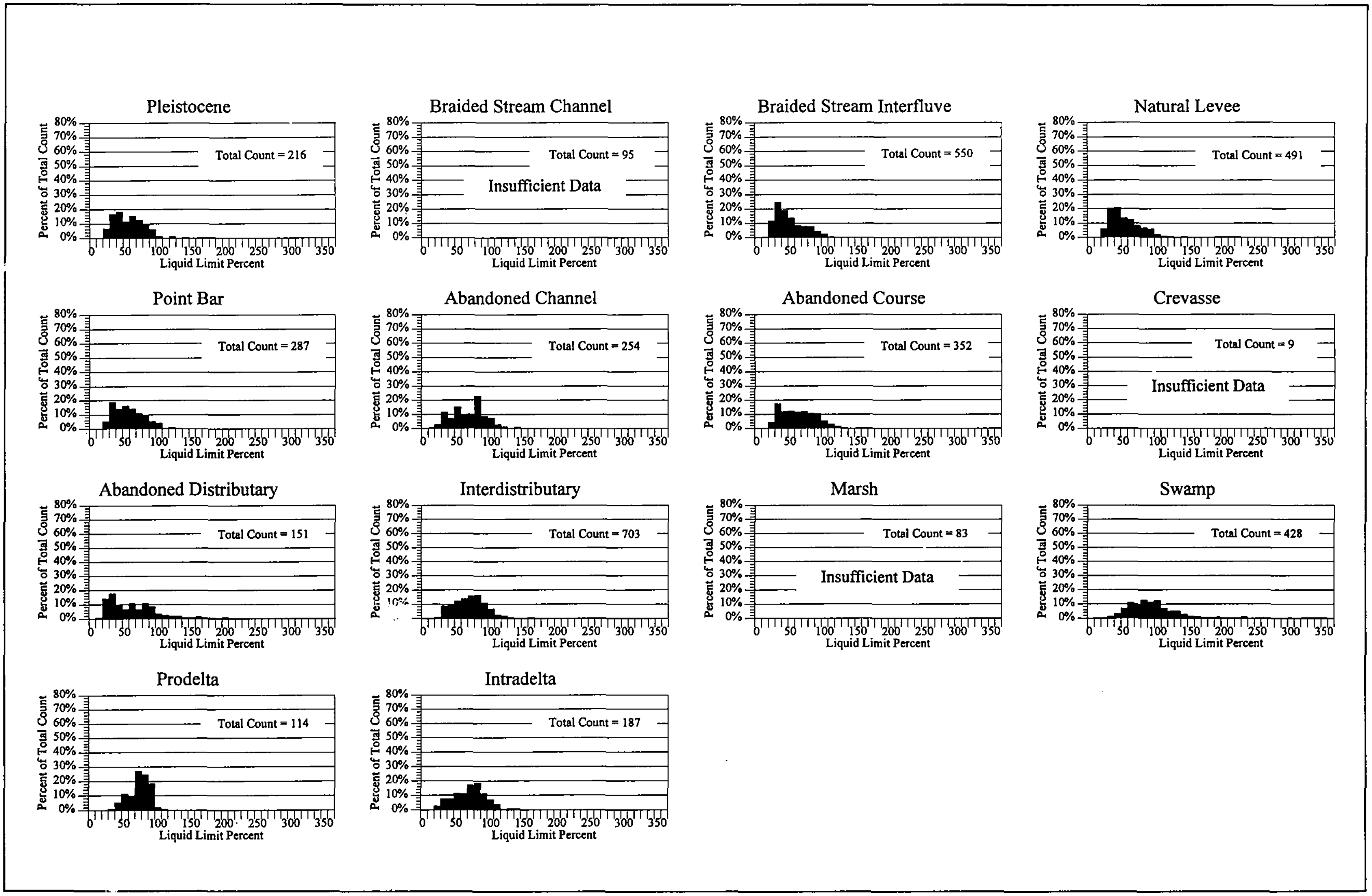

Figure A14. Comparison of liquid limit test data from various Pleistocene and Holocene fluvial and deltaic environments

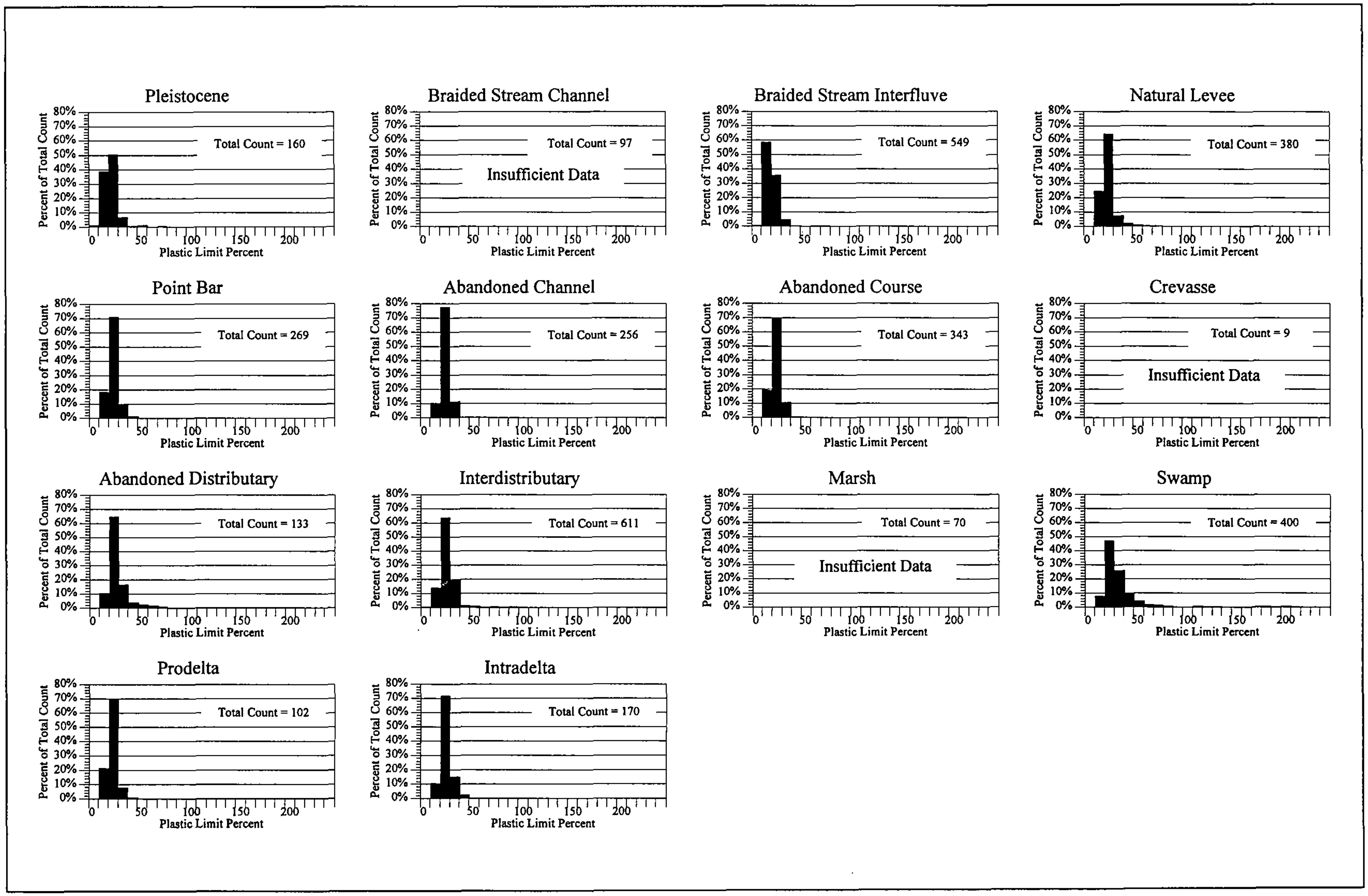

Figure A15. Comparison of plastic limit test data from various Pleistocene and Holocene fluvial and deltaic environments

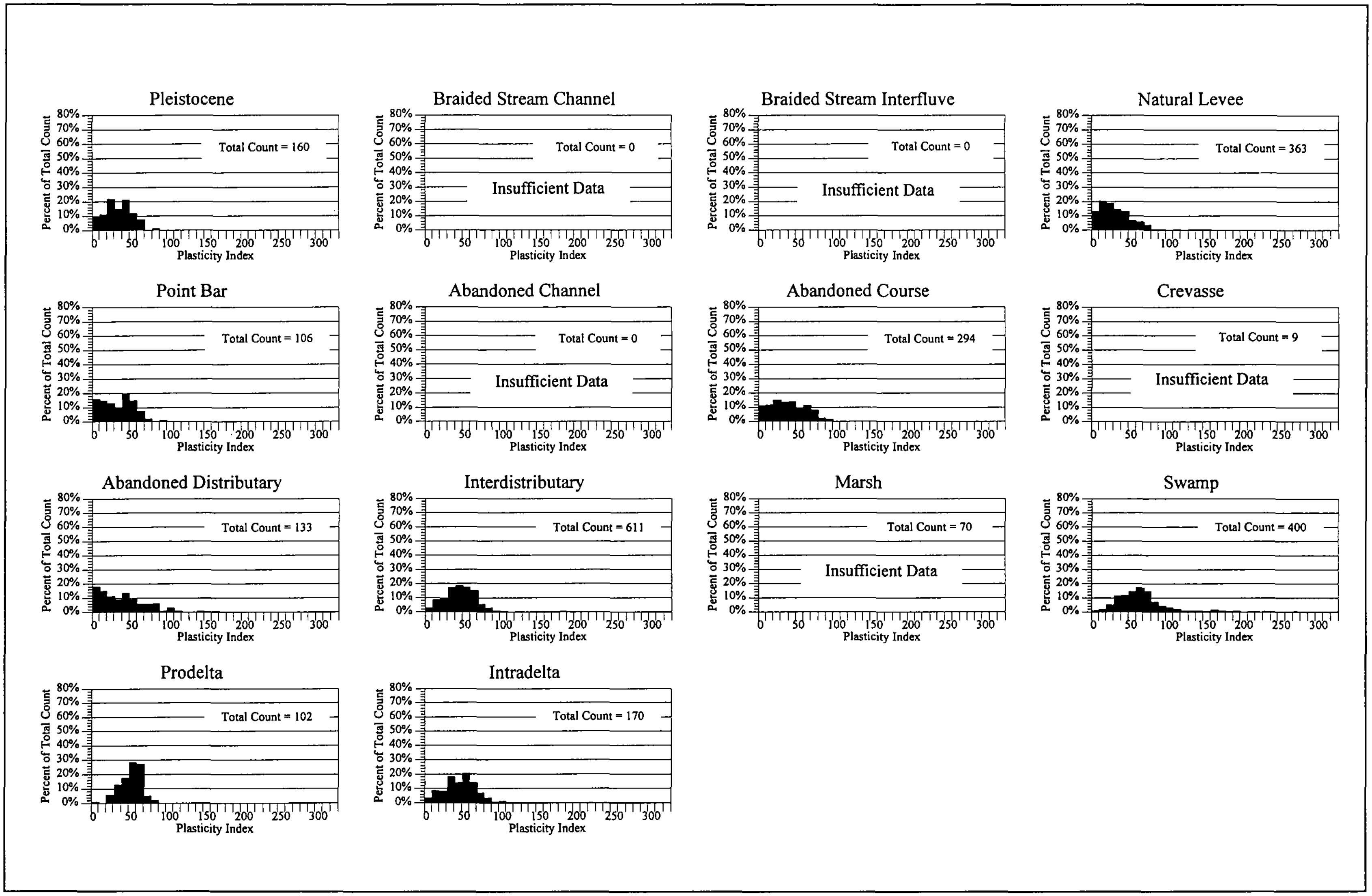

Figure A16. Comparison of plasticity index test data from various Pleistocene and Holocene fluvial and deltaic environments

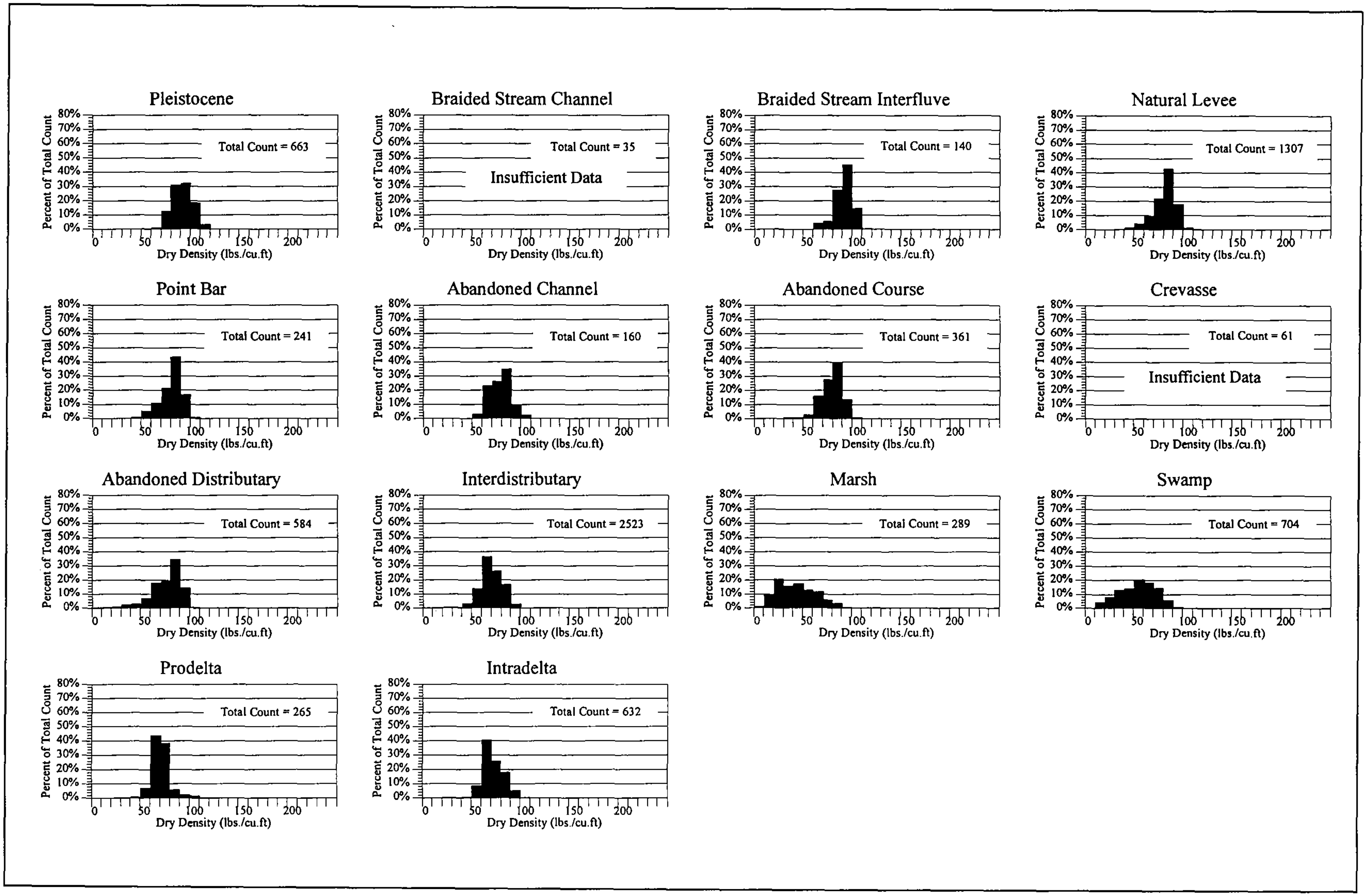

Figure A17. Comparison of dry density test data from various Pleistocene and Holocene fluvial and deltaic environments

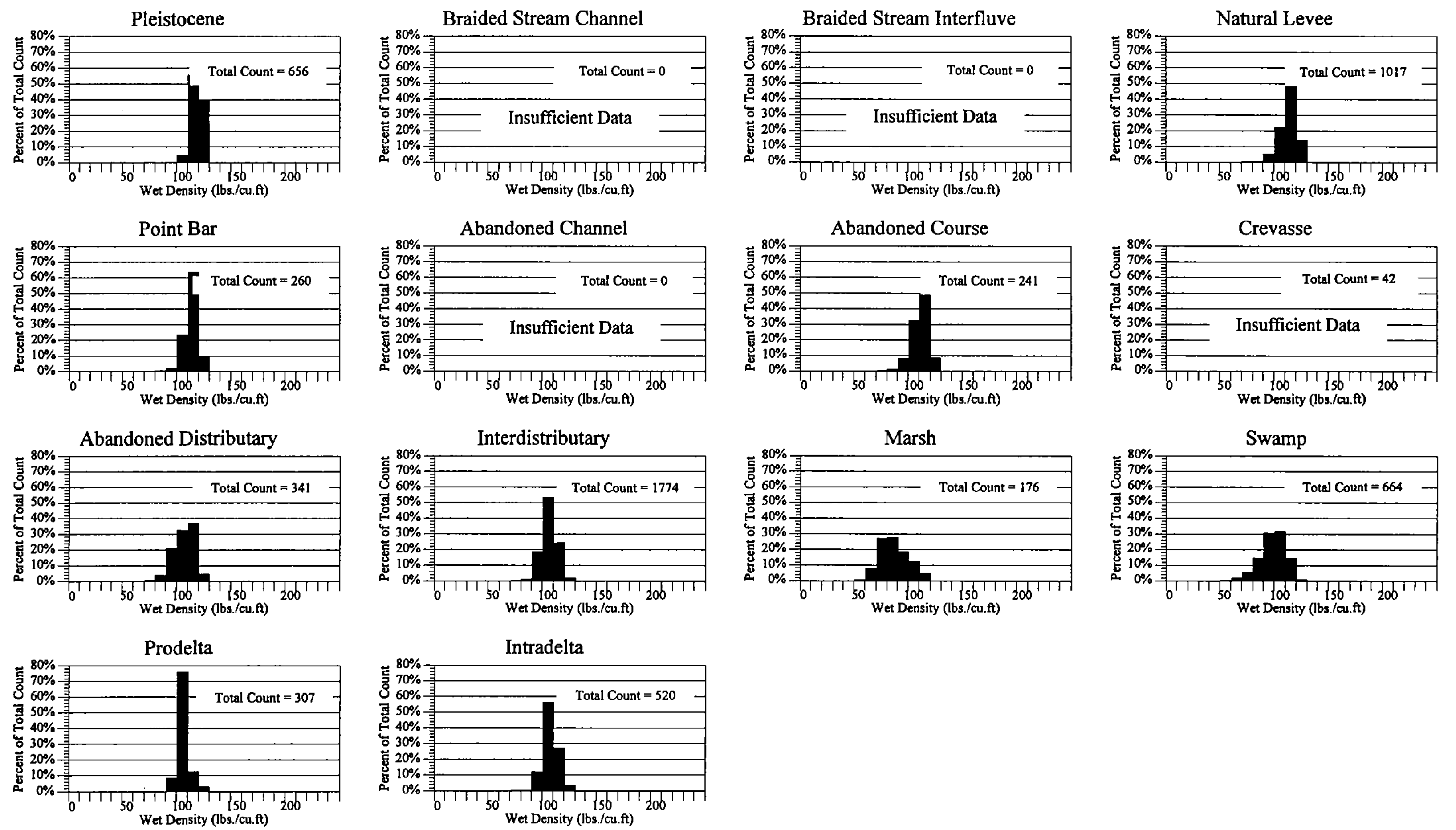

Figure A18. Comparison of wet density test data from various Pleistocene and Holocene fluvial and deltaic environments

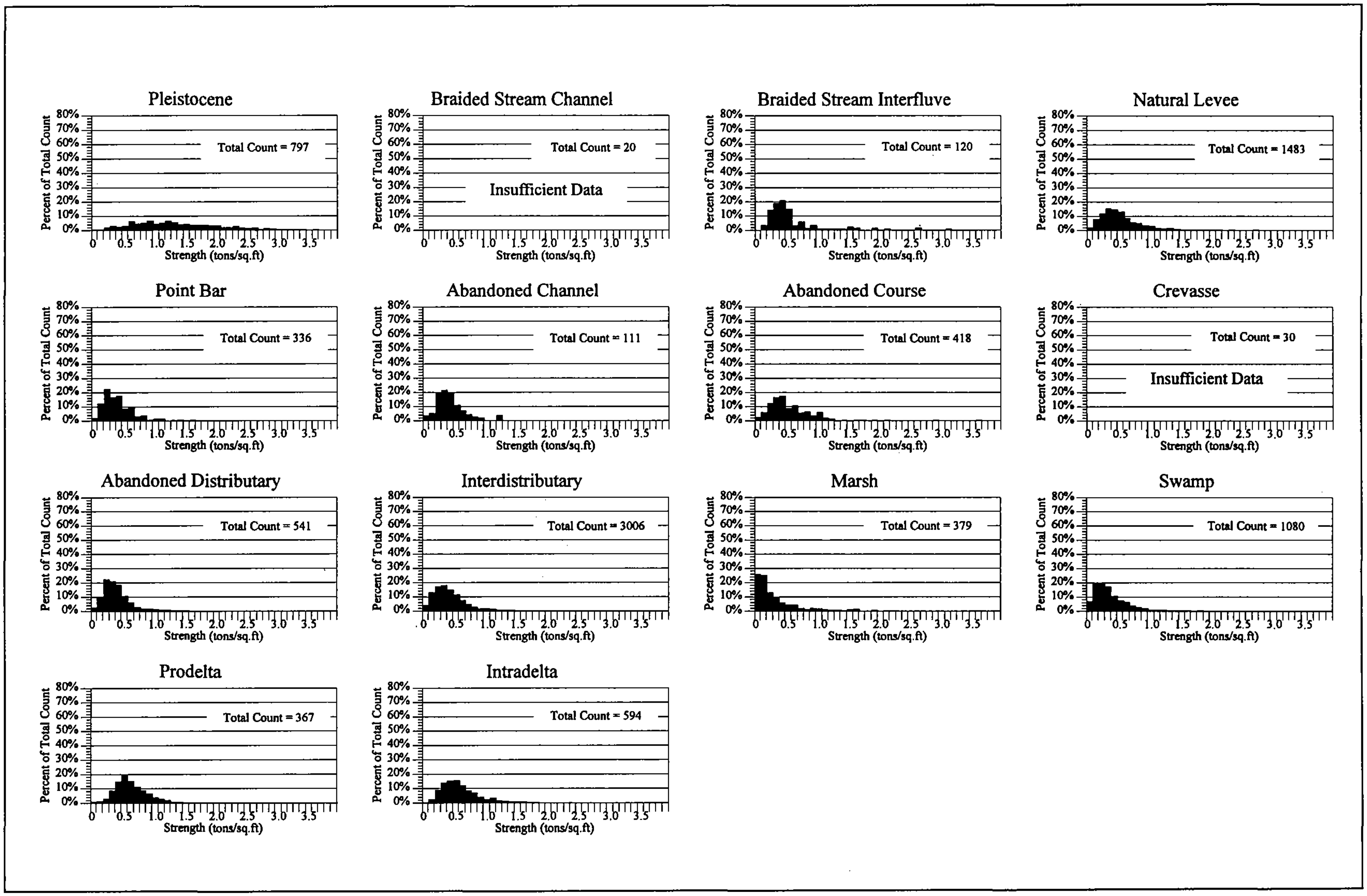

Figure A19. Comparison of cohesive strength test data from various Pleistocene and Holocene fluvial and deltaic environments

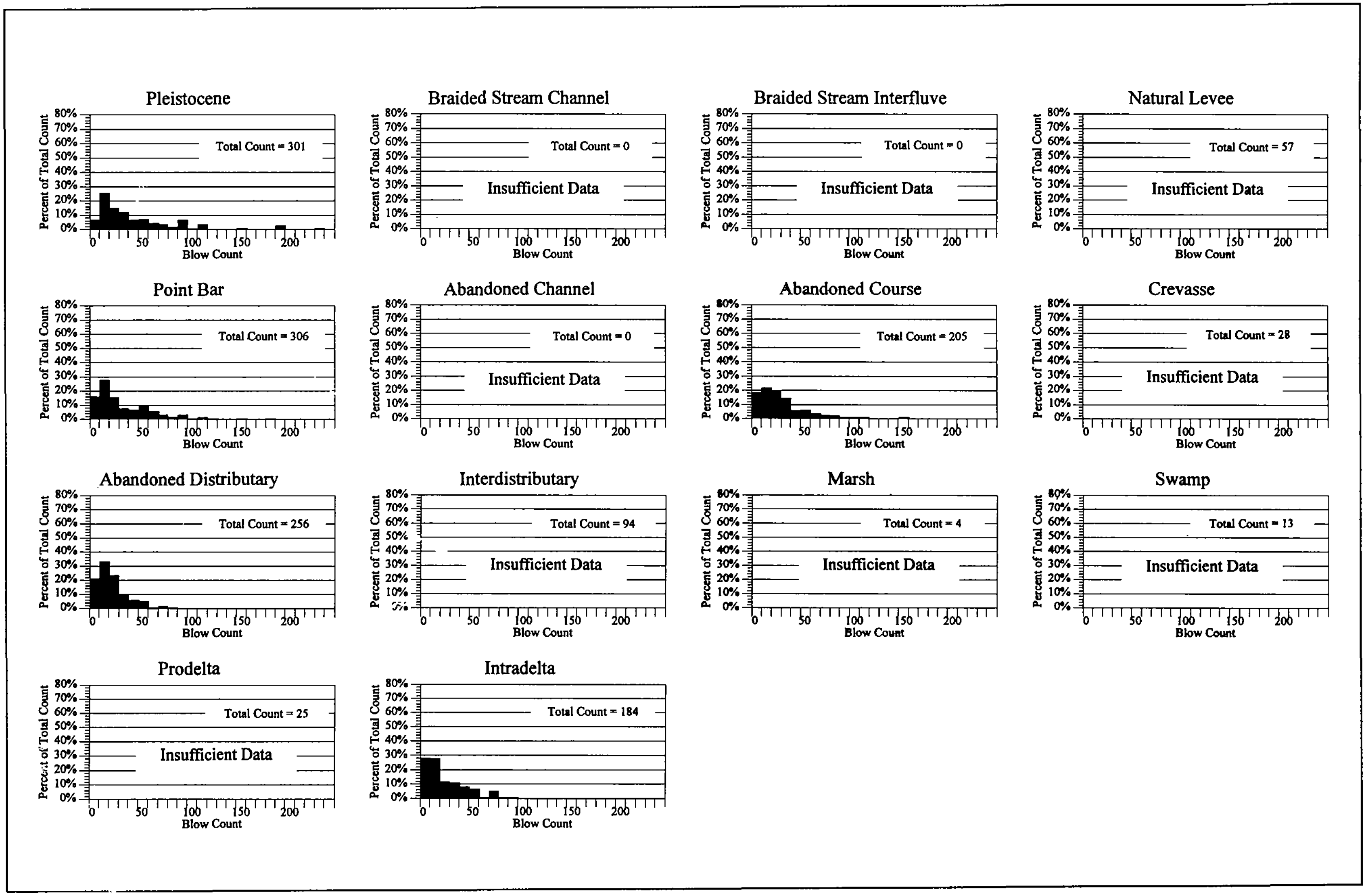

Figure A20. Comparison of blow count test data from various Pleistocene and Holocene fluvial and deltaic environment

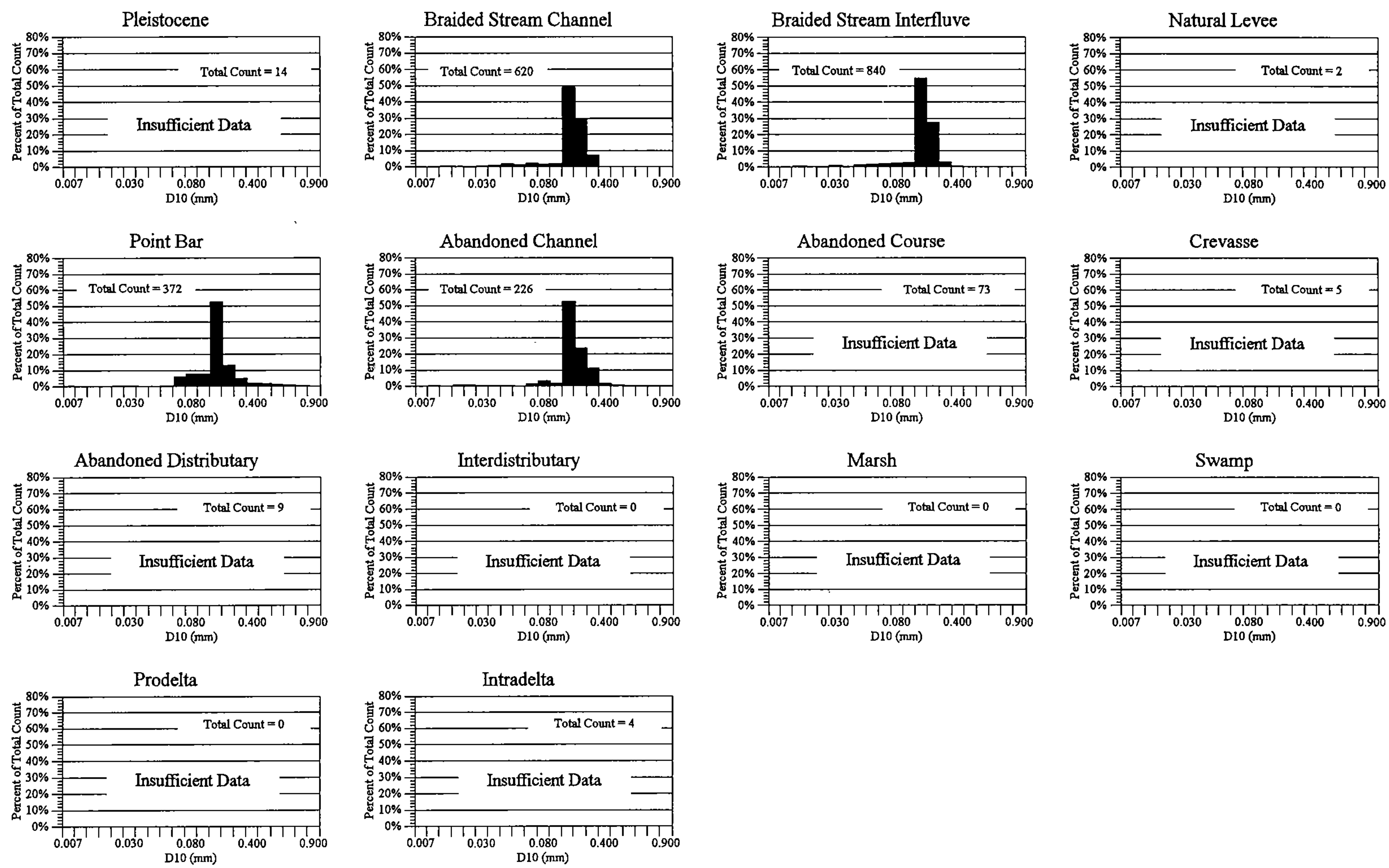

Figure A21. Comparison of D10 grain size test data from various Pleistocene and Holocene fluvial environments

Appendix B
Index

Note: Numerals in *italics* refer to page numbers of text figures and tables.

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13. ABSTRACT *(Maximum 200 words)*

This comprehensive, two-volume synthesis, the first in 50 years, is aimed at a multidisciplinary audience concerned with multiple aspects of water resources engineering and natural and cultural resources management. It presents at a scale of 1:250,000 the distribution of environments of deposition as compiled from more than 30 years of detailed geologic mapping, as well as a new interpretation and delineation of the eroded suballuvial surface. A detailed interpretation of the evolution of the alluvial valley and deltaic plain is presented and illustrated by a series of 13 paleogeographic reconstructions. The chronology of valley events is based on stratigraphic relationships and radiometric age determinations but relies heavily on archeological evidence.

The geologic processes and controls that affect the entire region include continental glaciations, climate, sea level variations, tectonics and diapirism, and subsidence. Both erosional and depositional landscapes are represented, and the lithology, soils, and geotechnical properties of the latter are presented in narrative and tabular form for the principal fluvial, lacustrine, eolian, deltaic, and deltaic-marine environments. Discussions of neotectonics in the region focus on the New Madrid Seismic Zone, and a section of the synthesis addresses special engineering considerations such as groundwater occurrence, mass movements, river meandering, and long-term stability.

14. SUBJECT TERMS

Chronostratigraphy	Geomorphology	Lower Mississippi Valley
Engineering geology	Geotechnical properties	Quaternary geology

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