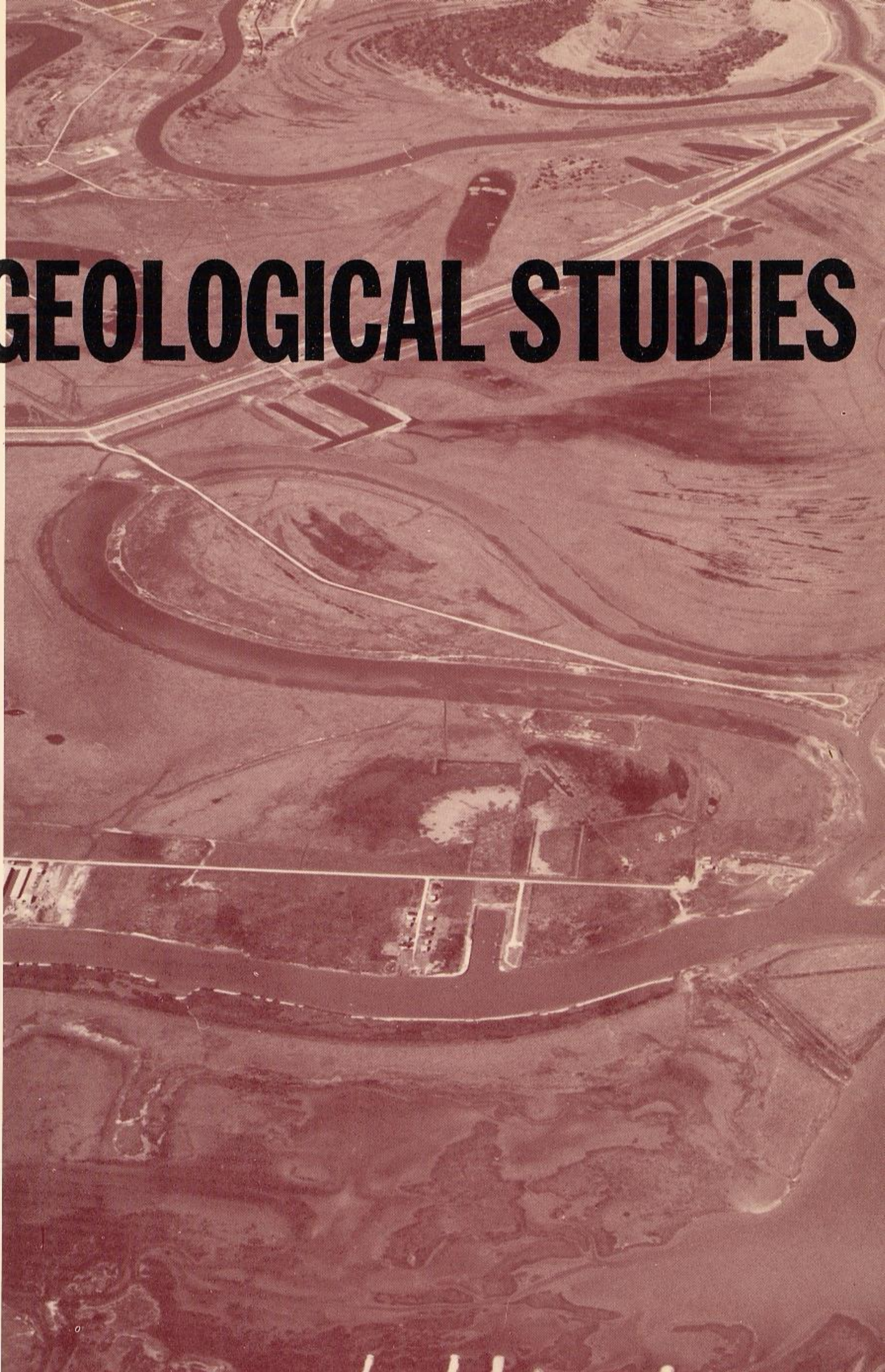
BAYLOR GEOLOGICAL STUDIES



SPRING 1973

Bulletin No. 24

A Geologic History of the Brazos River

LAWRENCE WARD EPPS

"Creative thinking is more important than elaborate equipment--"

Frank Carney, Ph.D.
Professor of Geology
Baylor University
1929-1934

Objectives of Geological Training at Baylor

The training of a geologist in a university covers but a few years; his education continues throughout his active life. The purposes of training geologists at Baylor University are to provide a sound basis of understanding and to foster a truly geological point of view, both of which are essential for continued professional growth. The staff considers geology to be unique among sciences since it is primarily a field science. All geologic research including that done in laboratories must be firmly supported by field observations. The student is encouraged to develop an inquiring objective attitude and to examine critically all geological concepts and principles. The development of a mature and professional attitude toward geology and geological research is a principal concern of the department.

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BAYLOR GEOLOGICAL STUDIES

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A Geologic History of the Brazos River

Lawrence Ward Epps

BAYLOR UNIVERSITY
Department of Geology
Waco, Texas
Spring, 1973

Baylor Geological Studies

The Baylor Geological Studies Bulletin is published semi-annually, Spring and Fall, by the Department of Geology at Baylor University. The Bulletin is specifically dedicated to the dissemination of geologic knowledge for the benefit of the people of Texas. The publication is designed to present the results of both pure and applied research which will ultimately be important in the economic and cultural growth of the State.

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A Geologic History of the Brazos River

Lawrence Ward Epps

ABSTRACT

The Brazos River is a major Texas stream. Meander scars in valley walls are much larger than those of the present river. Quantitative geomorphic analysis suggests that the larger meanders were associated with an ancestral stream having bankfull discharge 5 to 9 times the present river and a bankfull width of approximately 2600 feet.

Deposits of lateral accretion are the dominant depositional units within the modern flood plain. Approximately 2 feet per thousand years is the maximum rate of overbank sediment accumulation for the lower basin flood plain. Present average sediment load of the Brazos indicates that the entire basin is being lowered about 1 foot per thousand years. Comparison of modern point bar deposits to older alluvium within the flood plain suggests that the Brazos River has changed regime since deposition of the older alluvium. This may be related to the larger ancestral river.

At least four levels of alluvial deposits are related to the Brazos River, but only three are well represented throughout the basin. These terrace deposits range from 30 to more than 200 feet above the modern flood plain. Terrace ages are reflected by distinctive soil profiles found on the terraces. The most mature profiles are associated with the third terrace.

The Seymour Formation represents several periods of valley alluviation ranging in age from Sangamon to pre-Kansan. Slope orientation of the erosional surface below the Seymour Formation suggests that the Double Mountain Fork is a captured tributary of the Clear Fork.

High gravel near Throckmorton correlates with the Ogallala Formation thus marking the earlier distribution of this formation. Upland gravel near Lake Possum Kingdom is possibly related to flood plain development before incision of this area. Incision may have been caused by the capture of the Clear Fork from the Leon River in early Pleistocene.

Statistical analysis of gravel composition from the channel, first terrace, and second terrace indicates significant differences between levels which allows this criterion to be used as a correlating tool. The differences reflect changes in provenance as the Brazos basin evolved.

Large cobbles found in older alluvium suggest a river of greater competence than the present river. The tractive force needed for movement of these cobbles would require a river both deeper and steeper than the modern Brazos near Bryan, Texas. This supports the evidence from the meander scars.

The Brazos River may have been established as early as Eocene time, but definitely was established by Miocene. Pliocene deposits of the Ogallala and Goliad Formations indicate widespread arid conditions before Pleistocene time. This arid period was followed by deposition of the Willis Formation which represents a transitional unit between late Tertiary deposition and the multiple cycles of valley alluviation during Pleistocene time. Streams that deposited the Willis Formation had greater width-to-depth ratios and were steeper than streams that deposited the Brazos River terraces. The Willis and Uvalde Formations formed a broad alluvial blanket over much of the Cretaceous terrane during earliest Pleistocene, which explains the lack of limestone as a common component in gravel of the third terrace in the lower basin.

Eustatic sea level changes did not cause terrace development. The primary cause of terrace development was climatic changes associated with glacial stages. These climatic changes controlled river regime by influences exerted on sediment supply and runoff.

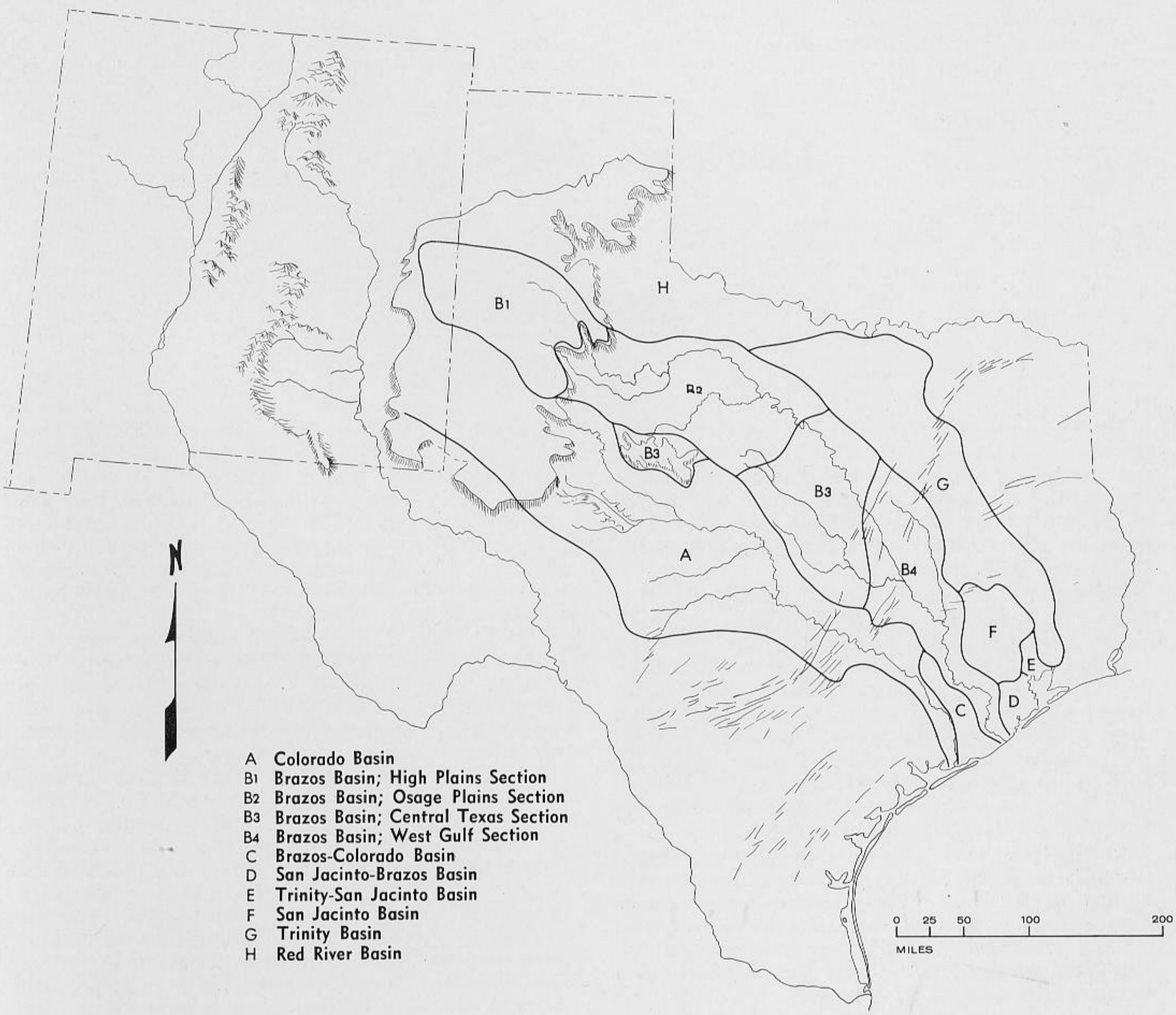

Fig. 1. Index map showing river basins in Texas and physiographic sections of the Brazos River basin.

INTRODUCTION*

LOCATION

The Brazos River basin, largest in the state of Texas, covers about 15 percent of the state. The drainage area totals 44,000 square miles, but about 9,000 square miles on the high plains are considered noncontributing (Irelan and Mendieta, 1964, p. 7). From the Brazos River headwaters on the High Plains of eastern New Mexico to the river mouth at Freeport, Texas it flows more than 1200 miles (about twice the airline distance between source and mouth) and drops 4600 feet. Divides separate the basin from that of the Colorado River on the southwest and the Trinity and Red Rivers on the northeast (fig. 1).

PURPOSE AND PROBLEM

The purpose of this investigation is to interpret the history of the Brazos River since it became established in approximately the present location. Study of drainage maps and field relationships suggests major problems associated with Brazos River history. Geomorphic features and sediment character suggest major changes in river regime. Terrace gravels include lithologies foreign to the present basin. These gravels are found in the terraces of the Leon and Navasota Rivers. Terrace sequences and lag deposits of gravel are evidence of the antiquity of the river. Geologic structure and stratigraphy commonly controlled the development of the drainage net. Soils, terrace deposits, and abandoned valleys suggest important drainage diversions.

METHOD

This study is based on analysis of data obtained from field reconnaissance, flow records of the modern Brazos River, and analysis of topographic maps. Maps and flow records provided data for an analysis of stream hydraulics and meander geometry. Cross-valley traverses were made at most of the bridges crossing the Brazos River and its major tributaries, and longitudinal traverses were made paralleling the river on its divides. This procedure gave a general knowledge of type and distribution of alluvium and terraces. Point bars were trenched to determine the size range of the river load and nature of typical sedimentary structures. Numerous gravel pits in both the modern floodplain and older alluvium were studied to determine changes in river load through time. Random samples were obtained of channel bed load and terrace gravel at locations between Seymour and Hempstead for use in a statistical analysis to determine whether the gravel composition of the modern Brazos differs from that of its terraces.

*A thesis submitted in partial fulfillment of the requirements for the M.S. degree in Geology, Baylor University, 1972.

PREVIOUS WORK

There have been few large-scale studies of the history of the Brazos River. References to Brazos deposits are usually found as part of larger reports on paleontology, county geologic surveys, and industrial mineral surveys.

Hill (1901, p. 357), one of the first to publish about Brazos River deposits, included them in a discussion of major river deposits in central Texas. He recognized the multilevel alluvium of the Brazos River and named the "upland alluvium" in the Black and Grand Prairie subprovince the "Uvalde Formation," which he considered Neogene in age.

Deussen (1924, p. 114) described the terraces of the lower Brazos River in some detail. He recognized six levels ranging from 30 to 310 feet above the river. His presentation included several valley profiles emphasizing paired terraces.

Brazos deposits in central Texas are well known because of numerous gravel pits and the wealth of vertebrate remains which have been recovered from them. In the McLennan County area terraces were described by Pace (1921), Adkins (1924), Deussen (1924), Bronaugh (1950), Burket (1965), Hopkins (1961), Taylor (1962), and Stricklin (1961). Of these, the study by Bronaugh was the most extensive. He grouped the terraces into two groups, a lower group including levels from 20 to 50 feet above the river, and a high group including levels from 70 to 200 feet above mean low water. Because his collections of fossils from the high terraces contained no Blancan (early Pleistocene) species, he concluded that the oldest terraces are late-early to early-middle Pleistocene in age. The Pleistocene fauna persists in the lower terraces down to and including the 50-foot terrace.

Cox (1950) and Bernard, LeBlanc, and Major (1962) recognized four Quaternary surfaces of the coastal plain, equivalent to paired Brazos River terraces which extend inland up the Brazos valley. Fisk (1944), the mentor of the previous authors, had shown a similar relation for the lower Mississippi River and considered the phenomenon glacially controlled. Thus each plain and equivalent terrace was correlated with an inter-glacial stage of alluviation.

Descriptions of terraces in the northern portion of the state are found in two county reports. Hendricks (1935, p. 49) described terraces in Parker County. Plummer and Hornberger (1935, p. 218) found three terrace levels in Palo Pinto County, the lowest terrace 50 feet above river level, the middle 90 to 120 feet, and the highest 210 feet above valley bottom. The high terrace was considered early Pleistocene or Pliocene and was thought to contain more quartz and chert than lower levels. The middle terrace was dated as late Pleistocene on the basis of mammal remains.

Study of the middle Brazos River between Seymour and Waco, Texas, by Stricklin (1961) produced two important age determinations. (1) He found Pearlette Ash of Kansan or early Yarmouthian age interbedded with deposits of the high terrace near Seymour. (2) He found a molluscan fauna in the same area which substantiated this date. He also suggested the probable capture by the Wichita River of drainage previously belonging to the Brazos River.

Lewand (1969), in a study of the Leon River (a major tributary of the Brazos River), suggested that the Leon valley was once occupied by Brazos drainage. Terrace soils and terraces of the present Leon valley show affinities with similar soils and terraces of the Brazos River. Quartzite cobbles from the terraces were correlated with similar lithologies that crop out in the Manzano Mountains of New Mexico, yielding evidence of caprock drainage in the former Leon basin. Petrographic comparison of metaquartzites from the divides near Dallas, Texas, to outcrops in the Manzano Mountains showed similarity in composition and thus support the provenance suggested by Lewand (Menzer and Slaughter, 1970, p. 292).

Recently, the "Uvalde Formation" of Hill has been regionally studied by Byrd (1971). He concluded that the gravels (Uvalde) capping the divides of the Brazos River represent alluviation of major drainage courses extending eastward from the original margins of the High Plains (Ogallala) during Pliocene time. Their present high elevation is a result of a reversal of topography related to the armoring effect of the coarse siliceous gravel.

ACKNOWLEDGEMENTS

The writer expresses his gratitude to those who contributed moral and physical support to this study. O. T. Hayward, Department of Geology, Baylor University, guided the study and aided it by discussion of field evidence and manuscript revision. R. L. Bronaugh, Department of Geology, Baylor University, critically reviewed the manuscript and advised the author. Howard Rolf, Department of Mathematics, Baylor University, provided advice and assistance on computer applications. Thanks go to the Texas Water Development Board for providing topographic maps of the drainage basin, to the United States Geological Survey Hydrologic Branch for providing copies of field measurements of flow records and numerous useful publications, to G. C. Mick of Micro Incorporated for providing aerophotography of the Brazos River delta, and to Al Redder for contributing radio carbon dates from the Brazos floodplain deposits. Appreciation is also extended to Thomas Maddock, Jr., assistant chief of the Hydrology Division, Bureau of Reclamation, for encouragement to continue the study of a large river basin and to Earle F. McBride and Victor R. Baker, Department of Geological Sciences, The University of Texas at Austin, who kindly reviewed the manuscript before publication.

EVIDENCE OF RIVER HISTORY
PRESENT RIVER

INTRODUCTION

To understand the history of a river system, one must be familiar with the dynamics of the present stream system. This understanding of present river processes forms a basis for conclusions about the ancestral river.

River hydraulics has become a strong tool for geomorphologists of the past three decades. Data derived from maps and flow records are analyzed to determine empirical relationships that might be applied to the problem. This "quantitative empirical method" has been little used in prior Brazos River studies.

Bates (1939) listed measurements from four Brazos River localities in an effort to relate width and meander-belt width (twice the amplitude). He generalized that the ratio of stream width to meander-belt width was possibly linear for streams of a given type (incised, etc.). Recent work has shown that the relationship is nearly linear (Table 1).

Goines (1965) studied flow characteristics of the Brazos River to determine the frequency of any specific flow at a particular station. This data is used to plot flow duration curves which describe long term discharge characteristics at a recording station.

Proctor (1967) applied Horton's morphometric analysis to the North Bosque River, a major tributary of the Brazos. His purpose was to determine basin stage, stability of stream geometry, and effect of geology on geomorphic evolution.

Lewand (1969) studied the meander geometry of the Leon River. He concluded the river was underfit thus suggesting major diversion of the Leon River's prior drainage.

MEANDER GEOMETRY

Since the beginning of this century geomorphologists and hydrologists have attempted to establish fundamental relationships between measurements of a natural channel and its corresponding meander dimensions (Table 1). A review of the equations shows that meander amplitude and wave length are consistently related to the square root of discharge. The relationships of meander wave length and amplitude to channel width and mean radius of curvature are essentially linear, though the coefficients vary considerably between similar equations. King (1966, p. 90) believed this variation in coefficients to be a function of the bed condition, load, seasonal discharge variation, and nature of bank material. Thus the coefficients are unique for each river. Leopold, Wolman, and Miller (1964, p. 297) generalized that coefficients relating meander length to channel width range from 7 to 10. Regardless of the source of data from which the relationships were derived, the coefficients are of the same order of magnitude (the equations of Leopold *et al.* are probably the best for American rivers).

Leopold and Wolman (1960, p. 774) equated channel width and mean radius of curvature (from their

TABLE 1 — MEANDER GEOMETRY EQUATIONS

Source	Meander Length (L) to Channel Width (W)	Amplitude (A) to (W)	(A) to Discharge (Q)	(L) to Radius Curvature (R)	(L) to (Q)
Inglis (1949, Jefferson's data)	$L = 6.6W^{.99}$	$A = 17.4W$	$A = 84.7Q^{.5}$		$L = 28Q^{.5}$
Inglis (1949, Bates data)		$A = 14W$			
Leopold and Wolman (1960)	$L = 10.9W^{1.01}$	$A = 2.7W^{1.1}$		$L = 4.7R^{.98}$	
Foweraker and Brice (1963, first order meanders)	$L = 34.2W^{.89}$	$A = 8.3W^{.86}$	$A = 13.2Q^{.51}$		$L = 58Q^{.52}$
Foweraker and Brice (1963, second order meanders)			$A = 24.2Q^{.5}$		$L = 199Q^{.49}$

original relationships of meander length to width and meander length to radius of curvature). Assuming the exponents to be unity and solving for the ratio of radius of curvature to width: $R/W = 10.9/4.7 = 2.3$ where R is radius of curvature and W is channel width. The investigators then calculated the ratios for 50 rivers of varying size and province. The mean value of their findings was 3.1. This value also bears an important relationship to resistance of flow, which indicates the hydraulic basis for the observed geometric similarity among channels of different sizes (Leopold and Wolman, 1960, p. 774). The ratio is similar for model streams in flumes, ice channels, submarine channels, canals, and natural streams. Bagnold (1960, p. 135) noted that the resistance of a liquid flowing through a pipe bend is minimal when the curvature-ratio is between 2 and 3.

Schumm (1963, p. 1089) divided meandering river patterns into three classes, determined by an index of sinuosity (P), defined as the ratio of channel length to valley length (Table 2).

TABLE 2 — SINUOSITY CLASSIFICATION

Class	P
Tortuous	2.1
Irregular	1.7
Regular	1.5

The sinuosity of the Brazos River (calculated from the Army map service Austin sheet 1:250,000) is 1.8. The curvature ratio of this irregular meandering pattern describes the meander geometry of the Brazos River within the coastal region, calculated by measuring the width and mean radius of curvature at bankfull stage. (Bankfull width was assumed to be the distance from a cut bank to the top of the adjacent point bar.) Measurements were taken from 1:24,000 topographic maps at 21 localities (Appendix A). The mean curvature ratio for the Brazos River is 2.8, thus the radius of curvature is 2.8 times the channel width or $R = 2.8W$ where R is the radius of curvature and W is the bankfull channel width.

A series of curved escarpments flanks the Brazos River from approximately 50 miles above its mouth to Waco, Texas. The escarpments are similar in appearance to cut banks of the present Brazos River in a meandering reach, except that they are of greater height and radius of curvature. The scarps rise 30 to 100 feet above the modern flood plain and most are topped by terrace deposits. A representative example of these curved escarpments can be seen east of the river on the Hearne South quadrangle (U.S.G.S. 7½ minute quadrangle). (Normally the scarp face is dissected by youthful drainage.)

These escarpments are probably meander scars belonging to an earlier cycle of erosion, cut by a river larger than the present Brazos River. The value of the mean radius of curvature can be approximated for the large scars by swinging an arc of best fit on several map examples and averaging radius measurements. The average for 16 meander scar measurements is 8680 feet (Appendix B).

This value is a maximum figure because it is measured from the cutbank of the postulated meander bend. The relationship between the maximum radius of curvature and the mean is (1) $R_{max} - R_{mean} = .5W$ where W is channel width. If the assumption is made that the previously derived relationship between mean radius of curvature and channel width, $R = 2.8W$, is applicable to the ancient Brazos River then by substitution into equation (1) the relationship between maximum radius of curvature and width can be developed: (2) $R_{max} - 2.8W = .5W$ or $W = R_{max} /3.3$. The solution of equation (2) by inserting the average radius of curvature for the meander scars into the expression gives a quantitative statement about the dimensions of the ancient Brazos River. The channel width responsible for the observed meander scars was approximately 2600 feet.

An independent method of determining the width of the ancient Brazos River is by use of equations in Table 1 (Leopold and Wolman, 1960, p. 769). One set of arcs, on the Hearne South 7½ minute quadrangle, is preserved well enough to measure a wave length and amplitude of a full meander. The length is 26,400 feet and amplitude is 10,000 feet. Inserting these values in the equations relating them to channel width: W = 2240 feet and W = 1760 feet. Results of the wave length to width method correlate reasonably well with the radius of curvature method. They differ by only 14 percent. The amplitude value correlates poorly (error equal 35 percent). Since amplitude is not only a function of hydraulics but also is influenced by the erosion characteristics of stream banks (Leopold and Wolman, 1960, p. 773), this is not too surprising.

"The most meaningful discharge for any discussion of channel morphology is that which forms or maintains the channel. This discharge can often be approximated by bankfull discharge. In many rivers the bankfull discharge is one that has a recurrence interval of about 1.5 years" (Leopold *et al.*, 1964, p. 241). Thus the 2600-foot channel width probably represents the bankfull stage.

HYDRAULIC GEOMETRY

Leopold and Maddock (1953, p. 1) use the term "hydraulic geometry" in reference to the fundamental relationships between channel dimensions and discharge. In their study, they graphed the variations of water surface width, mean depth, and mean velocity at a point as discharge changed. The results showed that each of the parameters varied as a power function (Table 3). These empirical relations are called "at a station" relations. By plotting the stream data as log coordinates the result is a straight line: Log W=b Log Q + Log a, etc. A best fit line drawn through the points will give the values of the constants for each equation.

TABLE 3

$$W = aQ^b$$
$$D = cQ^f$$
$$V = kQ^m$$

where: W = width of water surface; D = mean depth of water; V = mean velocity; Q = discharge in second feet; a, c, and k are constants; b, f, and m are exponents.

Similar relationships exist for cross sectional dimensions proceeding downstream providing the measurements are taken at a discharge of equal frequency for each station. These relationships at a set frequency are called "downstream" relations. The frequency of discharge at a station is determined statistically (Goines, 1965).

Because width, depth, and velocity are each a function of discharge, the three equations in Table 3 can be related one to another:

$$Q = \text{area} \times \text{velocity, or, } Q = WDV$$

substituting from Table 3

$$Q = aQ^b \times cQ^f \times kQ^m$$
$$\text{or } Q = ackQ^{b+f+m}$$
$$\text{therefore } b + f + m = 1$$
$$\text{and } a \times c \times k = 1$$

The constants b, f, and m represent slopes of the plotted lines. The constants a, c, and k are values of W, D, and V at a discharge of unity. For any given discharge the width and depth will vary widely from one cross section to another, therefore the intercept values will vary. However, the slopes show some consistency as can be seen from the average values in Table 4 (after Leopold *et al.*, 1964, p. 244).

TABLE 4

	At a Station Relations			Downstream Relations		
	b	f	m	b	f	m
Midwest U.S.	.26	.40	.34	.5	.4	.1
Brandywine Cr. Pennsylvania	.04	.41	.55	.42	.45	.05
Semiarid U.S.	.29	.36	.34	.5	.3	.2
Appalachians				.55	.36	.09
158 stations U.S.	.12	.45	.43			

The Brazos River "at a station" empirical relations were calculated for Bryan, Hempstead, and Richmond. These long record gaging stations provided 520 measurements for Bryan, 65 for Hempstead, and 644 for Richmond. Table 5 and figure 2 give a summary of the results.

TABLE 5 —AT A STATION RELATIONS

Location			
Bryan	$W = 61.1Q^{.181}$	$D = .148Q^{.451}$	$V = .113Q^{.368}$
Hempstead	$W = 48.2Q^{.205}$	$D = .117Q^{.493}$	$V = .191Q^{.294}$
Richmond	$W = 102Q^{.135}$	$D = .088Q^{.519}$	$V = .163Q^{.300}$
Average	$W = 70.4Q^{.174}$	$D = .118Q^{.488}$	$V = .156Q^{.321}$

The exponents for the Brazos River equations correlate with those average values given in Table 4. Generally it can be seen that the width rate-of-change is less than either the velocity or depth changes at increasing flow. As discharge increases, the width-to-depth ratio decreases, which is generally true for all rivers with elliptical to trapezoidal cross sections.

Records from eleven gaging stations along the main Brazos River channel were used to define the change of channel parameters in a downstream direction at a particular frequency of flow (Appendix C), mean annual discharge, which occurs approximately 20 percent of the time. Table 6 lists the equations and figure 3 shows the graph of the values.

TABLE 6 — DOWNSTREAM RELATIONS

$$W = 12.6Q^{.39} \qquad D = .08Q^{.51} \qquad V = 1.1Q^{.10}$$

The equations show that as discharge increases downstream the river cross section is adjusted by faster deepening than widening. Velocity increases downstream (fig. 3). Intuitively one would expect velocity to decrease as slope decreases. However, a modification of the Manning equation ($V = cD^{.66}S^{.5}$) shows that mean velocity is proportional to mean depth as well as slope and because the Brazos River's depth is increasing downstream the velocity is also increasing.

The real value of downstream relations is their application as estimators of ancestral Brazos River hydraulic parameters. The bankfull discharge of the ancestral Brazos, 861,000 cfs, can be approximated by inserting the river width value determined from meander geometry (2600 feet) into the width-discharge relation. The other downstream relations can be solved using this discharge: $D = 85$ feet and $V = 4.3$ ft/sec. These values describe the ancestral Brazos River at bankfull stage.

Calculation by this means raises several questions. The equations have been extrapolated far beyond the upper limit of the data from which they were derived. As one determines values further from the mean of data values the confidence limits expand rapidly. It will also be recalled that the line equation is a mean annual flow relation but the width value is for bankfull stage. The frequency of bankfull flow is less than 1 percent while mean annual flow occurs about 20 percent of the time. The effect of lower frequency of flow on the width-discharge line tends to decrease the value of the y-intercept but the slope remains essentially the same.

Little data is available for bankfull flow; however, calculations for a frequency of 1 percent indicate that bankfull discharge may have been larger than 861,000 cfs by a factor of 2.6. A better estimate could be obtained using the mean annual flow relation with a mean annual width corresponding to the ancestral Brazos River. An average relation between bankfull and mean annual width for the modern lower Brazos is $W_m = W_b/1.6$. This relation can be applied to the ancestral Brazos assuming that the cross-sectional geometry was similar to the present river. Calculation of mean annual width gives 1625 feet (6 times larger than Brazos at Bryan). Corresponding mean annual discharge would have been 257,000 cfs. Velocity and depth of this river would have been 3.8 ft/sec and 53 feet respectively.

Comparative figures from the Mississippi at mean annual flow values follow: at St. Louis discharge = 166,700 cfs, width = 1586 feet, velocity = 3.8 ft/sec, and depth = 28 feet; at Vicksburg discharge = 554,600 cfs, width = 2610 feet, velocity = 5.3 ft/sec, and depth = 40 feet.

Another estimate of bankfull discharge can be obtained from the equation $L = 30Q^{.5}$, where L is meander wave length (Dury, 1965, p. 5). Using the earlier described wave length of 26,000 feet in the equation gives a discharge value of 774,000 cfs. A final method was suggested by Thomas Maddock of the Bureau of Reclamation (1969 oral communication). Bankfull at Bryan, Texas today is 110,000 cfs. The width at this stage is 500 feet. Therefore the average flow per unit width is 220 cfs. Multiplying this factor times the postulated bankfull width of the ancestral Brazos gives a discharge of 572,000 cfs.

Values of paleodischarge for the Brazos River's lower basin range from 570,000 cfs to 860,000 cfs for bankfull stage and may possibly have been as great as 2,000,000 cfs, but the former figures are more probable. This range is 5 to 8 times greater than modern bankfull discharges. Maximum channel dimensions suggested by discharge relationships are 2600 feet wide and 85 feet deep, while velocity is 4.3 ft/sec. It is more probable that velocity was greater and depth was less. For example, if velocity is increased to 5.3 ft/sec then depth becomes 62 feet ($Q = WDV$). The postulated depths would represent bottom scouring at large bankfull flow.

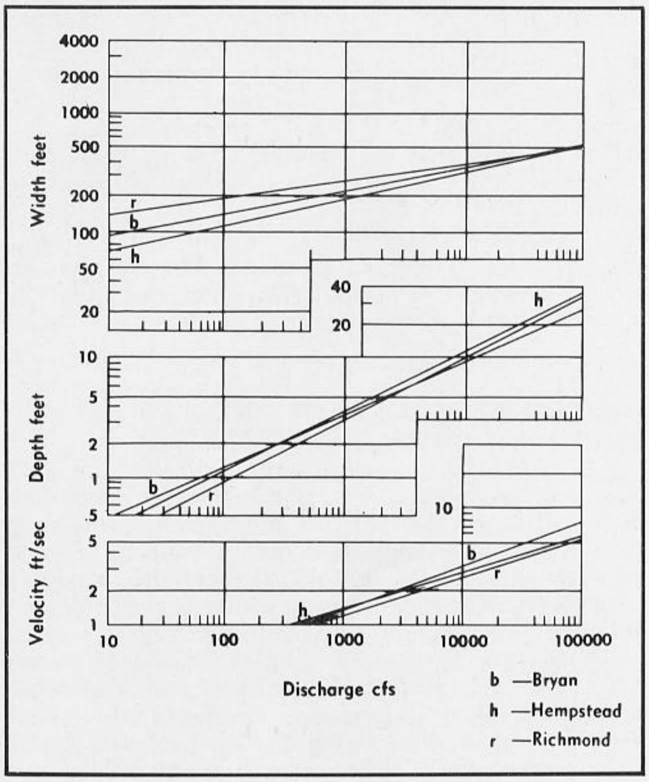

Fig. 2. At a station relations.

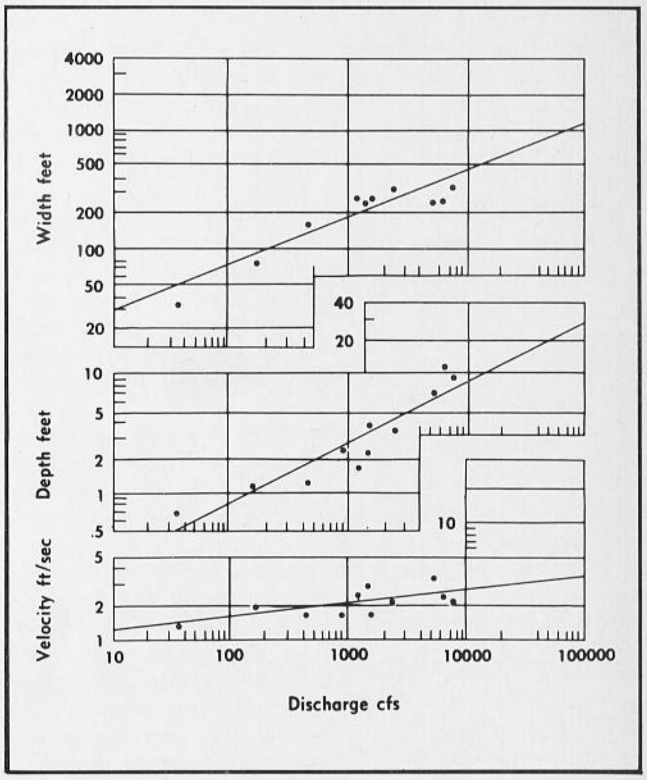

Fig. 3. Downstream relations.

EVIDENCE OF RIVER HISTORY ANCESTRAL BRAZOS

INTRODUCTION

The most direct evidence of the existence of an ancient Brazos River is contained in the terraces and high-level gravels preserved in and adjacent to the present valley. These not only give evidence of a previous river by their presence, but contained sedimentary structures, gravel composition, and grain size yield evidence of river character, basin geology, and river competence at the time of alluviation. For these reasons a survey of the distribution, geometry, and structure of the Brazos River's flood plain and each major terrace is imperative to an interpretation of basin history.

Terraces are remnants of former flood plains. Recognition of terraces is of primary importance in any interpretation of river history. Physiography, sediment type, and soil type are the three basic criteria used for terrace recognition in this study.

A terrace is composed of two parts, the scarp and the level surface above and behind it. These features are rarely preserved on higher or older terraces. Sheet erosion and stream dissection combined with colluvial deposition tend to destroy or conceal terrace scarps and flats or cause two or more terraces to merge. However, distinction may be accomplished by use of soil and sediment distribution.

Occasionally a combination of alluviation on the modern flood plain and erosion of the adjacent terrace allows the older deposits to be flooded by the modern river during maximum stages. Well data is needed to resolve problems of this nature.

Rock-defended terraces are developed over a basal stratum of resistant bedrock. Commonly these terrace deposits are stripped and the only evidence of their previous existence is a broad flat area with scattered lag gravel. This gravel terrace expression is commonly found in regions where the Brazos River flows over limestone.

Another evidence for prior existence of a terrace is the presence of a lag gravel at the base of soils developed on bedrock. This usually occurs at high points along a divide where formerly a terrace had existed. Montmorillonitic soils tend to move gravel downward by expansion and cracking during wet and dry periods. Sandy soils are often intensely reworked by gophers with the same results.

The types of sediment included in a terrace are those common to any flood plain of a modern river system. Deposits are of two types. Those of lateral and vertical accretion, more commonly called point bar and overbank deposits.

Point bar deposits form on the inside or convex bank of a river bend. Deposition is caused by the existence of helicoidal flow associated with the channel bend (Wolman and Leopold, 1957, p. 91). Circulatory currents cause points of maximum and minimum turbulence to occur at the concave and convex banks respectively. Points of erosion and deposition correspond to turbulence distribution. The balance between material eroded from the cut bank and that deposited on the point bar tends to maintain a constant channel width as the system shifts.

The rate of lateral migration is variable. Large rivers tend to have a more rapid rate; however, the data show no consistent relation between river size and rate of migration. Measurements of one Brazos River bend found on the Clay 7½ minute quadrangle show no shift from 1910 to 1926, but a shift of 1500 feet occurred between 1926 and 1958 or an average of 46 ft/yr. Even at slower rates over a period of several thousand years the river could easily move from one valley side to the other, as the broad valley of the Brazos River indicates.

Deposits of vertical accretion include very fine sands, silts, and clays laid down by overbank flow. Overbank flow is that which tops the height of the concave bank. Fine sandy silt is often deposited near the channel boundary because of a rapid loss of velocity away from the channel. Vegetation on the banks aids this process by acting as a sediment trap. Natural levees are a common feature resulting from overbank deposition. Clay is deposited in poorly drained areas behind levees and in the distal portions of the flood plain. Oxbow lakes and sloughs created by neck and chute cutoffs or migration of point bars are abundant on the Brazos River flood plain and are the principle sites of accumulation of overbank clay and organic muds. The thickness of vertical accretion deposits in a stable flood plain (one associated with a non-aggrading or degrading river) is minor compared to that of lateral accretion. This is because of the low frequency of overbank flow and the small amount of deposition associated with an overbank event. An approximate rate of deposition for the area near Clay, Texas, Brazos County is 0.002 ft/yr (locality 1, fig. 4). This rate is based on discovery of an assemblage of Indian artifacts which have a minimum age of 1,000 A.D. (fig. 5) and which were unearthed by deep plowing (approx. 2 feet). The Ohio, Connecticut, and Kansas Rivers have been observed to deposit during major floods 0.008, 0.114, and 0.098 feet of sediment respectively (Wolman and Leopold, 1957, p. 970). However, overbank flow may scour instead of deposit, a factor often determined by the vegetal condition of the flood plain (Morisawa, 1968, p. 87).

MODERN FLOOD PLAIN AND CHANNEL CHARACTERISTICS

The modern flood plain is not a terrace by definition, but it will be considered with the terraces because it contains the record of post-Wisconsin alluvial history and when it is abandoned will become the youngest terrace level. The flood plain is widest between Waco, McLennan County, and the coast. It ranges from 1 mile wide near Hempstead, Waller County, to more than 8 miles wide north of Bryan, Brazos County. The postulated width of the ancestral Brazos in this region (2600 feet) fits easily into the limiting minimum width of the modern flood plain.

The thickness of flood plain alluvium at a particular cross section in the lower valley (below Waco, Texas) ranges from 9 to more than 75 feet, the average being about 45 feet. Upstream from Waco thickness is about 35 feet or less. South of Rosenberg, Ft. Bend County, the thickness is about 100 feet and increases toward the Gulf of Mexico (Cronin and Wilson, 1967, p. 21). The flood plain deposits of the upper Brazos along the Clear Fork near Lueders in Shackelford County have a maximum thickness of 56 feet (Preston, 1969, p. 42). Flood plain deposits along the Salt Fork in northeast Throckmorton County are at least 40 feet thick (Preston, 1970, p. 33).

Grain size of flood plain sediments ranges from clay through cobble (1/256 mm to 256 mm) with some boulders of local origin. Generally, the coarsest sediments (those of lateral accretion) occur in the lower part of the flood plain and grain size decreases upward. This upward-fining sequence rarely is gradational. Size changes are usually abrupt both vertically and laterally, a common characteristic of fluvial deposition. The presence of fine sedimentary units between coarser ones sometimes gives the appearance of reverse grading. Figure 6 shows an example of reverse grading found near the junction of the Brazos and Little Brazos Rivers (locality 2, fig. 4). This sequence from bottom to top consists of (1) well sorted massive fine sand containing clay inclusions; (2) pea gravel in red clay matrix grading to pure red clay; (3) fine-to-medium sand interlaminated with clay and containing clay chips; (4) low angle trough cross-stratified sand and pea gravel with clay balls and pea gravel lenses; and (5) poorly sorted gravel in silty sand matrix with clay seams. The gravelly unit is at the same elevation as the upper point bar across the river. The top gravel unit may be related to the proximity of gravel from the first terrace level.

A section through a modern Brazos River point bar in the lower basin reveals an upward-fining sequence from gravel to silty clay. The associated sedimentary structures are similar to those described by Bernard and Major (1963, p. 350). The lower point bar is characterized by two zones. The first is a broad flat composed of poorly bedded gravel and cobbles. This area is bounded by the scour pool at the concave bank and the gently sloping portion of the convex bank, and is only exposed at very low stages of river flow. The second portion of the lower point bar is the sloping zone, which rises about 6 feet. It is composed of sand and gravel with a higher sand-to-gravel ratio than the portion adjacent to the scour pool. The average gravel diameter is in the pea gravel class. Large trough cross-stratification and thin foreset cross-stratification are the dominant structures. A 3- to 5-foot scarp facing the channel marks the beginning of the upper point bar (fig. 7, locality 3). Vegetation also begins here. Silt-size sediment dominates the upper point bar but fine sand and clay are also abundant. The surface of the upper point bar is described as ridge and swale topography. The depressions or swales are sites of clay deposition during falling stage. Upper point bars are locally cut by numerous transverse gullies developed between flooding intervals. These gullies are seldom mentioned in the literature but they are important features to consider when doing paleocurrent studies on ancient point bars. Sedimentary structures of the upper point bar include small trough cross-stratification, parallel laminae, and climbing ripple laminae (figs. 8, 9). Roots and other organic debris are common constituents of these deposits.

Fig. 5. Brazos flood plain dart points (Locality 1) Archaic stage. Top row: left to right—Broken knife, Palmillas, Yarbrough?, Pedernales, Bulverde. Bottom row: left to right—Gary bases?, Ensor, Elam?, Crudely worked dart point.

The most complete exposure of the flood plain in the lower Brazos is at the Gifford Hill gravel pit near Benchly, Brazos County (locality 4). These pits near the east valley wall are not yet flooded. The section consists of 14 to 30 feet of black clay representing oxbow or swamp fill, overlying 20 to 30 feet of sandy gravel (fig. 10). The pit bottoms in the Stone City Formation of the Claiborne Group. The sandy gravel shows no upward-fining in gravel size but the unit is capped by 1 to 3 feet of lightly oxidized sand beneath the black clay. It has been suggested that the oxidized zone may represent a buried soil. The upper surface of the gravel body appears to have been veneered by a coarse gravel (cobble size) lag, as is common to many point bar surfaces where the fines have been winnowed. The dominant structures are tabular sets of foreset cross-stratification 1 to 2 feet thick that tend to fine upward within each set (figs. 10, 11, locality 4). The tabular sets give the impression of horizontal bedding when viewed from a distance. Sections at right angle to the dip of foresets show broad, shallow, trough cross-stratification. One exposure in the pit (fig. 11) showed that the basal 4 feet of gravel below the tubular sets is coarser and structureless. The top of the gravel unit is near the same elevation as the tops of the silty clay units on modern point bars in the adjacent river reach. There are other areas where coarse gravel crops out at the same elevation as upper point bars in the modern river. Approximately 2 miles south of the U.S. Highway 79 bridge in Robertson County in the east cut bank

Fig. 6. Point bar stratification, Brazos County (Locality 2). Lower Unit, moderately sorted medium sand overlain by six inches of red clay. Upper Unit, poorly sorted cross-bedded gravelly sand with clay balls overlain by coarse gravel in a clay matrix.

of the Brazos River (locality 3) a large lens of gravel composed of chert cobbles crops out between low water and 10 feet above low water. The lens is capped by approximately 1 foot of sand overlain by 15 feet of silty clay. These coarse point-bar deposits suggest a major change in flow regime since their deposition.

Drillers' logs of test holes in the Brazos flood plain from Waco to Richmond, Texas, show other areas where thick gravel deposits are only 10 to 15 feet below the surface (Cronin and Wilson, 1967, fig. 27).

Another source of flood plain data is provided by Bernard *et al.* (1970). Under sponsorship of Shell Development Company an extensive coring program was conducted between the coast and Richmond, Texas. Analyses of trenches, cores, and electric logs gave a three-dimensional picture of depositional facies within the Brazos-Oyster Creek meander belt and the deltaic plain. South of Sugarland, Texas, two SP logs indicate a stacked sequence of three point bars capped by channel fill deposits. These point bars were dated from bottom to top as deposits of recent low sea-level stage, deposits of rising sea-level stage, and deposits of standing sea-level stage. It was also shown that a modern meander near Richmond has cut-and-fill relationship to older point bar deposits. Numerous carbon-14 dates are shown for core samples. Many of these dates appear to be anomalous and the fact that they are not discussed by the authors probably supports this conclusion. The dates do generally show that the modern meander belts are of standing sea-level stage (3,000 or 5,000 years B.P. to Holocene) and that basal deposits are from late Wisconsin to post-glacial (35,000 to 5,000 years B.P.).

Fig. 7. Point bar zonation. Foreground, coarse channel gravel. Middle, gravelly sand unit containing sets of large trough cross-stratification and foreset cross-stratification. Upper point bar, silt and fine sand unit containing small scale trough cross-stratification and climbing ripple lamination. Discontinuous clay seams are common along with interbedded organic debris. Background, tall sycamore trees growing on older upper point bar deposits. Robertson County (Locality 3).

Another absolute date associated with the modern flood plain comes from Horn Rock Shelter, a Paleo-Indian site found along an undercut Edwards Limestone bank approximately 8 miles southeast of Lake Whitney Dam. Radiocarbon dates indicate habitation of the site 10,800 years ago. This level is located approximately 7 or 8 feet below a fill of collapsed limestone roof and floodplain silty clay immediately adjacent to the river channel.

South of Brazoria, Brazoria County, the Brazos River flows onto its broad deltaic plain. The present drainage area extends about 25 miles from San Luis Pass to the divide of the San Bernard River, but the combined Colorado-Brazos deltaic plain is over 50 miles wide. In this area fresh water swamps and lakes, coastal marsh, and meander belts are equally important environments of deposition. The most mature meander belt crossing this area is that of Oyster Creek. It can be traced up the east side of the valley to Sealy, Austin County, where its flow was apparently diverted into the modern meander belt. Indian sites are abundant along Oyster Creek point bars with dates circa 500 B.P. It is doubtful that these would have been preserved intact if the major flow was through Oyster Creek at that time; therefore, diversion was at least prior to this date. Bernard *et al.* (1970, p. 8) date abandonment at approximately 1,000 years ago. Figure 12 shows Oyster Creek where it entered the Gulf of Mexico. Marine erosion has destroyed all physiographic evidence of its associated delta.

The delta of the modern Brazos River is a high-destructive wave dominated type (Fisher, *et al.*, 1969) (fig. 13). It has developed entirely since 1929 when the flow of the Brazos River was diverted through a dredged channel to improve Freeport harbor which is located along the original river mouth. Figures 13, 14 and 15 are a series of photos showing delta development over the past decade. For earlier photos see Bernard *et al.* (1970, p. 13). From these photos it can be seen that the shore line has prograded almost 2 miles in 40 years. The process has been accomplished by redistribution of delta-front sands and channel-mouth bar deposits as a series of beach ridges. The areas between beach ridges become small lagoons which are gradually filled with washover and organic deposits. The littoral drift is dominantly southwesterly thus the delta cusp grows faster on that side of the channel (fig. 15). The cusp on the east side of the mouth has been reduced by storm waves. A model depositional sequence for the New Brazos Delta based on core data from Bernard *et al.* (1970, p. 14) is shown in figure 16. The facies in this model are rarely completely represented in a single core.

Suspended sediment load associated with the delta has averaged 26,725,833 tons per year or 760 tons per square mile of Brazos basin since 1925 (Cook, 1970, p. 26). This rate of suspended sediment flow could lower the basin 8 feet in 10,000 years. The assumed sediment density is 1525 tons per acre-foot of sediment (Cook, 1970, p. 2). Addition of solution and traction

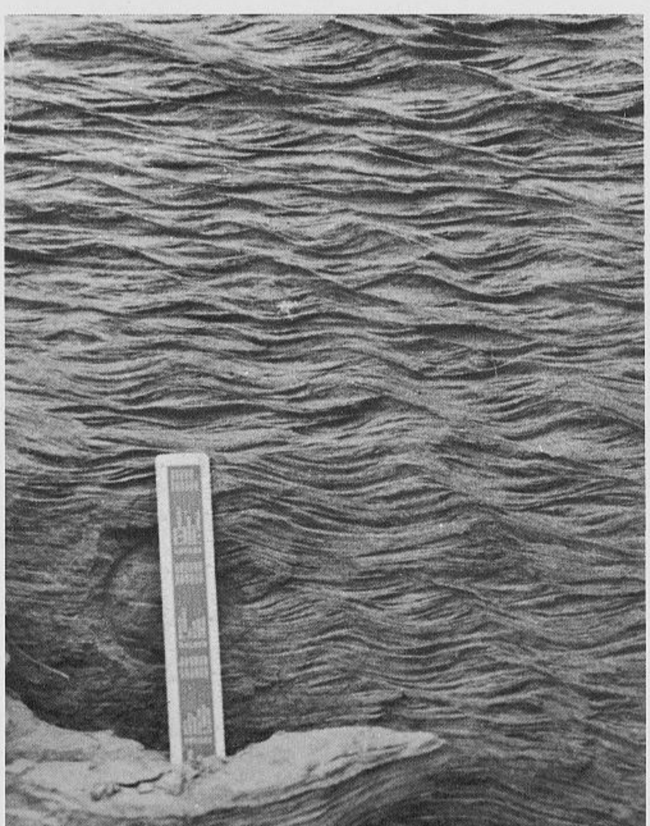

Fig. 8. Upper point bar sedimentary structure, Brazos County (Locality 2). Small scale trough cross-stratification and climbing ripples.

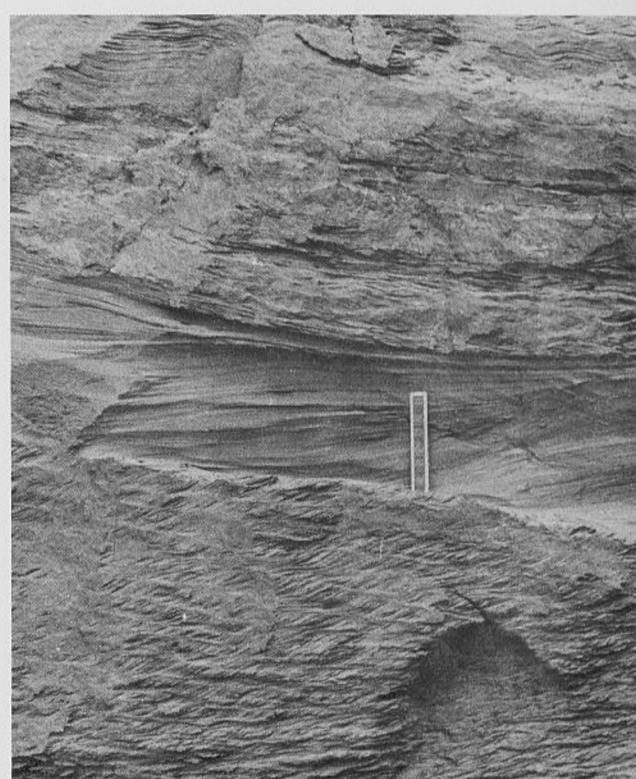

Fig. 9. Upper point bar deposits reflecting changing flow regime, Brazos County (Locality 2). Typical small scale cross-stratification and climbing ripple lamentation separated by a scour trough containing large scale cross-stratification unit of fine sand.

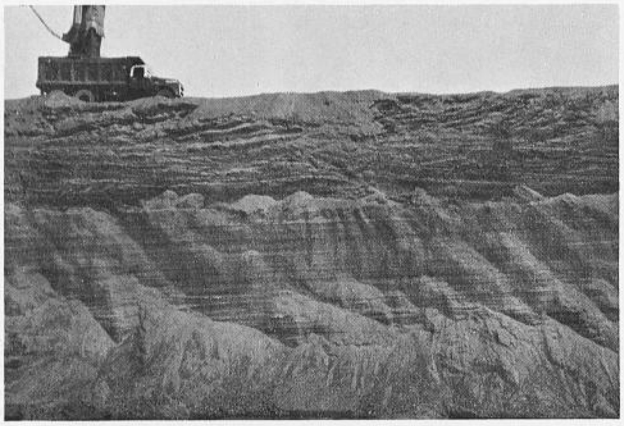

Fig. 10. Point bar sequence, Robertson County (Locality 4). Lower unit, large sets of trough cross-stratified sandy gravel and sandy gravel foresets. Contact zone, thin discontinuous slightly oxidized medium sand. Upper unit, organic rich black clay. The thick clay unit represents back-swamp and oxbow deposition.

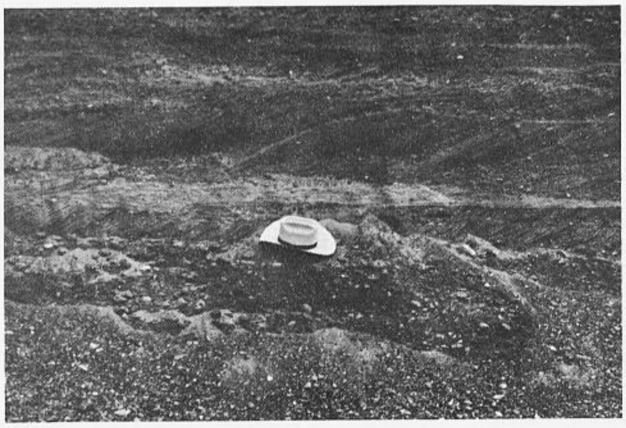

Fig. 11 Point bar basal deposits, Robertson County (Locality 4). Basal unit consists of coarser unstratified gravel representing the scour pool. The upper unit represents the lower point bar environment and is characterized by sets of trough and foreset cross-stratification.

Fig. 12. Oyster Creek meander belt, Brazoria County. This view looks north at the highly developed Oyster Creek meander belt. Oyster Creek is an old Brazos channel that parallels the modern meander belt for many miles upstream. Marine erosion has destroyed all physiographic expression of the Brazos-Oyster Creek delta.

Fig. 13. New Brazos River delta 1960, Brazoria County. This is the cuspate delta of the New Brazos River before hurricane Carla. Note the old shoreline, beach ridges, channel mouth bar, and mixing pattern of turbid fresh water with marine water. This photo looks northward toward the Oyster Creek meander belt and the Old Brazos River. The distance from the channel mouth bar upstream to the original shoreline is approximately two miles.

load would increase this value. The U.S. Soil Conservation Service has estimated that bed-load sediment entering Possum Kingdom Reservoir is 30 percent of the suspended load at South Bend sampling station (*ibid.*). This amounts to an average of 116 tons per square mile of basin above South Bend. However, the proportion of bed load is highly variable from time to time and from place to place in a given stream. It is probable, considering total stream load, that present lowering of the Brazos basin averages about 1 foot per thousand years.

The Little Brazos River is a striking feature of the flood plain between the north Robertson County line where it enters the flood plain and flows along the eastern valley wall to its junction with the Brazos west of Bryan, Brazos County. This 45-mile segment has been cited as an example of a Yazoo stream and it is tempting to consider it an old channel of the Brazos that is in some way linked to the Oyster Creek meander belt. However, geomorphic evidence indicates that the Little Brazos River is not an old channel of the Brazos (for location see fig. 4). The river lacks the meander belt pattern typical of the present Brazos River and other old channels such as Oyster Creek. Neither oxbows nor ridge-and-swale topography are associated with the Little Brazos River. The meanders of the Little Brazos River are poorly developed and have a smaller radius of curvature than the modern river. The Little Brazos River represents a flood plain drainage channel developed by short, steep-gradient creeks that cut the east valley wall at regular intervals. These creeks are unable to cross the flat between the east valley wall and the Brazos River which runs near the west wall. Therefore, they turn south near the eastern wall and link with the next creek down valley. This process apparently created the Little Brazos River. At its mouth the Brazos River has cut across valley to the east wall thus allowing a junction. During overbank flow the Little Brazos carries significant amounts of basin runoff. A diversion channel between the Brazos and Little Brazos Rivers northwest of Calvert, Texas takes advantage of this feature to aid flood drainage.

A longitudinal profile of the Brazos River from its mouth to the head of White River on the southern High Plains, Curry County, New Mexico, exhibits a concave upward curve, a result of the generally downstream decrease of slope (note: White River is called Running Water Draw in its upper reaches). Two important factors related to this decrease are a gradual decrease of median grain size and an increase in discharge downstream. The manner in which mean annual discharge increases in a downstream direction can be derived from Appendix C. Data on grain size variation is too limited to permit valid generalization. However, it is notable that the river near Sealy, Austin County has a gradient of 0.6 ft/mile and is still able to transport grain sizes in the cobble class (fig. 17). Some segments of the river have a convex profile which indicates downstream increase of slope. One such section begins at the foot of the Ogallala escarpment (fig. 18) and continues approximately 80 miles down the Salt Fork of the Brazos. Average gradients of the upper and lower reaches are 4.3 ft/mile and 6.3 ft/mile respectively. A combination of factors is responsible for this feature. A short distance upstream from the head of this section the White River enters the Salt Fork. White River crosses the High Plains at an average gradient of 8.3 ft/mile. It enters the relatively low gradient Salt Fork through Blanco Canyon at a gradient of 14 ft/mile. Normally flow is insignificant

Fig. 14. High flow sediment load, 1965, Brazoria County. This photograph shows the high discharge conditions and suspended sediment load of the Brazos River on May 29, 1965. A short distance upstream at the gaging station near Juliff, Texas, discharge was measured at 90,200 cfs and represents overbank flow conditions. Suspended load in this photo is approximately 343,354 tons/day or 6.5 times greater than the daily average for 1965. Note the influence of the littoral drift on sediment dispersal.

Fig. 15. New Brazos River delta, 1969, Brazoria County. This view looks northeast up the coastline. The first lineation to the right of the intracoastal canal is the original beach line of 1929 before diversion of the Brazos River. The cusp in the background is the old river mouth with its jettied entrance to Freeport harbor. Marine currents have truncated the beach ridges associated with the older delta. Dow Chemical Company is located on the large point bar in the background and beyond this area the beach is visible where it truncates the Oyster Creek meander belt. This photo was taken eight years after hurricane Carla struck the coast line. Compare it to pre-Carla appearance of the delta as shown in figure 13 and to figure 14 which was taken four years after Carla. Note that the channel mouth bar is not a large as before and the depressions between beach ridges have become more sediment filled.

(averages 4 cfs), but when a major storm occurs flow has reached peaks of 12,000 cfs. High flows deliver sand loads to the Salt Fork beyond the capacity of the stream to transport. The situation is compounded by the geology of the local Salt Fork basin: easily eroded Permian red beds which weather to bad-lands topography. Other causes of local steepening of slope are related to channel bed rock. For example, where the Brazos River crosses the Edwards Limestone below Lake Whitney it has a gradient of 2.5 ft/mile. Upstream and downstream from this area the gradient is approximately 1.4 ft/mile. Apparently the flat resistant surface of the Edwards Limestone causes the river to flow along a component of its regional dip slope.

The average gradient of the Brazos is 1.3 ft/mile from its mouth to the junction of the Salt and Clear Forks. The channel pattern of this river segment is dominantly meandering; however, above Waco the river bends are often controlled by regional joint patterns and below Waco, west of Hempstead, Waller County, lineation of flat-sided meanders and oxbows suggest fracture control.

Some meanders are deeply incised. Six miles northwest of Palo Pinto, Palo Pinto County, the river has cut more than 400 feet into Paleozoic sandstone and shale (locality 5). Figure 19 contrasts the asymmetrical profile of one of these incised meanders to that of a coastal plain meander (locality 6, fig. 4). The upper surface and the outlier of the surface on the north side of the river in the lower basin profile (fig. 19) are part of the upper delta plain in the Beaumont Formation.

The Salt and Clear Forks exhibit both meandering and braided segments. A typical braided section is found along the Salt Fork west of Seymour, Baylor County. Here the channel is broad and shallow with a local gradient of 4.4 ft/mile. Numerous sandy islands in the channel are sometimes stabilized with a growth of "Salt Cedar" (*Tamarix* sp.).

Flood-plain deposits can be distinguished by distinctive soil types. Miller, Norwood, and Yahola are the most common soil types found in both the upper and lower valley. Miller soil is predominantly dark reddish brown clay whereas Norwood soil tends to be more silty. Yahola soil is light reddish brown, very fine sandy loam, and is usually found on natural levees. Most soil profiles are poorly developed because of the youthfulness of the flood plain deposits. All three varieties are calcareous (U.S. Dept. Agriculture, 1958).

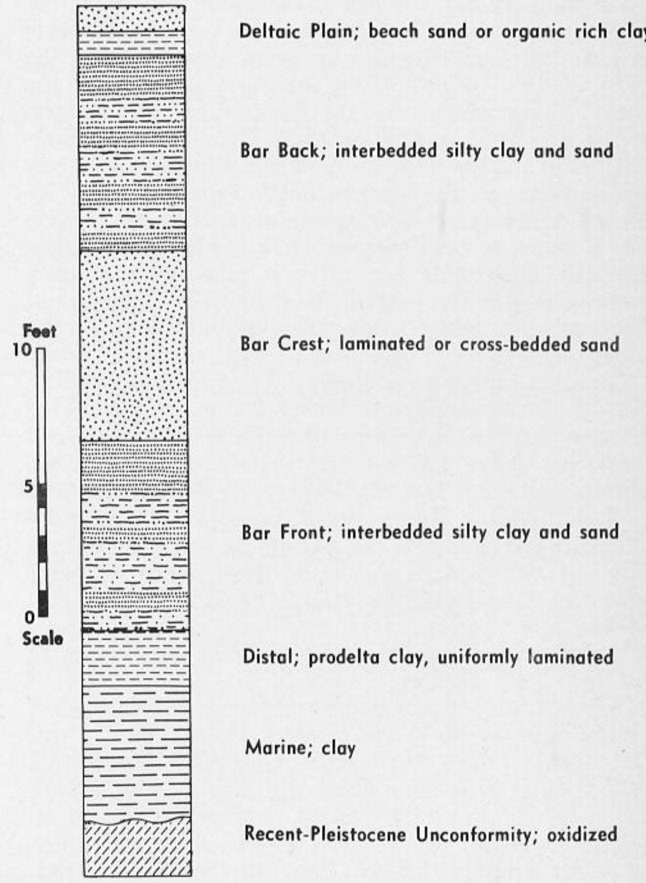

Fig. 16. Model core sequence for Brazos delta.

TERRACES OR ABANDONED FLOOD PLAINS

First Terrace. The first or lowest terrace of the Brazos River is still physiographically distinct; its scarps and flats are relatively undissected; swampy depressions and incompletely filled oxbows are still discernable on the terrace surface. Its proximity to the modern flood plain also aids recognition. Figure 20 shows a series of valley profiles along the Brazos River from the Salt Fork and Clear Fork to the coastal region from which terrace heights can be approximated. Average elevation of the first terrace above the adjacent flood plain is about 30 feet. Near Bryan, Brazos County, (fig. 20, profile F) this terrace is paired but more commonly it alternates from the east to west valley sides as it is traced upstream. Near Calvert, Robertson County and Asa, McLennan County, (fig. 20, profiles D and E) the terrace surface grades into the flood plain without a distinct scarp.

Test drilling on a gravel prospect east of Robinson, McLennan County, (locality 7) revealed alluvial deposits in excess of 35 feet but the average thickness was between 25 and 30 feet. A typical log of this area is shown in figure 21. The units encountered are laterally discontinuous and the sequence is that of point bar deposits. The data for a well drilled in the first terrace between Knox City, Knox County, and the Salt Fork of the Brazos River is also logged in figure 21 (Ogilbee and Osborne, 1962, p. 157).

Good exposures of the first terrace are difficult to find because of limited dissection. The gravel pits south of Waco along the Downsville road (about 3 miles north of Downsville on the east and west sides of the road) are among the best but the lower section is flooded. Conglomerates with carbonate cement are common in deposits of the first terrace.

The width of the flood plain associated with the first terrace level can be seen in figure 20, profiles F and G. In the Bryan, Brazos County, and Hempstead, Waller

County, area it is 6 and 9 miles respectively.

To compare terrace gradients, elevations from the base of the modern floodplain and the first terrace were projected orthogonally to a common axis following the trend of the Brazos valley between Waco and the coast. The gradients for the flood plain and the first terrace are both approximately 2 ft/mile. This similarity explains the uniformity of relief between the terrace surface and the modern flood plain at various localities. Only as the first terrace nears the coastal plain does it begin to converge with the modern flood plain (figs. 19, 20, profile L). This is probably the result of steepening of the coastal plain gradient as the result of subsidence before subsequent entrenchment of the modern river below this coastal plain surface.

The delta plain equivalent of the first terrace has been mapped as the Beaumont Formation in Texas, where soil type and surface slope distinguish the Beaumont Formation from the modern deltaic plain (Barton, 1930, p. 1303 and Metcalf, 1940, p. 698). The modern flood plain is 5 to 10 feet below the adjacent Beaumont surface. The slope of the Beaumont surface is from 2 to 3 ft/mile and is oblique to the modern drainage valley in an easterly direction (fig. 20, profile H). Physiographic features such as delta distributary channels, oxbows, and meander belts are still discernable on the Beaumont surface.

Interdistributary and flood basin areas are characterized by soils of the Lake Charles series (U.S. Dept. Agriculture, 1957). This series consists of dark gray, acid to neutral, flat-lying clays over alkaline clay subsoil that is lighter gray and mottled with yellowish brown. Edna soils mark the position of distributaries which are often slightly higher than surrounding terrain (*ibid.*). Edna soils are light gray fine sandy loams over red and yellow mottled subsoils containing iron and carbonate concretions below 2 or 3 feet. Bernard soils are transitional between Edna and Lake Charles soils and often mark levee deposits. They are grayish brown to dark gray friable clay loam.

The geometry of the distributary channels delineated by soils and the slope direction of the Beaumont surface indicate that the delta plain trended more easterly toward Galveston Bay than modern drainage. A conservative estimate of the area covered by the Brazos-Beaumont delta plain would be larger than the modern Brazos delta plain by a factor of 3.

Soils of the first terrace are distinct from those of the flood plain. Generally their profiles are better developed than soils of the younger flood plain. Burleson soils are commonly found on the first terrace (Table 7; U.S. Dept. Agriculture, 1958a, 1961, 1962b). These include dark gray clay that is slightly acid and grades into unweathered calcareous reddish-brown clay at 5 feet. Bell clay (fig. 21) and Bastrop fine sandy loam are also characteristic of the low level flats of the first terrace (Table 7; U.S. Dept. Agriculture, 1958b). The vegetation of the soils on the first terrace is mostly bunch grass, mesquites, and scattered trees. The Miles fine sandy loam is associated with the first terrace of the upper valley near Throckmorton (U.S. Dept. Agriculture, 1963b) and Seymour, Texas (U.S. Dept. Agriculture, 1963a). This reddish-brown soil has a zone of caliche 5 feet below the surface approximately 4 feet thick.

Fig. 17. Bedload, Austin County. Extremely low flow conditions exposed this view of the channel gravel adjacent to Stephen F. Austin State Park. Average gradient in this river reach is 0.6 ft/mile. Figure 29 shows fossil vertebrate material that was found in this gravel.

SECOND TERRACE. The base of the second terrace level is 45 to 60 feet above the Brazos River flood plain over much of its course. Along the Clear Fork of the Brazos River near Fort Griffin it is 90 feet above flood plain level (fig. 20, profile B) and just before it merges with the coastal plain south of Sealy, Austin County it is 35 feet above flood plain level.

Erosion has generally destroyed the scarp and flat of the second terrace surface thus physiographic expression is of limited value in determination of gradient. This fact is compensated by the abundance of good exposures in which alluvial sediments can be easily recognized.

The basal contact crops out on the west side of State Highway 6 about 0.5 mile south of its junction with Farm Road 159 near Millican, Brazos County (locality 8, fig. 4). The section includes about 9 feet of highly oxidized clay, sand, and gravel which exhibit typical point bar sedimentary structures. The fluvial deposits overlie light gray, laminated, micaceous, silty clay belonging to the Tertiary Jackson Group. This is a Navasota River terrace, but identical terrace deposits can be seen in a pit 5.5 miles southwest of this area (locality 9, fig. 4) just over the Brazos-Navasota divide (fig. 22). Soils on this divide are developed on a thin veneer of deposits of the third terrace. Gravel in this pit is composed of quartz, cryptocrystalline quartz, quartzite, and petrified wood. The wood is the coarsest material and commonly reaches lengths of a foot or more. Average size gravel is in the 0.25- to 0.5-inch

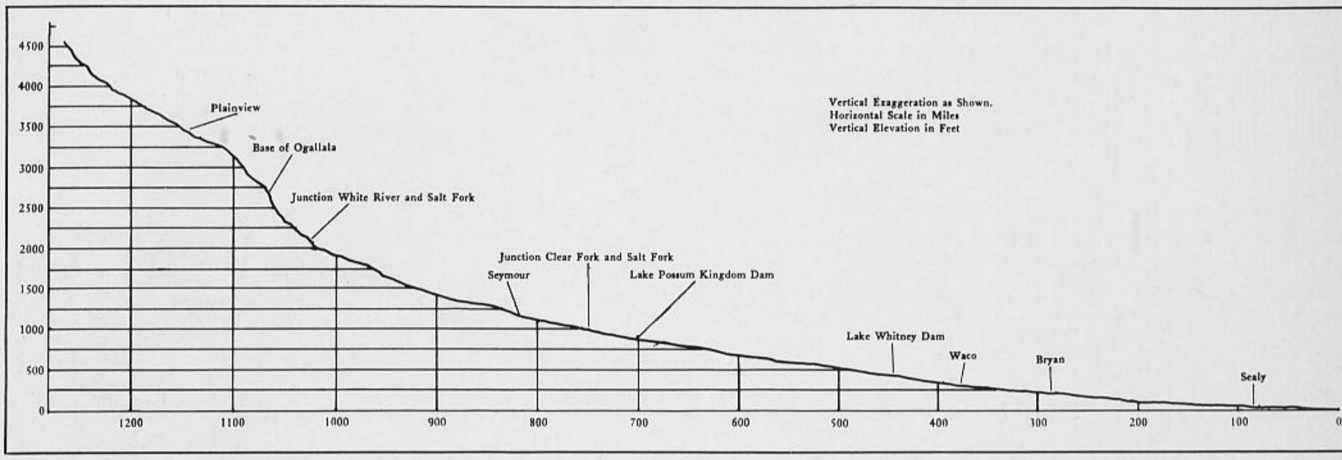

Fig. 18. Brazos River longitudinal profile.

diameter class but the coarser fraction includes 2-inch diameters. Poorly sorted red sand and clay are the dominant matrix materials. Discontinuous clay seams are laterally replaced by trough fill sand and gravel. Large-scale trough cross-stratification is the characteristic sedimentary structure. Outcrops of the second terrace are usually mottled by light green to yellow zones that surround fractures and roots.

Exposures of this terrace east of Calvert show discontinuous conglomerate lenses with iron-oxide cement. These iron-oxide cemented conglomerates are similar in appearance to Triassic conglomerates found in the Dockum Group along the divide between the Double Mountain and Salt Forks of the upper Brazos basin, though the pebbles of Triassic conglomerate are less well rounded than Brazos pebbles and the Triassic conglomerates appear to contain more white quartz.

Second terrace deposits are paired west of Hempstead (fig. 20, profile G) and they indicate that the flood plain may have been 10 miles wide in this area. South of Waco the paired second terrace is 7 to 8 miles wide. Elevations of the base of the second terrace when projected to a line following the valley axis between Waco and the coast suggest that the average gradient for this erosion surface was 2.6 ft/mile as compared to 2 ft/mile for the bases of the modern flood plain and the first terrace. One point of interest on the longitudinal profile of the second terrace is an anomalous bulge in the Millican-Courtney, Brazos and Grimes Counties area (fig. 23). In this area the base of the terrace is about 40 or 50 feet too high. This suggests that the active Millican salt dome has elevated the terrace. The first terrace and the modern flood plain do not reflect this activity thus giving some indication of the most recent activity of the dome. This suggests that the accurate plotting of coastal stream gradients and their associated terraces could be a useful tool for interpreting structural evolution in the lower gulf coast.

Outcrops of the second terrace level are abundant in the upper Brazos River basin. Generally the approaches to many of the major bridges crossing the Brazos River are cut through material of the second terrace on one or both of the banks. This is also true for bridges crossing the Double Mountain, Salt, and Clear Forks of the Brazos River.

One notable upper basin outcrop occurs west of Seymour, Baylor County, on the south side of the bridge crossing the Salt Fork between Benjamin and Knox City, Knox County (fig. 20, profile A). In the scarp of the second terrace a fauna of fossil molluscs was collected by Stricklin (1961, p. 31) and identified by Leonard as correlative to Sappa markers (late Kansan-early Yarmouthian) in the Kansas Pleistocene. This

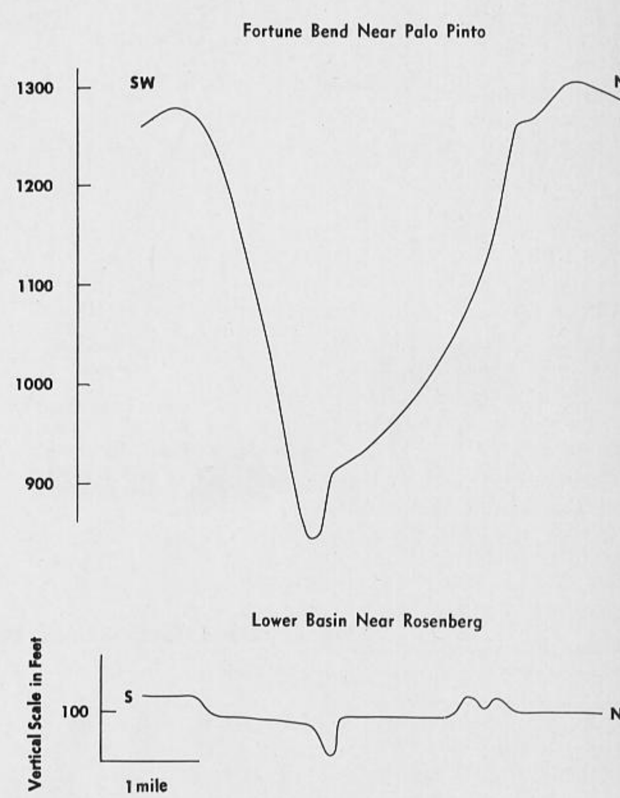

Fig. 19. Brazos Valley profiles.

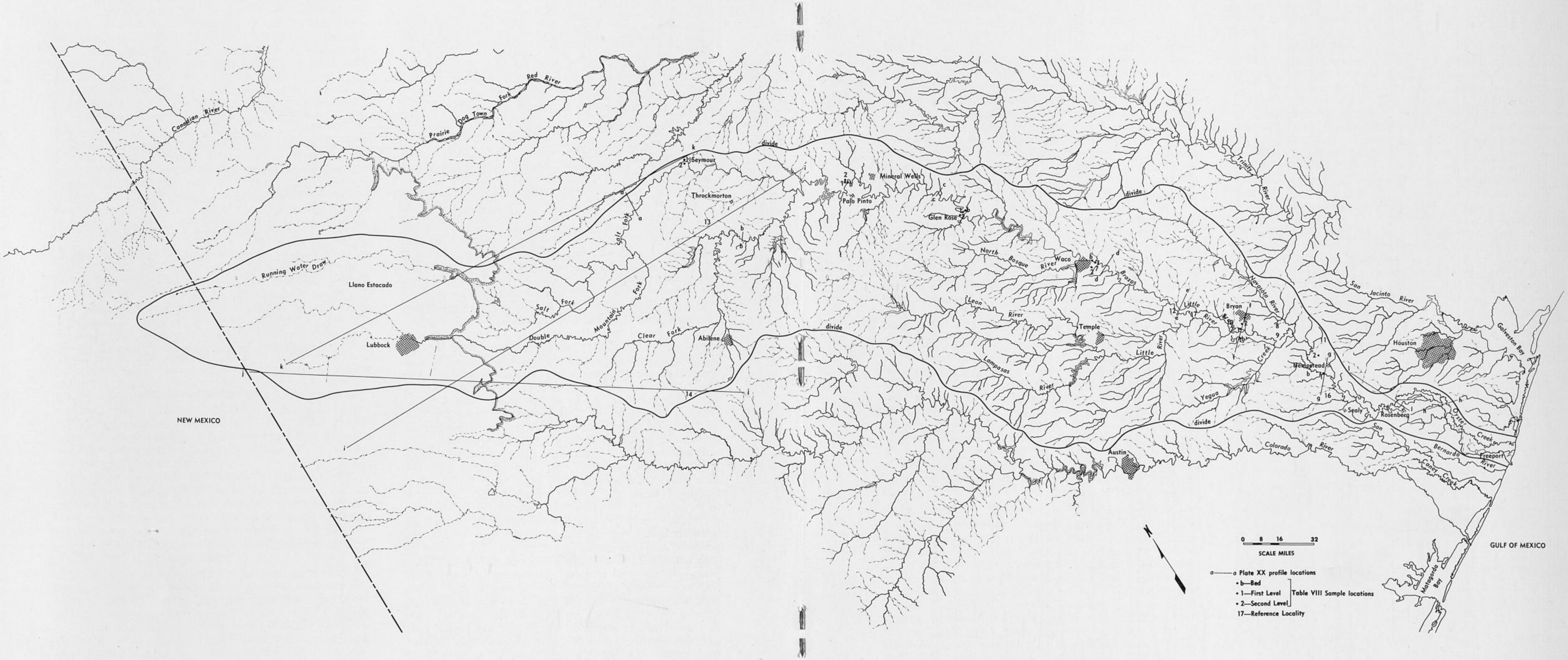

Fig. 4. Brazos River drainage basin.

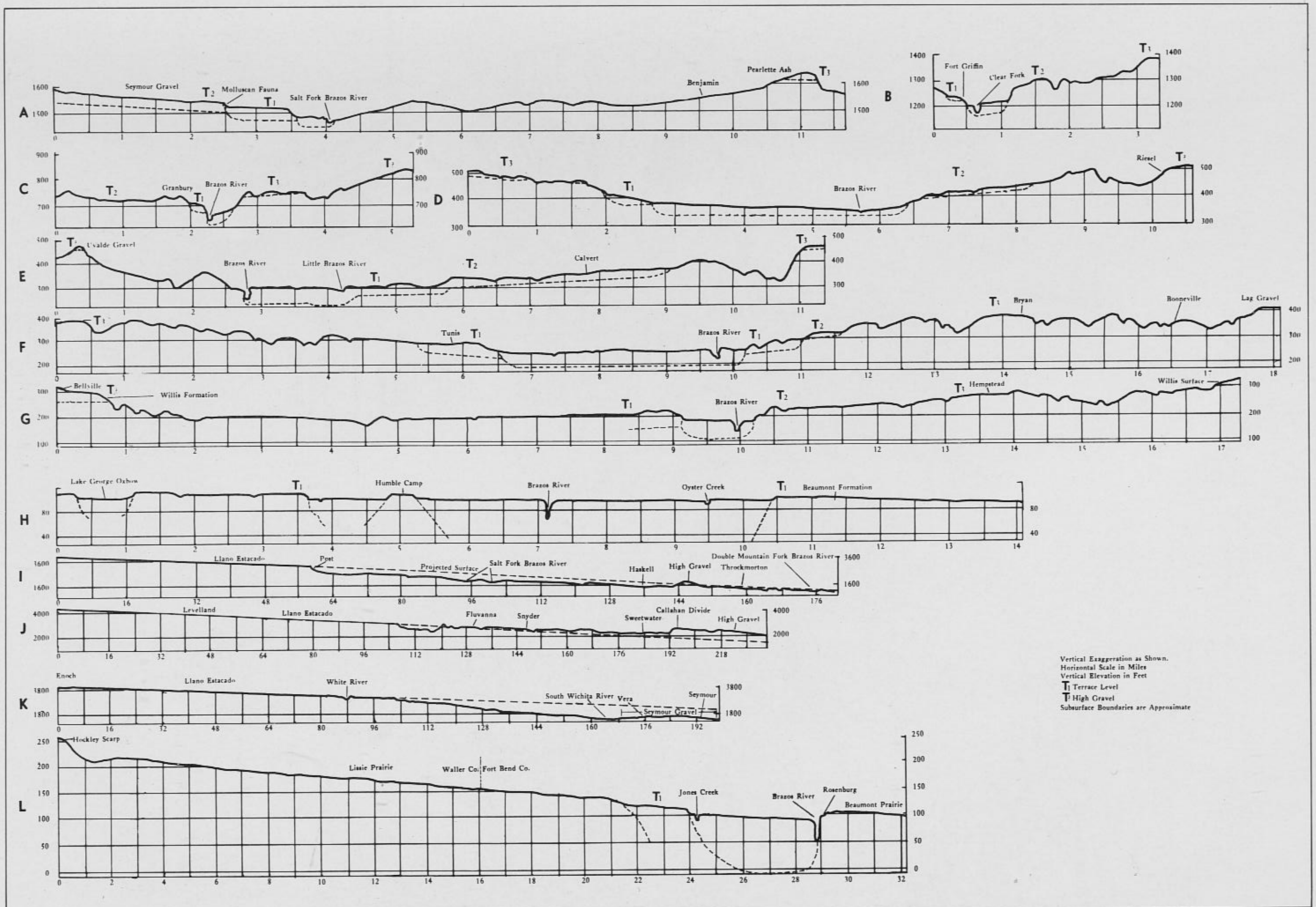

Fig. 20. Brazos Basin profiles

outcrop is in the "Seymour Gravel" approximately 60 feet above the present flood plain. North of this outcrop, on the divide between the Salt Fork and South Wichita Rivers, lenses of Pearlette Ash crop out in alluvial deposits which are mapped as "Seymour Gravel" (fig. 20, profile A). These deposits are 170 feet above the present flood plain. The Pearlette Ash is also dated as late Kansan to early Yarmouthian* (Frye and Leonard, 1952, p. 104). The disparity between the height of the Pearlette Ash and the adjacent molluscan fauna of the same age can be interpreted as follows: (1) The molluscan fauna is dated correctly and the Pearlette Ash is also dated correctly. This would mean that the Pearlette Ash fell into ponds occupying an elevated terrace level. One would expect to find Pearlette Ash deposits included in the alluvium of the lower second terrace which contains the molluscs. However, the ash appears to be confined to deposits on the higher divide. (2) The molluscan fauna is dated properly, but the ash deposits belong to an earlier volcanic episode. (3) The ash is indeed Pearlette, but the molluscs are younger than Sappa. A misinterpretation of the age of the molluscs could result if climatic conditions or some other unknown factor caused the fauna to be extant in Texas alluvial deposits while in Kansas it was viable only during Sappa age. Bob Slaughter of Southern Methodist University (oral communication, 1972) indicated that he has radiocarbon dates from deposits which contain the so-called Sappa markers in other areas of Texas which are late Wisconsin in age.

A late Kansan vertebrate fauna was collected by Hibbard and Dalquest (1960, p. 21) in deposits correlative with the elevation of those which contain the Pearlette Ash previously mentioned. This locality is only a few miles north of Benjamin, Knox County, across the South Wichita River valley and about 3.5 miles south of Gilliland, Knox County. This evidence suggests that the third hypothesis is most probable. Thus the high level (170 feet +) "Seymour Gravel" is of late Kansan age and the lower level (60 feet +) "Seymour Gravel" is considerably younger (Sangamon?).

At Stephen F. Austin State Park in Austin County, near the southern limit of second terrace exposures, the base of the terrace crops out above middle Pliocene deposits of the Goliad Formation in the west bank of the river. The upper portion of the Goliad Formation at this locality is sand cemented with white caliche and resembles a truncated paleosoil. The Goliad Formation grades downward into cross-bedded sand and pea gravel. The coarser gravel fraction consists of rounded caliche fragments. These deposits are reminiscent of the deposits and sedimentary structures associated with the broad shallow channel of the Salt Fork of the Brazos River near Seymour, Baylor County. The coarser gravel fraction at Seymour also consists of well-rounded caliche nodules which have been eroded from the subsoils of adjacent cut banks. The terrace sediments above the Goliad contact include red clay balls, trough cross-bedded pea gravel, and moderately sorted medium-to-fine sand. Immediately south of this area the second terrace level grades into the Lissie coastal prairie (fig. 20, profile L).

The Lissie Prairie has an average gradient of 4 ft/mile. It has been subdivided into two units by many investigators based on a subsurface "boundary" and surface gradient changes. Bernard, Leblanc, and Major (1962, p. 176) subdivided it into a coastward depositional surface called "unnamed second terrace" and a more interior depositional surface which dips beneath the "unnamed second terrace" called the Lissie surface. The gradient of the "unnamed second terrace" is 2.5 ft/mile and that of the Lissie surface is 3.5 ft/mile (*ibid.*). This change in gradient is not recognized in a profile drawn across these two surfaces from Rosenberg, Fort Bend County, to the Hockley Scarp, Waller County (fig. 20, profile L). The subsurface boundary between the "unnamed second terrace" and the Lissie is recognized in water wells drilled on the "unnamed second terrace" which penetrate two upward-fining sequences with a total thickness of approximately 100 feet (Bernard, *et al.*, 1962, p. 208). Bernard, *et al.* (1962) believe each of the sequences to be correlative with the "unnamed second terrace" and the Lissie; however, an alternate hypothesis is that both sequences belong to only one unit, either the "unnamed second terrace" or the Lissie. Wells drilled in the flood plain of the lower Brazos River basin support the latter, since they penetrate multiple upward-fining sequences that are all radiocarbon dated as Wisconsin (Bernard *et al.*, 1970).

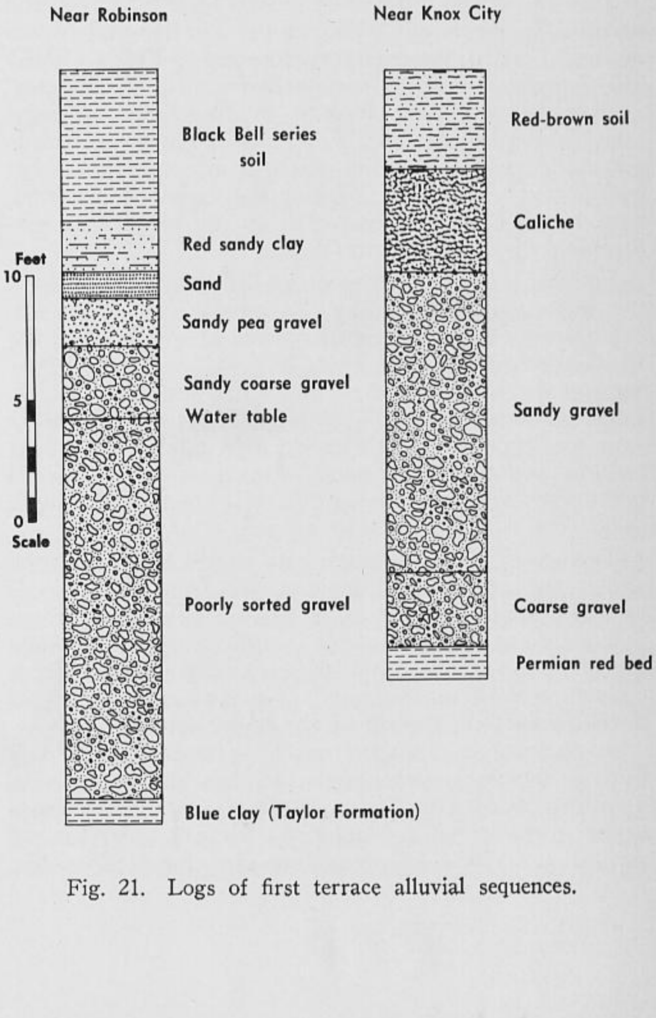

Fig. 21. Logs of first terrace alluvial sequences.

*The precise age of the "Pearlette" Ash is open to some controversy. See Izett and Wilcox, 1971, GSA Abstracts with programs, v. 3, no. 7, p. 610.

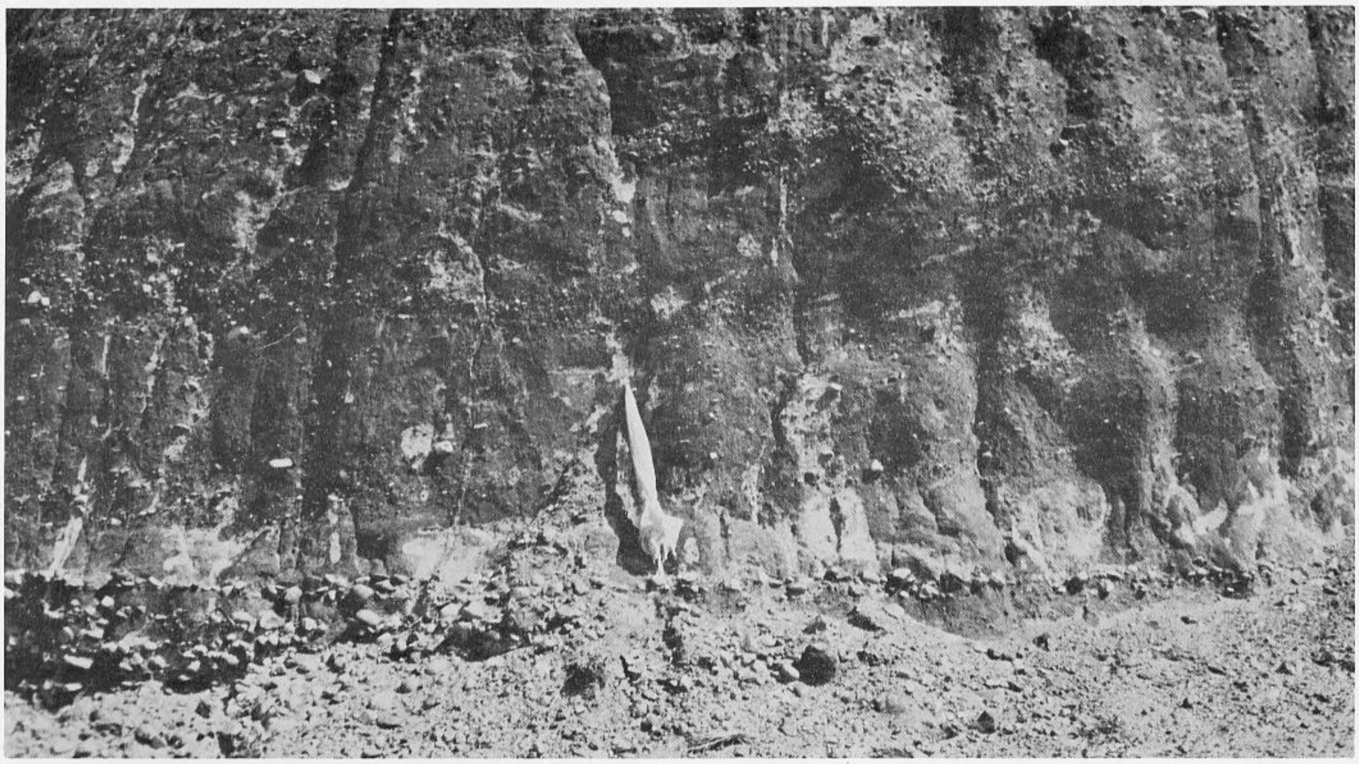

Fig. 22. Second terrace alluvium, Brazos County (Locality 9). Highly oxidized clayey sand and gravel. Note basal coarse gravel and the large scale cross-stratification above this unit. Compare this figure with figures 11, 24.

The Lissie and "unnamed second terrace" have been mapped by Aronow (1968) as the Bentley and Montgomery Formations, which correspond to Fisk's (1938) old Louisiana terrace nomenclature. C. V. Proctor, University of Texas Bureau of Economic Geology, (oral communication, 1972) has mapped numerous genetic depositional units, such as oxbow muds, that cut across Aronow's Bentley-Montgomery mapping boundary. This casts suspicion on the validity of subdividing the Lissie Formation.

Soils of the Lissie Prairie are different from soils of the Beaumont Prairie (fig. 20, profile L). However, soils alone do not provide a means of subdividing the Lissie Prairie into two units. The surface is characterized by the Katy, Hockley, and Edna soil series (U.S. Dept. Agriculture, 1957, 1961, 1970). Katy-Hockley soils are gray to grayish-brown acid sandy loams over red and yellow mottled sandy clays grading downward into yellow and red clay. Edna soils are similar though they have a dark gray clay subsoil.

Terrace alluvium belonging to upper levels (second and third) are easily distinguishable from lower levels by soil type. Because of a close similarity between soils of these two terraces it is difficult to discriminate one from the other with confidence. Planosols and red-yellow podzols, indicative of more advanced stages of development, are typical of the upper terraces.

A planosol is characterized by a well developed B horizon which is more compacted than associated soils. Typically it develops upon nearly flat upland surfaces under grass or forest vegetation. Mature soils that are similar and occur in areas that are climatically alike

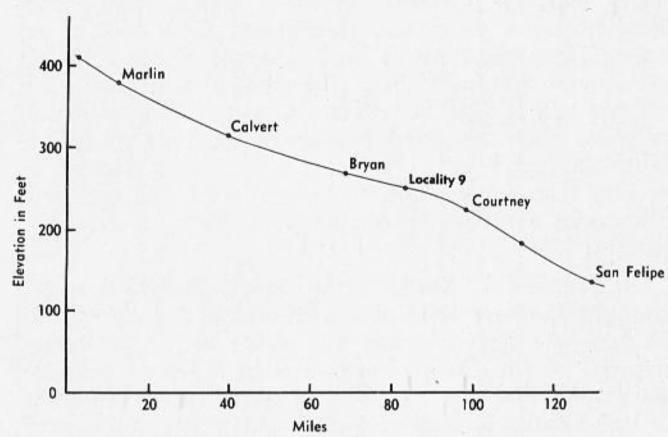

Fig. 23. Second terrace longitudinal profile.

can be grouped into "zonal soils." Planosols are called "intrazonal soils" because they reflect effects of local conditions more than "zonal" climatic conditions. The planosols of the terraces reflect the dominance of a clayey parent material (U.S. Dept. Agriculture, 1958a, p. 38). A typical planosol of widespread occurrence is the Axtell series:

Axtell Series Soil Profile

A horizon—Varies from 4 to 12 inches thick; light brown fine sandy loam (65 percent sand, 25 percent silt, 10 percent clay); friable

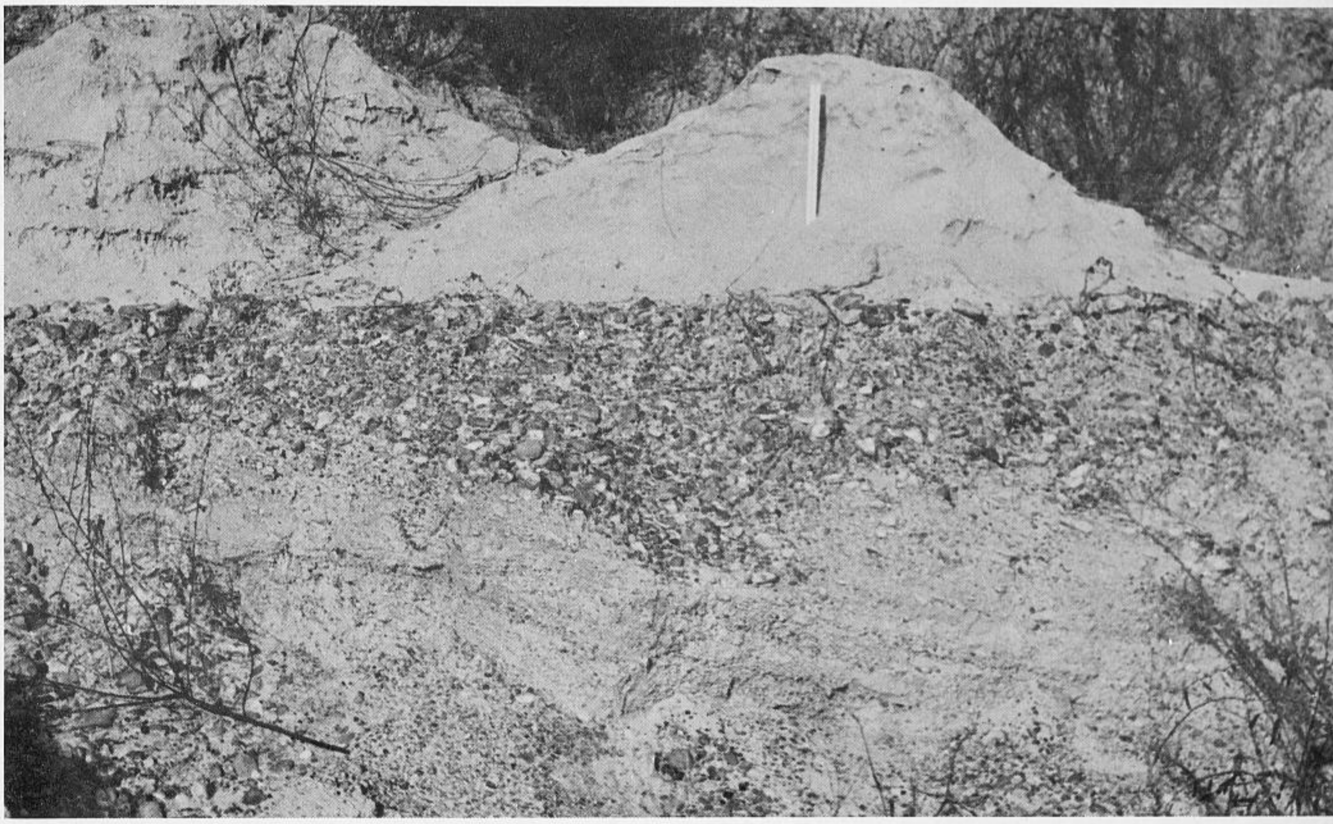

Fig. 24. Third terrace alluvium, Brazos County (Locality 10). Typical fining-upward point bar sequence. Note the coarse gravel scour fill between the light-colored sand unit and the cross-stratified lower unit.

when wet, hard when dry; pH 6; abrupt undulatory boundary with B horizon.

B horizon—Interval from 12 inches below surface to 45 inches; reddish-brown to yellowish-red clay; sticky when wet, very hard when dry; mottled pale brown and reddish-brown spots; pH increases downward from 5 to 8; often contains calcium carbonate concretions near base.

C horizon—Interval from 45 inches below surface to 66 inches; pale brown to yellowish-red calcareous clay.

Red-yellow podzolic soils are characterized by well developed A, B, and C horizons. The A horizon is highly leached of iron and carbonates thus giving it a light color from which the group derives its name (Thornbury, 1966, p. 77). Usually this soil group is developed under conifers and hardwoods in a cool humid climate. The B horizon is a bright red with mottled yellow resulting from the concentration of ferruginous oxides leached from the A horizon. The soil is strongly acid. Thornbury (1966) indicates that red-yellow podzols occur in areas with a mean annual temperature of approximately 75° F and annual rainfall from 64 to 128 inches. The Lakeland series includes the maximal red-yellow podzolic soils in the Brazos basin (U.S. Dept. Agriculture, 1958a, p. 43). The Lakeland series appears to be more commonly but not exclusively associated with the third terrace.

Lakeland Series Soil Profile

A horizon—Varies from 30 to 72 inches thick; very pale brown loamy sand (80 percent sand, 15 percent silt, 5 percent clay); loose when dry, friable when wet; pH 6.

B horizon—Interval from 53 inches below surface to 97 inches; mottled yellow and red clay; very hard when dry, sticky when wet; upper zone has concretions of manganous and ferruginous oxides; pH 5.

C horizon—Interval from 97 inches below surface to 120 inches; white sandy clay; mottled red; friable when wet, hard when dry; pH 5.5.

The association between the third terrace and the red-yellow podzol suggests that time, not process, is the key soil-forming factor. Red-yellow podzols should represent senile soils in an area where neither glaciation nor transgressing seas have interfered with weathering in the last million years (Carter and Sokoloff, 1954, p. 13). It has been suggested that 500,000 years is the time required to produce podzols similar to the Lakeland series. This suggests Yarmouthian or greater age for the upper terrace levels. The extended weathering period appears intuitively more acceptable as an explanation for formation of red-yellow podzols than the alternative option of increased annual rainfalls from 2 to 4 times the present rate. Greater antiquity of the

Lakeland series would also explain the unusually thick soil profile (8 feet). A list of some terrace-related soils is given in Table 7.

TABLE 7 — TERRACE-RELATED SOIL SERIES

Low Terrace	High Terraces	Contain Lag Gravel*
Bastrop	Abilene	Crockett gravelly loam
Bell	Axtell	Eddy gravelly loam
Bernard	Hockley	Tabor gravelly loam
Brewer	Irving	
Derby	Katy	
Eufaula	Kenney	
Hortman	Lakeland	
Irving	Milam	
Ivanhoe	Norge	
Lake Charles	Payne	
Milam	Sawyer	
Payne	Stidham	
Springer	Travis	
Stidham	Wichita	
Vanoss		

Third Terrace. The base of the third terrace is between 90 and 120 feet above the flood plain of the Brazos River between Granbury, Hood County, and Hempstead, Waller County. Along the Salt Fork near Benjamin, Knox County, the basal terrace deposits are 160 feet above flood plain level; near Fort Griffin on the Clear Fork they are approximately 180 feet above the flood plain (fig. 20, profiles A and B).

Lag deposits occur five miles west of Tunis, Burleson County, as well as four miles east of Bryan, Brazos County, at an approximate elevation of 360 feet. This corresponds to the height of the third terrace in this area. The separation of these two localities suggests an immense flood plain 18 miles wide. However, this value may not represent a measurement at 90° to the old flood plain trend. No other area provided an opportunity for a second estimate. Elevations projected to a line following the valley axis between Waco and the coast suggest an average gradient of 2 ft/mile for the third terrace which is similar to that of the modern valley.

The terrace is found on the rolling uplands and on divides between lower order drainage channels. A well preserved surface is located north of Waco around Connally Air Force Base. In many areas the surface is heavily vegetated with stands of hardwood and brush (Yaupon). Exposures can be seen in numerous small gravel pits developed on the third terrace. Deposits vary in thickness from a thin veneer of gravel to clay, sand, and gravel 20 feet thick. Many of these pits expose deposits that are similar to the oxidized sediments described from the second terrace and illustrated in fig. 22. These sediments can be seen in the numerous pits between Hearne, Robertson County and Hempstead, Waller County.

The best exposure is located in a gravel pit 6 miles northwest of Bryan on the Mumford Road, a light-duty county road (fig. 4, locality 10). The soils developed on this locality are Tabor gravelly loam and Edge fine sandy loam (U.S. Dept. Agriculture, 1958a, sheet 18). The pit exposes 20 feet of alluvium representing channel, lower point bar, and upper point bar deposits. Fig. 24 shows a sharp boundary dividing the trough cross-bedded sand and gravel of the channel facies from the clean sand of the lower point bar. Deposits in this pit are not as highly oxidized as is common to third terrace sediments in nearby localities. The presence of ferruginous conglomerates is probably related to local water quality. (Water associated with many of the iron oxide-cemented conglomerates is high in sulfates). A massive ferruginous conglomerate from the third terrace can be seen in a U.S. 190 roadcut approximately 10 miles northwest of Bryan, Brazos County.

A problem associated with the third terrace in the lower river basin is its confusion with the deposits of the Willis Formation. Wilson (1962, p. 349) considered third terrace deposits 1 mile east of Navasota, Grimes County on State 105 to be Willis (late Pliocene or early Pleistocene). However 8 miles farther east on Texas Highway 105, the Fleming and Willis contact can be seen south of the highway near Yarboro and Stoneham, Grimes County, at an elevation of 380 feet, 80 feet above the elevation of the outcrop at Navasota. Soils and sediments of these two localities are quite similar but topographic position strongly indicates that they are not correlative.

About 6.5 miles south of Yarboro at Pleasant Hill church in Grimes County (locality 11) the base of the Willis crops out 80 feet lower than the outcrop near Stoneham. The steeper gradient of the base of the Willis Formation causes it to converge with the base of the third terrace near Hempstead, Waller County. The basal surfaces are so nearly adjacent and the deposits are so similar near Hempstead that it is impossible to separate the two units with confidence (fig. 20, profile G). The Willis Formation dips beneath the surface south of Sealy, Austin County. South of Hockley, Waller County, just before it goes beneath the surface it forms an escarpment known as the Hockley Scarp (fig. 20, profile L). The Willis strikes more north of east than the terrace units and dips approximately 11 ft/mile; however, if elevations are projected to a line following the trend of the Brazos River between Navasota and Hempstead the average gradient is 5 ft/mile, twice as steep as any terrace gradient projected to the same line.

Another problem related to the third terrace is that in the upper river basin west of Seymour, Baylor County, it is mapped jointly with the second terrace and sometimes with the first terrace as the "Seymour Gravel." The age problem associated with the Seymour Formation has already been discussed. The Seymour Formation crops out on the uplands trending northward from the great northern bend of the Clear Fork in Jones County (fig. 4) to the Red River in Hardeman County. Within the Brazos basin the Seymour Formation is composed of interbedded clay, sand, and gravel. The formation is commonly topped with blow sands that have been heavily cultivated. The largest expanse covers about 50 percent of Knox and Haskell counties (Ogilbee and Osborne, 1962, p. 19). A stratigraphic section drawn from Haskell northward through Knox City shows that the surface slopes uniformly toward the Salt Fork at about 8 ft/mile (Ogilbee and

*Parent material not alluvium; usually associated with highest terrace.

Osborne, fig. 19, profile C). On this section there is no evidence of a scarp dividing the second and third terrace levels, however in the subsurface about 7 miles south of Knox City there is an escarpment which generally divides the base of the second terrace (elevation 1500 feet) from the base of the third terrace (elevation 1550 feet) (*ibid.*). Six miles south of this subsurface feature is another escarpment, but more subdued, which suggests a fourth terrace level above 1600 feet elevation at its base.

An east-west stratigraphic section between Rule and Haskell adjacent to and at right angle to a north-flowing segment of the Double Mountain Fork allows some interesting speculations (Ogilbee and Osborne, fig. 19, profile B). The base of the Seymour Formation is almost entirely graded in an easterly direction away from the Double Mountain Fork and parallel with the slope of the Salt Fork valley in a downstream direction. One would expect much of this surface to be graded westward toward the Double Mountain Fork if this river is contemporaneous with the age of the Seymour Formation at this locality. This evidence suggests that a tributary of the Salt Fork working southward by headward erosion parallel to the Haskell-Stonewall county line captured the Double Mountain Fork which had previously drained eastward across southern Haskell County and entered the Clear Fork in Throckmorton County (fig. 4). This speculation agrees with the views of Lewand (1969, p. 21), who suggested that the Double Mountain Fork and the Clear Fork were linked through Paint Creek valley in Throckmorton County. The date of capture would have to be after deposition of the Seymour Formation in the Rule area by the ancestral Salt Fork. The elevation of these deposits in relation to occurrences of Pearlette Ash suggests that they are earliest Pleistocene or pre-Kansan. The projection of the High Plains surface (fig. 20, profile K) shows that the Seymour Formation is younger than Ogallala (Neogene).

Miscellaneous High Gravel. Throughout much of the Brazos basin, gravel deposits are found on high divide areas. Most of these deposits occur on points ranging from 200 to 500 feet above the local river level. Their high position indicates that they are older than the third terrace, but their range of elevation indicates that they are not all correlative with a graded fourth terrace level.

Some of the most abundant deposits occur in central Texas and are referred to as Uvalde Gravels, characterized by an abundance of cobble-sized chert and flint (fig. 25). A good exposure of this unit can be seen west of the Brazos River on a high point near Cross Roads, Milam County (locality 12, fig. 4 and profile E, fig. 20). Uvalde Gravel was apparently deposited as bed load in valleys extending from the eastern margin of the Ogallala plain before the development of modern drainage (Byrd, 1971, p. 29). The age of the Uvalde Gravel in central Texas is probably late Pliocene (*ibid.*).

A sparse lag of high gravel consisting of pea-size quartz and egg-size quartzite pebbles and cobbles is scattered over a broad area 12 miles west of Throckmorton (locality 13, fig. 4) 450 feet above the Clear Fork. This elevation intersects the projected High Plains surface (fig. 20, profile I) suggesting that these sediments correlate with the Ogallala 116 miles to the west. Higher lag deposits occur at an elevation of 2510 feet on the Callahan Divide south of Abilene (locality 14). A projected High Plains surface passes 600 feet below the top of the Callahan Divide (fig. 20, profile J) suggesting that the high lag is pre-Ogallala. A well developed Ogallala surface can be seen in numerous areas situated in valleys below the surrounding surfaces held up by lower Cretaceous deposits (Frye and Leonard, 1964, p. 23).

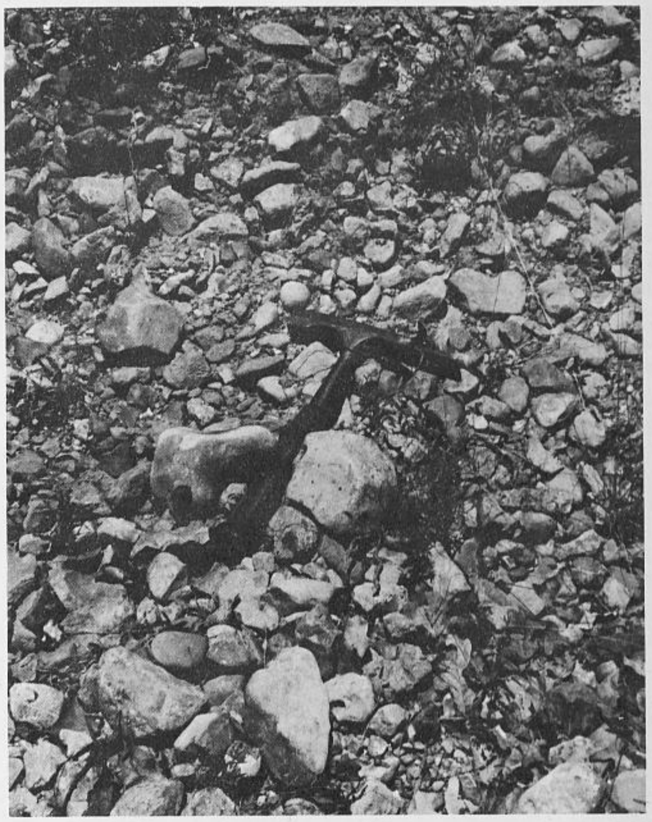

Fig. 25. Uvalde gravel, Milam County (Locality 12). Two or three feet of gravel composed predominantly of cryptocrystalline quartz.

South of Possum Kingdom dam, Palo Pinto County, pea gravel deposits of unknown age crop out 240 feet above the adjacent flood plain (locality 15, fig. 4). The river section through Lake Possum Kingdom is one of the most tortuous segments in the Brazos River basin, with a sinuosity factor of 2.6 (Table 2). Fortune Bend (Palo Pinto County), a well developed meander just below the dam (profile in fig. 19), has a ratio of radius of curvature to channel width of 5.6 (mean for the lower basin is 2.8). A third unusual feature in this area is the incision of meanders over 400 feet below a relatively flat upland surface.

First impressions suggest that incision is a result of rejuvenation. The high gravel lag thus might represent the mature flood plain upon which the first cycle meanders developed. Stricklin (1961, p. 22), because of asymmetrical valley profiles, believed that the meanders are primary and ingrown. However it appears that meander profiles may be of little value in distinguishing between first and second cycle meander development (Morisawa, 1968, p. 145).

The large radius of curvature suggests that the river may have been wider in its early history but incision has constricted its width while maintaining the original radius of curvature.

The reason for rejuvenation, if indeed it occurred, has never been adequately explained. A common cause cited for rejuvenation is epeiric uplift of the headwaters region, called "dynamic rejuvenation," but because of the localized nature of the area under consideration this explanation appears inadequate. "Static rejuvenation" may result from changes in stream load, increase in runoff because of increased rainfall, or increase in stream volume through acquisition of new drainage by stream diversion (Thornbury, 1966, p. 143). Of these three suggested mechanisms the last, increase in stream volume through acquisition of new drainage, may have been the most significant change affecting the Possum Kingdom area. Lewand (1969, p. 21) suggested that the Clear Fork originally entered the Leon River north of Eastland, Eastland County. The capture of the Clear Fork which at this time was also linked with the Double Mountain Fork may have triggered the change in river regime that caused incision of the Possum Kingdom segment. As the wave of degradation worked upstream following this event, the Salt Fork eventually may have captured the Double Mountain Fork. All of these major diversions would have occurred before development of the third terrace level, probably prior to Kansan time.

Areally extensive sand bodies 15 miles long and 10 miles wide mapped by Eifler (1967) as "blow sand" occupy the divides between the Double Mountain Fork and Salt Fork. The relationship of these upland sand units to the High Plains surface suggests a sandy plain beginning near the edge of the High Plains about 300 to 400 feet below its surface and sloping eastward approximately 11 ft/mile. This was later dissected by modern drainage development. It is not now known if all such sand deposits overlie fluvial sand and gravel, but the proximity to high lag gravel in the immediate area suggests that several sand areas must overlie basal gravel units, and may be equivalent in age to the highest deposits mapped as Seymour Formation further to the east. Eifler believes the "blow sand" overlies fluvial deposits (oral communication, 1972). These suggested relationships are speculative but detailed field investigation might eventually establish the age and nature of the first major alluvial surface developed from reworked Ogallala sediments.

The Willis Formation (discussed with the third terrace level) is situated on the divides in the lower Brazos River basin (Grimes, Waller, and Austin counties). The age of this formation is not certain, but Doering considered the Willis to be correlative with the Citronelle Formation of Alabama, "the basal member of the sequence of Pleistocene fluviatile formations (1956, p. 1853)." Bernard *et al.* (1962, p. 219) considered the Willis Formation to be Pleistocene in age because it overlies the Pliocene Goliad Formation unconformably and is overlain by the Pleistocene Lissie Formation. Wilson (1962, p. 349) suggests that the Willis is either latest Pliocene or early Pleistocene in age, and an eastern facies of the Uvalde Gravel.

Outcrops of Uvalde Gravel are separated from the northernmost Willis outcrops by the Oakville Escarpment. Projection of the Willis dip-slope northward suggests that Uvalde Gravel occupies a lower post-Willis depositional surface. Soils, gravel size, and composition are distinctly different in Uvalde and Willis deposits. These relationships suggest that the Uvalde is younger than Pliocene as implied by Byrd (1971, p. 20) and the age of the Willis may be older than Pleistocene. Both of these units probably occupy a transitional time period including late Pliocene and earliest Pleistocene. Good exposures of Willis deposits can be seen near Bellville, Austin County (locality 16, fig. 4).

COMPOSITION OF TERRACE GRAVEL

INTRODUCTION

If the Brazos River has always occupied the same basin or eroded the same geologic units, then the composition of the gravel in terraces of the river should be similar from level to level. If, however, the provenance has changed with time either by gains or losses of basin or exposure of different rock, then the composition of the terrace gravel should reflect these changes. For these reasons a study of gravel composition was undertaken.

Four basic components comprise the gravel of terraces along the Brazos River: (1) limestone, (2) chert, (3) quartz, and (4) quartzite. Qualitative inspection of the various terraces gives the impression that gravels of the upper terraces are more siliceous, because the coarser fraction, more obvious in upper terraces, is composed of chert and quartzite.

A statistical experiment was designed to determine whether differences in composition were real or imaginary. Six sampling areas were chosen at intervals along the Brazos River (fig. 4). Sampling stations were near bridges crossing the Brazos at Seymour, Palo Pinto, Glen Rose, Waco, Bryan, and Hempstead. At each locality samples were taken in the river channel, on the first terrace, and on the second terrace. These samples consisted of approximately 30 pounds of gravel collected within a 10-foot square.

Each sample was screened and only the fraction between 0.53 and 0.37 inches was retained for analysis. This small grain size was chosen because of its abundance and the hope that a more easily transported fraction might represent the total gravel population better than coarser material. Two samples of one hundred individuals each were randomly drawn, with replacement, from each sample bag. The number of components in each of four classes were counted: limestone plus others, quartz, chert, and quartzite. The "others" included mostly sandstone, caliche, and conglomerate.

These data were subjected to Chi-Square (X^2) analysis by constructing appropriate row-column contingency tables. (Chi-Square analysis is commonly used when the data consist of discrete variables such as the counts of compositional classes.) Table 8 shows the contingency table used for calculation of a Total X^2 value. It can be seen that the data in any one cell are classified by four criteria: mineralogy, locality, terrace level, and sample number.

TABLE 8 — TOTAL X^2 CONTINGENCY TABLE

Locality	Lev.	Sam.	Ls+oth. ob	Ls+oth. x^2	Quartz ob	Quartz x^2	Chert ob	Chert x^2	Quartzite ob	Quartzite x^2	Row Total
Seymour	Bed	1	61	20.8	5	7.8	24	2.0	10	3.2	100
		2	57	15.0	4	9.3	25	1.5	14	0.7	100
	1	1	13	13.2	22	2.0	41	2.6	24	2.4	100
		2	18	7.7	21	1.4	38	1.2	23	1.7	100
	2	1	16	9.7	16	0	19	5.2	49	56.7	100
		2	18	7.7	18	0.3	21	3.7	43	37.2	100
Palo Pinto	Bed	1	52	9.1	22	2.0	17	7.0	9	4.1	100
		2	67	31.2	13	0.7	11	13.7	9	4.1	100
	1	1	6	23.3	28	8.4	38	1.2	28	6.3	100
		2	5	25.0	24	3.6	49	9.2	22	1.2	100
	2	1	5	25.0	38	28.9	18	6.1	39	26.4	100
		2	14	12.0	32	15.1	17	7.0	37	21.7	100
Glen Rose	Bed	1	61	20.8	13	0.7	13	11.2	13	1.2	100
		2	73	43.7	8	4.2	10	15.0	9	4.1	100
	1	1	58	16.4	11	1.7	25	1.5	6	7.6	100
		2	64	25.7	6	6.5	26	1.1	4	10.4	100
	2	1	38	0.4	17	0.1	32	0	13	1.2	100
		2	41	1.3	13	0.7	38	1.2	8	5.2	100
Waco	Bed	1	90	90.5	2	12.5	6	21.0	2	13.7	100
		2	82	66.3	4	9.3	7	19.4	7	6.3	100
	1	1	79	58.3	7	5.3	9	16.4	5	8.9	100
		2	81	63.6	2	12.5	10	15.0	7	6.3	100
	2	1	5	25.0	30	11.5	46	6.2	19	0.1	100
		2	7	21.7	25	4.6	51	11.4	17	0	100
Bryan	Bed	1	46	4.0	14	0.3	32	0	8	5.2	100
		2	47	4.7	15	0.1	31	0	7	6.3	100
	1	1	5	25.0	25	4.6	56	18.2	14	0.7	100
		2	9	17.0	17	0.1	54	15.3	20	0.4	100
	2	1	6	23.3	19	0.4	48	8.1	27	5.2	100
		2	4	26.8	24	3.6	47	7.1	25	3.2	100
Hempstead	Bed	1	54	11.3	13	0.7	23	2.5	10	3.2	100
		2	43	2.2	8	4.2	30	0.1	19	0.1	100
	1	1	1	32.3	20	0,8	59	23.0	20	0.4	100
		2	4	26.8	23	2.8	52	12.7	21	0.7	100
	2	1	4	26.8	13	0.7	62	28.4	21	0.7	100
		2	1	32.3	16	0	63	30.3	20	0.4	100
Column Total			1235		588		1148		629		3600

Total X^2 = 1617.7 Degrees of Freedom = 105

STATISTICAL INTERPRETATION

Chi-Square tests are used in studies of association or comparison. In this experiment it is suggested that the mineralogy of a gravel sample is associated with the terrace level and geographic locality from which the sample was taken. To find an answer to this hypothesis one must determine if the observed differences in mineralogy are the result of sampling from different populations or whether the variation is a function of random fluctuation. A Chi-Square value allows determination of the probability that a difference in sample composition as great or greater could have occurred from merely random factors.

The calculation of X^2 follows a simple formula:

$$X^2 = \text{Summation } \frac{(o-e)^2}{e}$$

where: o = observed frequency;
e = expected frequency.

The observed frequency is generated by sample analysis, but the expected frequency is a value derived from the following equation:

$$e = \frac{r \text{ X } c}{n}$$

where: r = row total;
c = column total;
n = grand total.

For example, the column total for all quartz counts is 588, the row total is 100, and the grand total is 3600 (Table 8). Therefore, the expected frequency that should appear in each quartz cell if the classes are independent is:

$$e = \frac{588 \text{ X } 100}{3600} = 16.3$$

Using the basic formulas the Total X^2 value was found to be 1617.7. To find the probability of this value occurring randomly one must know the degrees of freedom (D.F.) involved in the Total X^2 value. This is determined by multiplying (r-1) (c-1). In this equation r and c stand for the number of rows and columns in the X^2 contingency table. Thus the degrees of freedom for the Total X^2 table is (3 X 35) or 105 D.F. By using a standard X^2 statistical table with the appropriate values of X^2 and D.F., it is found that the chance of X^2 = 1617.7 occurring randomly is remote (much less than 1 in 200 chances).

Since the total variation is not random it must be related to one or all of the classifying factors (sample, locality, terrace level) or to some unknown effect. Symbolically this means:

$$T = S + Lev + Loc$$

where: T = Total X^2; S = Sample X^2;
Lev = Level X^2; Loc = Location X^2.

To study the relative effects of each of these sources of variance, two more contingency tables were constructed and their X^2 value determined. Adding the observed frequencies of both samples at a particular level and locality caused removal of sample variance and allowed calculation of a Level-Location X^2 value. Similarly the effect of sample and level variation is removed by adding all observed frequencies at a particular locality for each class thus leaving only the Locality X^2 value. All other pertinent X^2 values can be derived from the previous three values. A summary of the results is given in Table 9.

The asterisk by the X^2 value denotes that they are significant. This means that it is highly improbable that this value would occur as a result of random fluctuations. The Sample X^2 value is not significant. This means within each level the sample variation can be explained by random fluctuation and thus sample differences within levels are unimportant in explaining total variation.

The F-distribution is used to determine the relative importance of Level and Location effect. It is defined as the distribution of variance ratios.

$$F = \frac{\dfrac{s_1^2}{S_1^2}}{\dfrac{s_2^2}{S_2^2}}$$

where: s_1^2 = sample variance

S_1^2 = population variance

The relation of X^2 to the F-distribution can immediately be seen if one divides its mathematical equivalent by its degrees of freedom.

$$X^2 = \frac{D.F.\,(s^2)}{S^2}$$

$$\frac{X^2}{D.F.} = \frac{s^2}{S^2}$$

The ratio of X^2 to D.F. is the Mean Square (Table 9). When using the F-distribution we assume that population variances are equal ($S_1^2 = S_2^2$), thus s_1^2/s_2^2 should approximate unity. If the value of the ratio of sample variances deviates farther from unity than can be explained by random fluctuation then the hypothesis that $S_1^2 = S_2^2$ is rejected and the alternate hypothesis that they are unequal is accepted. This implies that the estimators of population variance (s_1^2 and s_2^2) probably come from different populations.

An F-test for level effect is made by dividing the Mean Square of sample variance into that of sample plus level variance. The F value of 17.6 (Table 9) is highly significant at a 5 percent test level. This means there is a probability less than 1 chance in 20 that the gravel mineral suites of various terraces come from the same population. Therefore, it is established statistically that the terrace suites are different.

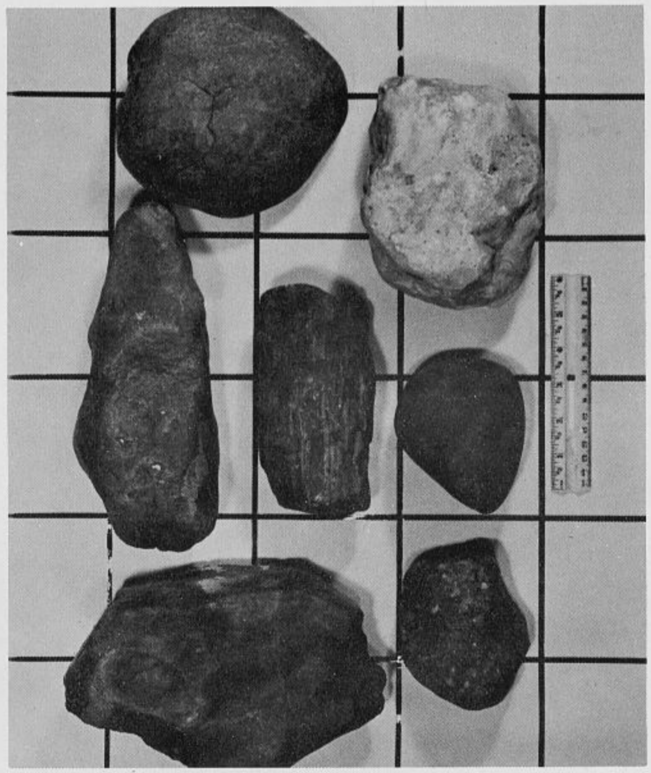

Fig. 26. Cobble Composition. Robertson County (Locality 17). These samples are from old abandoned gravel pits near Hearne, Texas. Top row: left, orthoquartzite; right, limestone. Center row: left to right, chert, petrified wood, metaquartzite. Bottom row: left, chert; right, conglomerate (probably from Cretaceous Trinity Group).

The F-test for location effect is calculated by dividing the Mean Square of sample plus level variance into the total variance Mean Square. The 1.3 value (Table 9) obtained is not significant at the 5 percent test level. The actual probability of this value occurring

TABLE 9 — ANALYSIS OF VARIANCE

Source of Variance	D.F.	X^2	Mean Square	F at 5%
Total (Sam + Lev + Loc)	105	1617.7*	15.4	1.3
Level + Location	51	1581.1*		
Location	15	554.4*		
Derived Variance				
Sample (Total — (Lev + Loc))	54	36.6	.67	
Sample + Level (Total — Location)	90	1063.3*	11.8	17.6*

*—denotes significant variance

in the F-distribution is about 1 chance in 10, a borderline probability. Generally it can be said that the terrane of a locality does not greatly influence the gravel composition. However, examination of the data shows this not to be true in all cases (Table 8), for the terrane does show an influence on the gravel composition, particularly in the limestone outcrop areas of Glen Rose and Waco. The lack of other influence is to be expected because the Brazos River basin contains no metamorphic or plutonic bedrock.

The experiment has shown that the mineralogic suite of each terrace level investigated is unique for the specific size fraction studied, suggesting that correlation of terraces might be aided by study of gravel composition and that geology of the provenance area changed as the river evolved. The general trends of composition are shown in figure 27. Each level is represented by 200 counts at six localities, thus the composition of each level is generalized by 1200 counts. Sedimentary material, mostly limestone, decreases while all other classes increase from the channel to the first and then second terrace level. The difference in abundance of chert between the first and second terrace is minor, possibly reflecting the manner in which chert was defined for the count. (All material of cryptocrystalline quartz whether wood, chert, flint, jasper, etc. was included in this category.)

The experiment does not include material from terraces above the second level because of the difficulty of finding localities where all levels are adequately exposed in a given area. Further, the uncertainty of level correlation increases in the upper terraces, which con-

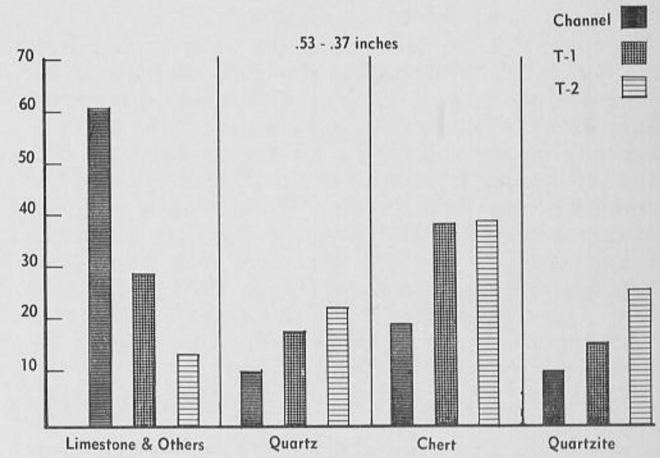

Fig. 27. Gravel composition.

tain gravel similar in composition to that of the second terrace. However, the gravel found along the highest points adjacent to the Brazos between Waco and Calvert is exceptional.

These deposits have long been known collectively as the Uvalde Gravel (Byrd, 1971). Cobble-size material is more common than in lower levels and chert seems to be more abundant. The Uvalde Gravel usually occurs as a lag, but at some localities it is 3 or 4 feet thick (fig. 25).

RIVER COMPETENCE

INTRODUCTION

Bedload gravels are measures of river competence, and competence can be related to a number of descriptive parameters relating to the river. Thus, terraces may contain independent information on river history in the form of the bedload fraction they contain.

Byrd (1971) traced the Uvalde Gravel from central Texas along the Brazos to the eastern edge of the Llano Estacado. He considered the Ogallala gravel to be the source of quartzitic material in the Uvalde Gravel. Quartzite and chert gravel in the Uvalde are commonly 4 to 8 inches at their maximum diameters (fig. 25). Gravel of this size is uncommon in the bedload of the present Brazos River. The conclusion appears to be warranted that the hydraulic parameters of the modern Brazos system and the Uvalde depositional system were different and that the earlier river was of greater competence. Thus, a study of relative stream competence of the modern and ancient Brazos River should desribe aspects of river hydraulics useful in interpreting river history. Competence is the measure of the maximum particle size of a given specific gravity which a river may move.

The solution of problems of stream competence involves a study of fluvial hydraulics, complex because of the numerous variables involved, their interdependence, and their relative randomness in time. An investigation of limited scope will normally include only a few variables and will rely on established hydrologic relationships in a theoretical framework. However, even such theoretical treatment offers potential for generation of information about past history of the Brazos River.

The ability of flowing water to transport debris depends on two forces: the gravitational impelling force, which is the downslope component of water weight and the resistance offered to this force by the channel. The effects of lithology and topography on the ability of flowing water to scour and transport are exerted principally through their relation to the resisting force. Because a river does not accelerate but moves with a constant velocity for a given set of conditions, the forces of gravity downslope and the frictional resistance of the channel are in equilibrium. This equilibrium velocity is actually an average of vertical and lateral differences.

The relation of the resisting bed or boundary shear stress to stream flow is best approached through the equations of Chezy and Manning (Leopold, *et al.*, 1964, p. 156). In a channel segment of unit width w, length L, and depth d, the force acting on the body of water is the downslope component of its weight.

$$1)\ F_1 = pgdLw\ \sin B$$

The angle B is the deviation of the water surface from the horizontal. For small angles sinB = tanB or the slope s of the stream. The symbols p and g represent fluid density and gravitational constant. The effect of viscosity on the ability of a stream to transport sediment is through its control of particle fall velocity, bed roughness, and fluid density. Water with a sediment concentration of 28,000 ppm or less has a specific weight (pg) of 62.4 lbs. per cubic foot when mass density is expressed in slugs (Cook, 1970, p. 2). The Brazos concentration is within this range (28,000 ppm is approximately the maximum sediment load that has been measured for the Brazos River).

The force (F_2) resisting the downslope weight component is the stress per unit area (t) times the boundary area over which it acts.

$$2)\ F_2 = t(2d + w)L$$

The term in parenthesis is the wetted perimeter of the channel. Because there is no acceleration, F_1 and F_2 are equated and solved for the stress t.

$$3)\ t = pgs\ \left(\frac{dw}{2d + w}\right)$$

The term in parenthesis is called the hydraulic radius and for streams that are much wider than deep it can be approximated by average depth (d). The product (pg) is the specific weight (y) of the fluid (62.4 lbs./cu. ft.).

$$4)\ t = ysd$$

Equation (4) shows that the bed shear stress is proportional to the depth slope product.

Shear stress (t) can also be expressed as a velocity (V). The relation, called the Manning equation, is widely used by engineers.

$$V = 1.49\ \frac{d^{.66}\ s^{.5}}{n}$$

This equation is empirical. Its successful use depends on the skill in estimating the value of the Manning number (n). This value is a measure of channel roughness and includes consideration of channel sinuosity, cross section, size of bed material, obstructions, and vegetation. Values range from 0.01 for smooth metal surfaces to 0.06 for very rough natural channels (Leopold, *et al.*, 1964, p. 158). The Mississippi River near Memphis, Tennessee has a value ranging from 0.025 to 0.045 (Chow, 1959, p. 105). The Brazos River near Bryan has a (n) value of about 0.03 based on calculations using stream parameters at flood stage.

FORCES ACTING ON A BED PARTICLE

Stress on a particle in a stream bed is influenced by: (1) particle weight, (2) bouyancy, and (3) drag or shear stress of water. The first two can be combined to give the immersed weight of the grain.

$$5)\ F_i = aD^3p'g$$

where: Fi = immersed weight; D = particle diameter; p' = immersed density of particle; g = gravitational constant; a = shape coefficient such that aD^3 equals the volume of particle; $a = \frac{\pi}{6}$ for a sphere.

The third force (shear or drag) can act on the particle in a number of ways depending on the Reynolds number (R) of the grain (Reynolds number can refer to a grain or a river depending on the definition of V in equation (6)).

$$6)\ R = \frac{Vd}{v}$$

where: V = fall velocity of grain; d = depth of stream; v = kinematic viscosity; (R) is the ratio of viscous forces to inertial forces and describes the nature of fluid flow around a grain. This flow is laminar if viscous forces dominate, a rare case in natural streams. It is turbulent if the reverse is true. When (R) is greater than 3.5 the shear forces acting on a grain pass through its center of gravity (Leliavsky, 1966, p. 55) and tangential components of drag are negligible. This case will be assumed for the present analytical model.

If the stress per unit area on the bed is (t) then the stress on a particle will be some fraction of (t) determined by the number and size of grains within a unit area.

$$7)\ F_d = t\frac{D^2}{c}$$

Where: F_d = drag force on grain; t = stress per unit area on bed; D = grain diameter; c = packing coefficient.

The packing coefficient (c) is defined for unigranular material as the number of grains per unit area times the square of the grain diameter. This study assumes 0.5 foot is the maximum diameter of any gravel found in the Uvalde Gravel. This is the diameter of the first approximation to shape which is a sphere. Therefore, one square foot of bed could contain from 1 to 4 grains. The packing coefficients for these extremes are 0.25 and 1 respectively.

The grain will not move as long as the resultant of the drag force and immersed weight falls within an angle defined by the angle of repose of the grain (fig. 28). The equilibrium condition is that in which the ratio of the two forces equals the tangent of the angle of repose.

$$8)\ \tan b = \frac{F_d}{F_i}$$

The critical value of stress (t_c) at incipient grain motion can be calculated by substituting the equivalent values of F_d and F_i from equations 5) and 7) into 8) and rearranging.

$$9)\ t_c = c\frac{\pi}{6}p'\,gd\,\tan b$$

INCIPIENT MOTION STUDY AND THE BRAZOS RIVER

A minimum stress for incipient motion is obtained by using 0.25 as the packing coefficient instead of one. The angle of repose for a grain 0.5 foot in diameter is about 40° regardless of angularity (Chow, 1959, p.

173). The only remaining value needed to solve for critical stress is the specific weight of the particle. Assuming the cobble is quartzose gives a specific weight of 165 lbs/cu ft. Using the previous values and solving equation 9) for (t_c) gives 5.6 lbs/sq ft. The mean value of stress needed for first motion is about half the calculated or 2.8 lbs/sq ft, because the turbulence produced in the wake of the grain causes the force to fluctuate and the maximum value reaches twice the mean (Kalinske, 1947, p. 616).

The critical mean drag force as calculated from the equation of Kalinske (1947, p. 617) is 2 lbs/sq ft. A U.S. Bureau of Reclamation graph relating grain size to critical stress gives a value of 2.5 lbs/sq ft (Chow, 1959, p. 173). Schoklitsch showed that gravel 2 inches in diameter is transported by about 1 lb/sq ft of stress (Leliavsky, 1966, p. 42). All of these values are close enough to the derived value to suggest that it is reasonable.

Equation 4) gives the shear stress per unit area of bed in terms of mean river depth and slope. The average slope of the Brazos between Waco and Calvert is about 1.7 ft/mile (3.2×10^{-4}). With this slope,* equation 4) suggests it would require a river approximately 140 feet deep to produce enough stress to move a cobble 0.5 foot in diameter. A test calculation using a grain size that is common in the Brazos River channel between Waco and Calvert would indicate the validity of the previous statement. Gravel 0.25 foot (3 inches) in diameter is commonly the coarsest bedload. Substituting the appropriate values into equation 9) predicts the stress needed for movement is 0.62 lbs/sq ft, or applying Kalinske's generalization about 0.31 lbs/sq ft. It will be noted that the relation between critical stress and grain size is not linear. This is because stress must overcome the immersed weight of a grain which is a function of the radius cubed. To achieve the required stress to transport a 0.25-foot-diameter cobble requires a river depth of 15 feet at the present slope. This is actually a minimal condition where there is only one grain per square foot of bed in this size class. With more than one grain per square foot it would take a deeper river. The mean depth of the Brazos River at bankfull stage near Bryan is 30 feet. The agreement between theoretical and actual depth in the test case seems to justify the theoretical 140-foot value and allow interpretation.

However, conditions in the modern Brazos River are such that it could not transport 6-inch cobbles. Furthermore, it is unlikely that a river 140 feet deep ever existed. The fact remains that the large cobbles are present in the highest terrace levels and even more significant, cobbles near this size (about 5 inches) have been observed in the coarse fraction of gravel pits in older alluvium of the modern flood plain (fig. 26). The

*The slope value used in equation 4) is the slope of the energy grade line which is approximated by the water-surface slope. For uniform flow channel bottom slope is a good approximation of water-surface slope. However, for rapidly shifting stages during flood flow the energy grade line can be steeper as the flood wave passes. This is particularly true for flood flow through tortuous channels with many channel obstructions. Under these conditions channel slope is not a good estimate of the energy gradient. The transport of large cobbles may be related to steeper energy gradients experienced during flood hydrology.

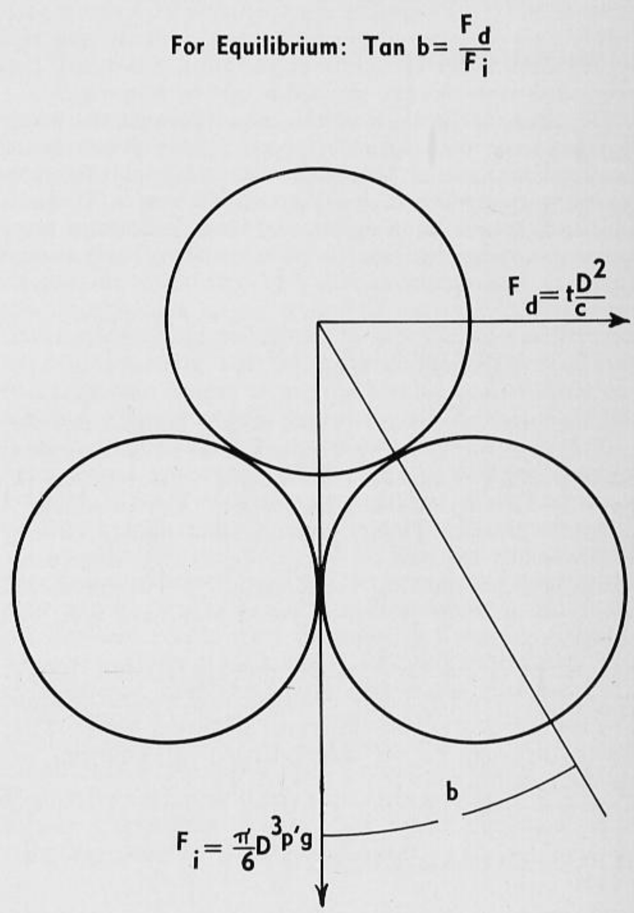

Fig. 28. Forces acting on a bedload particle.

cobbles in the flood plain (locality 17, fig. 4) suggest a river both deeper and steeper than the present one.

The stress required to transport a 5-inch cobble is 1.8 lb/sq ft. If the slope were increased by a factor of 1.5 or from a 1.7 ft/mile to 2.5 ft/mile (5×10^{-4}) the river depth required to produce the critical stress is 58 feet. This is within the realm of possibility since the present flood plain is 60 feet thick at points where the large cobbles are observed. However, the present river slope and bankfull depth of 45 feet (mean depth would be less) is not enough to produce the critical stress. These figures tend to support the concept of a larger ancestral Brazos River indicated earlier from analysis of meander geometry. The increase in gradient may be partially explained by decreased sinuosity. To suggest that a similar river deposited the Uvalde Gravel (that is a deep river flowing on a high gradient) is a possibility but is highly speculative. C. V. Proctor (oral communication, 1969) suggested that these high gravels may represent a plain of alluviation related to the Balcones Fault zone to the west of the Brazos River. The constant increase in elevation of the Uvalde Gravel toward Austin, Texas, from uplands on the west side of the river gives this hypothesis merit though the quartzites in the Uvalde gravels indicate that they were not locally derived (fig. 25).

IMPLICATIONS OF THE BRAZOS RIVER HISTORY

INTRODUCTION

The previous sections of this paper present the principal evidence with which to reconstruct the history of the Brazos River. Drainage net, river and terrace geometry, river form, alluvium and terrace soils, bedrock and terrace composition, and river hydraulics may be combined to describe the Brazos River at various stages in its evolution.

EARLY ORIGIN

Fisher and McGowen (1969, p. 39) pointed out the marked coincidence between positions of modern coastal streams and the locations of stacked delta systems in the early Eocene Rockdale Formation. The Guadalupe, Colorado, Brazos, Trinity, Neches, and Sabine Rivers all flow along the axis of Eocene deltas that were prograding toward the Gulf (*ibid.*, fig. 19). The trends of the major drainage basins in Texas may have been established as early as Eocene.

Contours on the pre-Ogallala surface (Cronin, 1969) suggest a pre-Ogallala dendritic, eastward-flowing drainage system during Miocene or earlier time. The modern drainage in the upper basin still conforms to this pattern.

DEPOSITIONAL EVENTS

The earliest depositional event that relates directly to the history of the Brazos River is the formation of the High Plains surface. The history of the High Plains is in turn synonymous with the origin of the Ogallala Formation. Frye and Leonard (1959, p. 7) indicate that the mechanism of Ogallala deposition was one of valley alluviation. "In the High Plains belt, as alluviation proceeded through Neogene time more and more of the existing erosional topography was laterally overlapped, the lower divides were buried, and ultimately much of the region northward from Howard County, Texas to at least southern South Dakota became a coalescent plain of alluviation with only local areas on the highest of the former divides extending to or slightly above the regional surface" (*ibid.*). After the surface was developed it remained stable for a considerable time as indicated by the thick caliche zones in fossil soil horizons.

The Goliad Formation may be a coastal equivalent of the Ogallala. Near San Felipe, Austin County, in the west bank of the Brazos River the Goliad exhibits a sequence of fluvial sand and gravel capped by a thick caliche zone. The larger gravel fraction is composed of reworked caliche. The age of the Goliad, Neogene, is equivalent to that of the Ogallala. Deposits of Ogallala and Goliad type suggest wide bedload streams. Apparently the climate was becoming increasingly arid during Neogene time and tectonism was dying out in the source area for Ogallala sediments during the final stage of High Plains development.

Investigations of high gravel outcrops east of the Cap Rock Escarpment indicate the original extent of the High Plains in Texas. Frye and Leonard (1963) correlate the highest gravel deposits in Montague County 180 miles east of the High Plains edge with the Ogallala Formation. It has been previously mentioned (discussed with miscellaneous high gravel page 29) that the high gravel near Throckmorton, (fig. 4, locality 13 and fig. 20, profile I) about 100 miles east of the High Plains edge, is an Ogallala remnant. These deposits suggest that the Ogallala once covered most of north-central Texas between the Callahan Divide which rose above it on the south and the Red River on the north. The present major drainage trends of the modern Brazos River basin also support this hypothesis. A line of best fit for the trend of the trunk stream of the Brazos River between the coast and Mineral Wells, Palo Pinto County trends in a southeasterly direction. This dominant drainage direction apparently results from the gulfward slope of topography and rock structure (fig. 4). Northwest of Mineral Wells the Brazos River divides into several major forks. The dominant flow direction is more eastward than southeastward. (One exception, a northeast-flowing segment of the Clear Fork, is related to a Permian escarpment. This segment of river probably did not exist during the earliest basin history.) The eastward trend of upper basin rivers correlates with the slope of the old Ogallala surface that once covered this area.

Not only did the paleoslope of the Ogallala determine the direction of drainage for the upper basin, but its sediments were a major component of early flood plain deposits east of the Ogallala outcrop. Byrd (1971, p. 29), Menzer and Slaughter (1970, p. 292), and Lewand (1969, p. 23) have shown the correlation between mineralogy-petrology of coarse gravel suites of the Ogallala and high gravels such as the Uvalde.

Pliocene through earliest Pleistocene marked the development of widespread alluvial cover over most of the Brazos basin. The lower basin was blanketed with deposits of the Willis fluvial system which overlap the older Cenozoic deposits from the Oakville Escarpment coastward. These sediments were derived mainly from erosion of updip Tertiary formations of the Wilcox and Claiborne groups. Interpretation of depositional conditions during this period is dependent on climatic assumptions.

Caliche deposits in the Goliad and Ogallala Formations indicate aridity. The Willis Formation is believed to reflect the cooling trend that marks conditions prior to the first North American glaciation. Schumm (1968, p. 43) shows that the effects of a change in temperature on runoff and sediment yield is greatest for a change of climate in originally arid or semiarid regions. Assuming the average temperature during Neogene time to be 70° F and that by Willis time it had dropped 5° F, the effect would be increased rainfall and runoff with decreased sediment yield (figs. 28, 29 in Schumm, 1968). For example, if the annual rainfall was 25 inches in Neogene time and temperature dropped to a 65° F average along with a 5-inch increase of annual rainfall, runoff would increase by a factor of 7 from

0.5 to 3.5 inches and sediment yield would decrease from 825 to 525 tons per square mile of drainage basin (*ibid.*).

The changed hydrologic regime would demand channel adjustments. Modern bedload streams (those with high bedload to suspended load ratios) which are originally stable and then experience increased flow with decreased sediment loads become unstable, downcutting is reduced and channel widening by bank cutting predominates (Schumm, 1968, Table 5).

A model for streams that deposited the Willis fluvial material is one of high width-to-depth ratio, relatively steep gradient, and low sinuosity. The areal extent of deposits, nature of sediments, and sedimentary structures in the Willis Formation support this interpretation.

The upper basin during or slightly later than deposition of the Willis received a covering of fluvial deposits now called the Uvalde Gravel. West of the Balcones fault zone, the influence of the Ogallala sediments on the composition of the Uvalde is more pronounced. East of the fault zone the influence of cobble-size chert and flint derived from Lower Cretaceous rocks is more prominent.

After deposition of the Willis and Uvalde formations, three well developed Brazos River terraces (possibly a fourth) indicate a cyclic history of valley alluviation followed by valley cutting. This period of Brazos River development is definitely Pleistocene in age. Pleistocene fossil vertebrates have been recovered from all three terrace levels. (Figure 29 shows bone fragments found near San Felipe, Austin County, tentatively from the second terrace which crops out nearby.)

Individual terraces are both paired and unpaired depending on their location within the basin. The distance between paired terraces suggests that each new flood plain was narrower than the earlier flood plain. However, this is not interpreted as indicating significant size differences in the rivers associated with each terrace, since the width of a flood plain is also dependent on the span of time during which the river regime is stable enough to allow extensive lateral planation and sediment accretion. The practice of correlating terrace levels to Pleistocene glacial stages is possibly valid for the Brazos River terraces. Since early interglacial stages were longer than more recent ones, it should be expected that these longer time intervals would produce wider flood plains.

The gravel composition of each terrace has been summarized in figure 27. Limestone decreases upward through successive terraces and quartz and quartzite increase. Although the third terrace was not studied quantitatively, it appears that the same relationships are true for this level. South of the Cretaceous terrane limestone is almost absent in the third terrace, suggesting that Cretaceous terrane was not being extensively eroded by the river which deposited the third terrace.

Stream-worn limestone gravel along with fossil fragments of Late Cretaceous invertebrates (*Durania* sp., *Inoceramus*, sp., *Exogyra* sp., *Gryphaea* sp.) were collected in Miocene fluvial deposits of the Fleming Formation in Grimes County. This indicates that Cretaceous rocks had been unroofed by Miocene time, possibly reflecting Balcones fault activity. It has been suggested that the greater age of this terrace has permitted complete leaching of all carbonates, but occasionally stream-worn *Gryphaea* sp. are found in third terrace deposits. If these small pelecypods could survive leaching, then other limestone material should also survive.

The absence of the carbonate gravel suite from the upper terrace in Brazos, Grimes, and Waller Counties is here interpreted as the result of burial of local geology under the fluvial mantle described earlier (Willis and Uvalde). During Miocene time the Cretaceous rocks were being stripped, but during deposition of the third terrace the Cretaceous outcrop in the vicinity of major streams was protected by a fluvial blanket and not until this deposit was largely removed did older Cretaceous terrane contribute sediments to the Pleistocene Brazos River. Generally the third terrace is rich in chert, quartzite, and quartz as are the Uvalde and Willis Formations. The problem caused by similarity of Willis deposits and the third terrace has been discussed earlier. Terraces younger than the third level show a gradual increase of limestone gravel as more and more of the protective cover on the Cretaceous is eroded. The composition of the modern channel gravel is 60 percent limestone (fig. 27). Percentages of quartz and quartzite in the second terrace, first terrace, and channel deposits successively decrease as a result of downward reworking of older more siliceous terrace material and increase in locally derived material. The net effect is to produce a distinct gravel suite for each terrace level. Local sources of siliceous material other than older terrace deposits include conglomerates of Triassic, Pennsylvanian, and Early Cretaceous age, of which the Lower Cretaceous deposits are most important, contributing pea-size quartz and chert.

MULTIPLE TERRACES AND EUSTATIC CHANGES OF SEA LEVEL

Flood plains are transformed into terraces by tectonic, climatic, or man-induced changes which alter the regime of the river, causing it to entrench itself below its established bed and associated flood plain (Wolman and Leopold, 1957, p. 106). Major changes have occurred in the Brazos River basin several times as evidenced by terrace positions.

Bernard, LeBlanc, and Major (1962, p. 179) considered eustatic changes of sea level the prime factors which have caused terrace development in the lower Brazos basin. Lowering sea level caused entrenchment of valleys as stream base level was lowered, while rising sea level caused valley alluviation. "Deposits of the low and rising sea level substages in the entrenched valleys are predominantly coarser than those of the standing sea level substage because river gradient and the capacity of the stream to transport material were greater" (*ibid.*). The most recent sea level changes are described by McFarlan (1961, p. 129), who confirmed by radiocarbon dating that sea level stood 440 feet below its present level more than 35,000 years B.P. and a rise of sea level to -250 feet also occurred before 35,000 years B.P. About 18,500 years ago sea level began to rise and reached its present position approximately 5,000 years ago (*ibid.*). Coastal streams must have been effected by sea level fluctuations, but the question remains how far upstream would this cause river adjustment?

To answer this question an analogy is drawn from the effect of rising base level on longitudinal profile of a stream after dam construction. Closure of the dam causes sedimentation in the reservoir. Leopold *et al.* (1964, p. 259) point out that the aggradation effect is not felt upstream from the intersection of the sloping river channel and the backwater curve. "The fact that deposition appears not to proceed upstream even though the slope is flatter below indicates that the channel has adjusted in such a way as to transmit the sediment across the depositional reach and into the reservoir" (Leopold *et al.*, 1964, p. 261). This suggests that eustatic changes of sea level are absorbed by adjustment in river deltation and limited aggradation for a short distance upstream. Therefore, it is probable that sea level fluctuation is not a prime cause of flood plain aggradation in the Brazos River.

The effect of tectonism on the Brazos River's history is little known. Subsidence of the coastal plain has allowed superposition of Brazos fluvio-deltaic deposits, but terrace gradients are remarkably similar to modern river gradients suggesting a long-term stability.

MULTIPLE TERRACES AND PALEOCLIMATES

A major portion of this paper outlines evidence indicating prior existence of a much larger river. Analysis of meander scars, hydraulic geometry, and stream competence suggests that the present valley once contained a river with bankfull discharge 5 to 9 times greater than the present river, and thus the modern Brazos River is an underfit stream. This idea is not new. Matthes (1941, p. 635) believed that the Brazos River near Possum Kingdom Dam was inherited from an older, larger river. Dury (1964, p. B 39) suggests that the relationship of stream to valley meanders in the Brazos valley is underfit. Dury also interpreted the Mission River of south Texas, where it enters Mission Bay, as representative of a group of manifestly underfit streams which possess drowned valley meanders in their coastal reaches and are common on the south Texas coast (*ibid.*)

To illustrate the significance of the postulated flow of the ancestral Brazos at bankfull stage a relationship between basin area (A) and bankfull discharge (Q) was derived. Plotting bankfull discharge at a gaging station against the contributing basin area gives a line of best fit with the equation: $A = 68Q^{.6}$. This means to have produced the high postulated flows under existing rainfall conditions the basin area above the mouth of the Brazos would have had to be five times the present area, an excessive figure, in consideration of existing probable basin configuration.

The average annual temperature of the Brazos basin is approximately 65° F and varies from 59° F in the High Plains section to about 70° F for the Upper Coast section (Carr, 1967, p. 8). Average annual rainfall is about 30 inches, ranging from 18.5 inches on the High Plains to 46 inches on the Upper Coast (*ibid.*). A ten-year average of annual basin discharge taken at San Felipe gauging station is 6,299,000 acre feet. This amounts to a yearly runoff of 3.4 inches from a basin area of 43,660 square miles of which 9,240 square miles are considered noncontributing. Runoff is approximately 11 percent of the average annual rainfall.

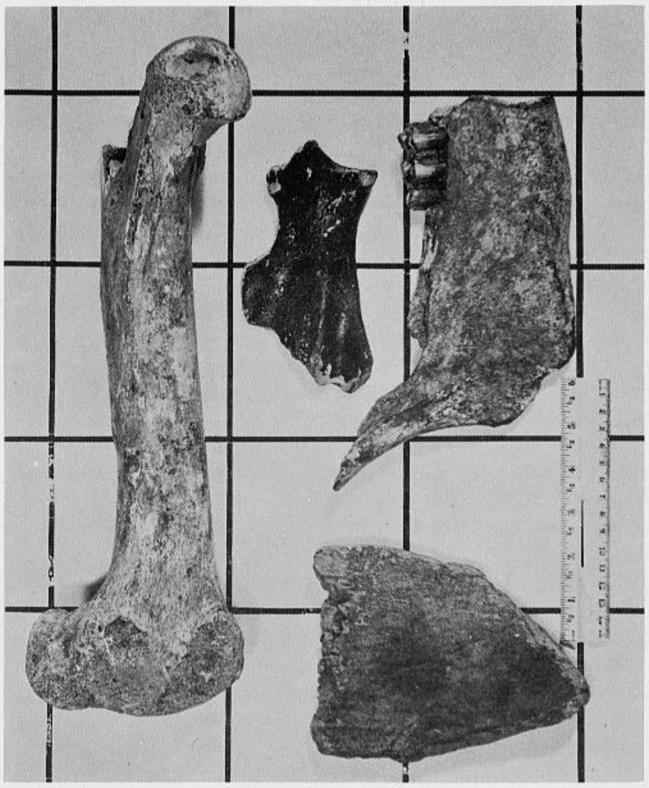

Fig. 29. Pleistocene bone fragments, Austin County. Found in channel in channel gravel (fig. 17) adjacent to Stephen F. Austin State Park. Left, modern horse. Top center, scapula; fossil of unknown affinity. Top right, jaw of extinct *Bison* sp. Bottom, fragments of carapace from extinct giant turtle *Geochelone* sp. This turtle could not survive unless the climate was frost free without a winter freeze (Slaughter, 1972, oral communication).

Bankfull flow of the ancestral Brazos was not related to present climatic conditions. Discharge during earlier periods represented the effect of climatic changes associated with glacial stages. Factors which operated during glacial advances to promote high discharges were: 1) reduced air temperature, 2) increased precipitation, 3) higher runoff, 4) increased storm frequency, 5) increased soil moisture, 6) changed vegetal cover, and 7) change of rainfall intensities.

These effects are most dramatically demonstrated by the existence of large lakes on the High Plains during periods of Pleistocene glaciation. Lake Blanco (Nebraskan), Lake Tule (Kansan), Lake Tahoka (Wisconsin), and Lake Lomax (Wisconsin) represent High Plains pluvial periods. The present runoff from the High Plains based on temperature-rainfall vs. runoff curves is 0.7 inch (fig. 28 in Schumm, 1968).

Dillon (1956) inferred reduction in mean annual temperature of 10° F at latitudes 35°-40° N during Wisconsin glaciation. Antevs (1954) suggested an 8-inch increase in annual precipitation and a 10°-16° F reduction in summer temperature to restore pluvial Lake Estancia in New Mexico. Reeves (1966, p. 287) believed it would require an 18° F decrease of summer temperature and an annual rainfall of 33-36 inches to return the High Plains section of the Brazos River to its pluvial conditions. Under these climatic conditions runoff from the High Plains would be approximately 10 inches. Near the Gulf of Mexico Wisconsin glaciation would lower the annual temperature 13° F (Manley, 1955), while at the same time causing an increase in rainfall.

The present pollen rain on the High Plains exhibits little pollen from conifers, but clay associated with Wisconsin Lake Tahoka in Lynn County contains 96 percent pollen (Reeves, 1966, p. 276). The presence of pines on the High Plains is botanical evidence of the higher rainfall conditions in pluvial periods.

Presently 900 square miles of the upper Brazos basin does not contribute to flow in the middle and lower basin. However, Reeves (1966, p. 281) suggested that some lake basins on the High Plains were developed along old stream courses during Pleistocene pluvials. These lakes were linear and appear to have been connected by drainage channels leading eventually to eastern drainage.

This source of runoff during Pleistocene pluvials, added to the increases in runoff in the middle and lower basin is offered as an explanation for postulated unusually high bankfull discharges associated with the large meander scars. Another important result of high rainfall would have been increased vegetal cover causing decrease of sediment yield and ultimately decreased stream load.

Valley entrenchment and flood plain abandonment within the Brazos River basin was apparently caused by changes in regime associated with glacial maxima. For each major glacial advance the river adjusted to increased bankfull discharge and decreased sediment load. Stream adjustment under these conditions involved down cutting below the adjacent flood plain. Channel widening accompanied bed erosion if the stream carried a mixed load (between 3 and 11 percent bedload) and minor channel widening occurred if the stream carried dominantly suspended load (Schumm, 1968, p. 40).

The interglacial stages were characterized by decreased rainfall and runoff with increased temperature and sediment yield. During the interglacial stages the Brazos River became underfit to the valley cut by the prior stream but comparatively more stable. This period of dynamic equilibrium allowed the formation of a thick valley fill and a broad flood plain which was then dissected by the river when it adjusted to the next glacial advance.

CONCLUSIONS

1. Meander scars in the Brazos valley between Waco and the coast indicate that the modern Brazos River is an underfit stream. The radius of curvature of the large scars suggests that the width of the ancestral Brazos was approximately 2600 feet at bankfull stage.
2. Ancient channel characteristics derived from meander scars in the valley walls inserted into hydraulic equations derived from modern flow records indicate the ancestral Brazos River had a bankfull discharge 5 to 9 times greater than the present river at Bryan, Texas.
3. Deposits of lateral accretion (channel gravel, lower point bar, and levee) are the dominant depositional units within the modern flood plain. Thick clay fills found in abandoned channels are the only significant overbank deposits. Indian artifacts that are associated with the flood plain deposits of the lower basin suggest a maximum rate of overbank deposition of 2 feet per thousand years.
4. Modern point bar deposits indicate that the Brazos River is a mixed load river (3 to 11 percent bedload). Comparison of these deposits to exposures of older alluvium within the flood plain suggests that the Brazos River has changed regime since deposition of the older alluvium.
5. Sites of Brazos River deltation and delta plain construction have progressively shifted toward the southwest. The Oyster Creek meander belt represents an abandoned Brazos channel which formerly terminated in a delta several miles northeast of the modern Brazos delta.
6. Annual suspended load for the Brazos River is 760 tons per square mile of drainage basin. This is equivalent to lowering the land surface of the whole basin one foot per thousand years, if total load is considered.
7. The Little Brazos River has a history of development related to creeks entering the Brazos valley along its eastern side and does not represent an old abandoned Brazos channel.
8. At least four levels of alluvial deposits are found within the Brazos River basin, but only three levels are well represented throughout the basin. The top of the first terrace escarpment is about 30 feet above the adjacent flood plain. The bases of the second and third terrace are 45 to 60 feet and 90 to 120 feet above the flood plain respectively. High gravel deposits occur on points ranging from 200 to 500 feet above the local river level.
9. Soils associated with terraces in the middle and lower basin reflect the age of the terrace by their maturity. High terraces have mature soils with well-developed profiles such as the red-yellow podzols, and lower terrace soils are thinner with poorly developed profiles.
10. The map unit called the Seymour Formation represents three and possibly four different periods of valley alluviation ranging in age from Sangamon to pre-Kansan. Slope orientation of the erosional surface below the Seymour Formation near Rule, Haskell County, suggests that the Double Mountain Fork was previously a tributary of the Clear Fork which was captured by the Salt Fork in early Pleistocene.
11. The High Plains surface once extended more than 100 miles east of the Cap Rock Escarpment. This conclusion is suggested by correlation of high lag

gravel 12 miles west of Throckmorton, Throckmorton County, with the projected position of the Ogallala Formation.
12. Incised meanders of the Possum Kingdom area are second cycle in origin. Incision of this section may have been caused by flow regime changes related to capture of the Clear Fork from the Leon River in early Pleistocene time.
13. Statistical analysis of gravel (0.37 to 0.53 inch fraction) from the channel, first terrace, and second terrace indicates significant differences in composition between these levels. This criterion can be used to correlate terrace levels at different localities and compositional changes also indicate that geology of the provenance area changed as the river evolved.
14. Large cobbles (5-inch diameter) found in older alluvium (possibly middle Wisconsin) of the modern Brazos River flood plain suggest a river of greater competence than the present Brazos River. The tractive force needed for movement of these cobbles would require a river both steeper and deeper than the modern Brazos near Bryan, Brazos County, at bankfull stage. This evidence supports that of the meander scars which also indicated a larger ancestral Brazos. Larger cobbles occur in the Uvalde Formation and indicate either prior existence of a completely different type of river (proportions of bedload and suspended load) or a river of much steeper gradient (approximately 5 feet per mile).
15. The Brazos River drainage trend was established as early as Miocene time, as indicated by buried valleys beneath Ogallala cover, and may have been established by Eocene time, as indicated by delta locations in the Rockdale Formation.
16. Caliche deposits in the Goliad and Ogallala Formations reflect the widespread arid conditions of pre-Pleistocene time. This arid period was followed by climatic cooling resulting in deposition of the Willis Formation. The Willis Formation represents a transitional unit between late Tertiary deposition and the multiple cycles of valley alluviation during Pleistocene time. Streams that deposited the Willis Formation had greater width-to-depth ratios and were steeper than streams that deposited the Brazos River terraces.
17. Although the Cretaceous terrane was unroofed as early as Miocene time, the rarity of limestone gravel in the high terrace of the lower valley indicates that a thick alluvial mantle protected this area during late Pliocene through early Pleistocene time. The mantle included deposits of Uvalde Gravel and the Willis Formation.
18. Eustatic sea level changes did not effect stream development except near the point of changing base level. Therefore, sea level fluctuations are not the cause of terrace development within the Brazos valley.
19. Evidence suggests that the ancestral Brazos was much larger than the modern river. Cyclic alteration of the Brazos River's size was caused by climatic changes at glacial maxima. Higher runoff and lower sediment yield associated with glacial stages caused valley entrenchment and flood plain abandonment. Interglacial stages were characterized by decreased runoff and increased sediment yield which resulted in the Brazos River becoming underfit to the valley of the former stream. During interglacial times broad flood plains were developed which were dissected by the river when it adjusted to the next glacial advance.

SUGGESTIONS FOR FUTURE RESEARCH

This study offers a minimum outline of developmental history of the Brazos River system and facts pertinent to that history. Throughout the investigation problems beyond the scope of the present work were recognized. Some of these are listed below in order to stimulate continued investigation of the Brazos River and related problems.

1. At the beginning of this project, Horton morphometric analysis of drainage-basin development was attempted using 1:500,000 base maps. After considerable labor it was established that the drainage network at this scale was not detailed enough to provide significant quantitative data. I believe 1:250,000 base maps would remedy this problem and allow comparison of subbasins in the Brazos River system. Preliminary investigation of the Navasota River showed enough peculiarities for this river basin to merit special attention. This basin is unusually long and narrow and appears to be deficient in low order stream drainage. This river is the last major tributary of the Brazos River that is still uninfluenced by man. The proposed Millican Dam project may soon alter this situation.
2. The Navasota River has a well-developed set of terraces. Lithologies of high terrace gravel in the upper reaches are foreign to the basin. A study of the terraces and composition of their gravel might reveal important drainage diversions.
3. The upper terrace of the lower basin contains abundant pebbles of jasper veined with white quartz. Determination of the source of this distinctive gravel might be valuable in establishing basin configuration.
4. Detailed examination of gravel from the upper terrace near Bryan, Brazos County, showed that some of the gravel contains silicified fusulinids and tabulate corals. Thin section study of this fauna might reveal provenance and thus relate the terrace deposits to a specific Paleozoic terrane.
5. A detailed study of cover sands and possible buried alluvium on divides near the Salt and Double Mountain Forks might reveal stratigraphic relationships and relation to river history.
6. A detailed surface and subsurface study on the history of the Seymour Formation would contribute significantly to understanding the Pleistocene history of the upper basin.

7. The significance of the high gravel on the Callahan Divide is as yet uncertain. It clearly relates to a much earlier fluvial history than any considered here.
8. The same may be said of the dissected high surface near Possum Kingdom Dam. Is this surface possibly correlative with the lag on top of the Callahan? If so, what relationship does it bear to earlier Tertiary Gulf Coast stratigraphy?
9. Does the Uvalde Gravel east of the Balcones fault zone reflect late Neogene tectonism or climatic conditions characterized by torrential flow?

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APPENDIX A

MEAN CURVATURE RATIO OF BRAZOS RIVER

Measured Meander Locations U.S.G.S. 7½ minute maps	R_m:W*
1. Daniels	1.5
2. Daniels	2.7
3. Daniels	1.8
4. Daniels	1.7
5. Daniels	1.6
6. Marlin	2.0
7. Gause	2.7
8. *Clay*	*2.8*
9. Clay	3.1
10. Howth	3.3
11. East Columbia	3.0
12. Buckhorn	3.7
13. Brazoria	3.3
14. Otley	2.9
15. Thompsons	3.3
16. Cedar Springs	2.5
17. Mumford	3.0
18. Baileyville	3.6
19. *Baileyville*	*2.8*
20. Wellborn	2.2
21. Sunny Side	4.5
Average	*2.8*

* R_m = mean radius of curvature
W = width at bankfull stage

APPENDIX B

MEANDER SCAR MEAN RADIUS OF CURVATURE

Meander Scar Locations U.S.G.S. 7½ minute map	Radius of Curvature
Baileyville and Maysfield	3 1/16 inches
Cedar Springs	5 8/16 inches
Cedar Springs and Marlin	3 inches
Marlin	5 8/16 inches
Riesel and Marlin	5 8/16 inches
Maysfield and Calvert	5 8/16 inches
Hearne South	4 1/16 inches
Hearne South and Bryan West	4 8/16 inches
Mumford and Bryan West	3 4/16 inches
Tunis	4 2/16 inches
Chances Store and Tunis	5 11/16 inches
Clay and Snook	4 4/16 inches
Clay	2 6/16 inches
Millican	4 14/16 inches
San Felipe and Sunny Side	3 12/16 inches
San Felipe	2 10/16 inches
Gholson*	6 3/16 inches

Average radius in unconverted scale --- 4.34 inches
Map scale: 1 inch = 2,000 feet.
Average radius of curvature of meander scars = 8680 feet.
*Meander scar is in bed rock and may not be representative.

APPENDIX C

HYDRAULIC GEOMETRY OF BRAZOS RIVER

Station Location	Mean Annual Flow cfs	Width ft	Depth ft	Velocity ft/sec
Double Mt. Fork at Justiceburg	36	37	.70	1.45
Double Mt. Fork near Aspermont	179	82	1.10	1.94
Brazos River at Seymour	418	171	1.41	1.70
Brazos River near South Bend	934	203	2.70	1.80
Brazos River near Palo Pinto	1128	289	1.80	2.59
Brazos River at Glen Rose	1585	251	2.20	3.00
Brazos River at Whitney	1669	261	3.90	1.74
Brazos River at Waco	2510	331	3.50	2.16
Brazos River near Bryan	5407	269	6.30	3.33
Brazos River near Hempstead	6676	274	11.0	2.27
Brazos River at Richmond	7310	346	10.0	2.15

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