

# Record of an Early Cretaceous Marine Transgression—Longford Member, Kiowa Formation

Bulletin 219

Paul C Franks

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# Paralic to Fluvial Record of an Early Cretaceous Marine Transgression—Longford Member, Kiowa Formation, North-Central Kansas

By

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## EXECUTIVE SUMMARY

This report describes, names, and treats the origins of the Longford Member of the Kiowa Formation. Longford rocks form a distinctive part of the Kiowa Formation and include siltstone, relatively scarce sandstone, minor lignite, and abundant clay rocks. They mark the base of the Kiowa Formation in parts of north-central Kansas and are named for the excellent exposures near Longford, southwestern Clay County. The member also crops out in parts of Washington, Ottawa, Dickinson, Saline, and Marion counties. Longford sediments were deposited more than 100 million years ago during the Mesozoic Era, specifically during late Early Cretaceous time when much of what we now call Kansas was covered by the Kiowa sea.

Clay rocks of Cretaceous age have been an important economic resource in north-central Kansas, especially since World War II. Shale in the Kiowa Formation is used to make light-weight, bloated aggregate. Clay rocks in the Dakota Formation, which overlies the Kiowa Formation, are used to make bricks. Knowledge of Kiowa and Dakota stratigraphy has aided in the utilization of those ceramic resources. The essentials of Kiowa and Dakota stratigraphy were outlined by Norman Plummer and John F. Romary in 1942 (Kansas Geological Survey Bulletin 41, Part 9), and this report refines and extends some of the groundwork laid by them. Although this report does not deal specifically with the ceramic properties of Longford clay rocks, it does note that Longford rocks resemble and have been confused with those in the Dakota Formation. Longford clay rocks, like those near the base of the Dakota Formation, are demonstrably lenticular, and they contain variable proportions of the ceramically important clay minerals kaolinite and smectite.

Longford sediments were deposited in broad stream valleys, in estuaries, and in lagoons or bays along sandy barrier coasts as the Kiowa sea flooded into what is now north-central Kansas. The top of the Longford Member is marked by nearly white to pale yellowish-brown siltstone that formed from well-washed sediments that were deposited behind Kiowa barrier bars and islands along the landward, tide-influenced shores of lagoons or bays. The distribution of Kiowa shale and sandstone above the Longford Member offers insight into the mechanism by which Kiowa shorelines shifted landward, and that mechanism seems to differ significantly from the major mode of marine transgression that is inferred for many geologically young barrier shorelines, like those along the modern Atlantic coast of North America. Clay rocks in the Longford Member range from those that are light gray and blotched by striking, red, hematitic mottles to those that are nearly black and highly plastic. At least some of the clay rocks originated as floodplain or other overbank deposits of streams, but they were modified into immature tropical alluvial soils by the warm, humid climate that prevailed during Kiowa time.

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## ABSTRACT

A distinctive facies of the Albian Kiowa Formation is exposed along the eastern fringes of the Kiowa outcrop belt in north-central Kansas. The name "Longford Member of the Kiowa Formation" is proposed for the distinctive unit, which is as much as 100 ft (30 m) thick. Conspicuous, nearly white siltstone as much as 15 ft (4.6 m) thick marks the top of the member. The lower parts of the member consist of clay rocks, siltstone, sandstone, and lignite. Red-mottled clay rocks, red-mottled siltstone, and the nearly white siltstone at the top of the member form a striking contrast to drab shale and yellowish-gray to brown sandstone in other parts of the Kiowa Formation. The Longford Member rests with transgressive disconformity on Lower Permian rocks. Along most of the outcrop belt, it is overlain conformably by higher parts of the Kiowa Formation. About 30 mi (48 km) south of the Kansas-Nebraska border, near the northward pinchout of the Kiowa Formation, the member is overlain by the Dakota Formation (Albian? and Cenomanian). The siltstone at the top of the Longford Member is the northernmost Kiowa rock that can be identified with certainty as the formation pinches out.

Longford sediments were deposited as the Kiowa sea transgressed onto the eroded, gently dipping, western flank of the Nemaha anticline. Kiowa barrier-bar systems and the topography developed on Permian bedrock influenced both the thickness and the depositional environments of the Longford Member. The siltstone that marks the top of the unit stems from sedimentation along the landward, inner shores of lagoons or bays that formed behind Kiowa barrier bars. Sedimentary structures and fossil reeds and rootlets suggest that the inner shores were affected by tides.

The lower parts of the Longford Member formed from sediments that accumulated in fluvial and estuarine realms that developed in broad valleys eroded into Permian bedrock. Log-probability plots of grain-size distributions support a fluvial origin for most Longford sandstone. Red-mottled, kaolinitic and smectitic clay rocks and scarce red-mottled siltstone represent floodplain deposits. The red mottling accords with gleying in immature, tropical soils. A pedogenic origin for the red-mottled clay rocks also is indicated by plexoidal and domain clay fabrics similar to those common in soils, and by "clay skins" around detrital quartz grains. Lenses of dark, smectitic claystone may be fossil vertisols that developed in floodbasins or on estuarine bayhead deltas. Lignite seams and lenses represent floodbasin deposits, but some of them may have formed from detrital accumulations of plant debris on estuarine delta plains. Sequences of light-gray siltstone may also represent estuarine sediments.

Light-gray, smectitic, commonly silty Kiowa shale overlying the Longford Member probably formed from lagoon or bay sediments. Kiowa rocks above the lagoon or bay deposits indicate that transgression was accompanied by in-place growth and eventual submergence of Kiowa barrier systems. Submergence resulted in landward shifting of surf zones and in construction of new barriers near the former inner shores of

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lagoons or bays. Open-sea, illitic, Kiowa shale overlies the lagoon or bay deposits in many places. Shoreface erosion, or other processes of ravinement, did not lead to the development of either marked or extensive disconformities during transgression. Instead, remarkably complete transgressive records of fluvial, estuarine, lagoon, and bay sedimentation were preserved.

Fossil soils along the Permian-Cretaceous unconformity in central Kansas indicate that the kaolinite and smectite in Longford clay rocks were reworked from similar soils developed on Paleozoic source rocks in the continental interior. Clay-mineral assemblages in facies-equivalent Longford and Kiowa rocks accord with, but do not prove, derivation of the abundant illite in Kiowa shale from sources that were different from those that supplied kaolinite and smectite to Longford depositional systems. Longford sandstone was derived mainly from Paleozoic terrains in the continental interior, but staurolite grains in the sandstone suggest that some of the sand ultimately came from the crystalline terrains of the central Appalachian Mountains.

## INTRODUCTION

The Kiowa Formation (Early Cretaceous, Albian) of Kansas (Cragin, 1889, 1894) is a heterogeneous unit made up of shale and other clay rocks, siltstone, sandstone, and coquinoid limestone or mollusk-shell conglomerate. These rocks formed from sediments that were deposited in a complex of marine and marginal-marine environments as the Early Cretaceous sea of the southern Western Interior spread northeastward across gentle terrain developed mainly on Permian rocks, and onto the gently dipping western flank of the Nemaha anticline in central Kansas. A conspicuous siltstone marks the lower parts of the Kiowa Formation along most of the eastern fringes of its outcrop belt in north-central Kansas (Figs. 1, 2). The siltstone rests directly on weathered Lower Permian rocks in places, but, more commonly, it overlies distinctive assemblages of claystone, mudstone, siltstone, sandstone, and carbonaceous beds that rest unconformably on Permian rocks. The assemblage of rocks below the conspicuous siltstone locally encloses and contrasts sharply with lenses or tongues of more typical, olive-gray, facies-equivalent Kiowa shale. The name "Longford Member of the Kiowa Formation" is proposed for the conspicuous siltstone and the underlying assemblages of Kiowa rocks. Except for informal use of the name "Longford" to refer to the same body of rocks (Franks, 1966, 1975), the name has not been applied to stratigraphic units in North America (G. V. Cohee, 1969, written communication).

A formal stratigraphic name is appropriate for Longford rocks because they are important to an understanding of Kiowa stratigraphy and sedimentation:

(1) Longford rocks form a recognizably different and distinctive part of the Kiowa Formation.

(2) The member is of limited stratigraphic and geographic extent on the outcrop in north-central Kansas (Figs. 1, 2).

(3) Despite their restricted occurrence near the base of the Kiowa Formation, Longford rocks have been mistaken for parts of the Dakota Formation (Early? and Late Cretaceous), which overlies the Kiowa Formation. Schoewe (1952), for example, assigned lignitic beds in the Longford Member in Dickinson County (Fig. 1) to the Dakota Formation. Appreciable thicknesses of overlying Kiowa rocks were overlooked during geological mapping of Clay and Ottawa counties, apparently because Longford rocks were mistaken for parts of the Dakota Formation (Walters and Bayne, 1959; Mack, 1962). Had the Longford Member been recognized as a formal stratigraphic unit at the time those projects were under way, the confusion probably could have been avoided.

(4) Some Longford clay rocks and siltstone resemble rocks in the Cheyenne Sandstone (Early Cretaceous), which underlies the Kiowa Formation in parts of southern Kansas (Figs. 1, 2). It might be tempting, therefore, to treat Longford rocks as part of the Cheyenne Sandstone, even though the outcrop areas of the two units are isolated from each other, and no direct stratigraphic connection seemingly exists between the units in the subsurface (Franks, 1975; Scott, 1970b).

(5) The distribution of Longford rocks on the outcrop and the depositional strike inferred for the Kiowa Formation (Franks, 1966, 1975) indicate that Longford rocks, or their analogues, may be present in the subsurface of southern or central Nebraska. Recognition of the Longford Member in Kansas, therefore, may aid in unravelling Cretaceous stratigraphy and sedimentation in areas to the north of Kansas.

(6) Lastly, although environmental interpretations technically play no part in the definition of formal stratigraphic units, Longford rocks warrant special attention because they constitute the easternmost record of paralic and continental sediments that accumulated during transgression of the Kiowa sea.

Longford rocks and facies-equivalent parts of the Kiowa Formation form an unusually complete sedimentary record of marine transgression along a barrier-island coast. Papers by Fischer (1961), Kraft (1971), Ryer (1977), and Swift (1968) describe some of the sedimentary records left by marine transgression along barrier-island coasts. One of the chief elements of transgression noted by those authors is erosion of barrier-island, lagoon, bay, inner shore, and even continental deposits as barrier systems migrate landward. The erosion leads to the development of transgressive disconformities within and beneath the transgressive

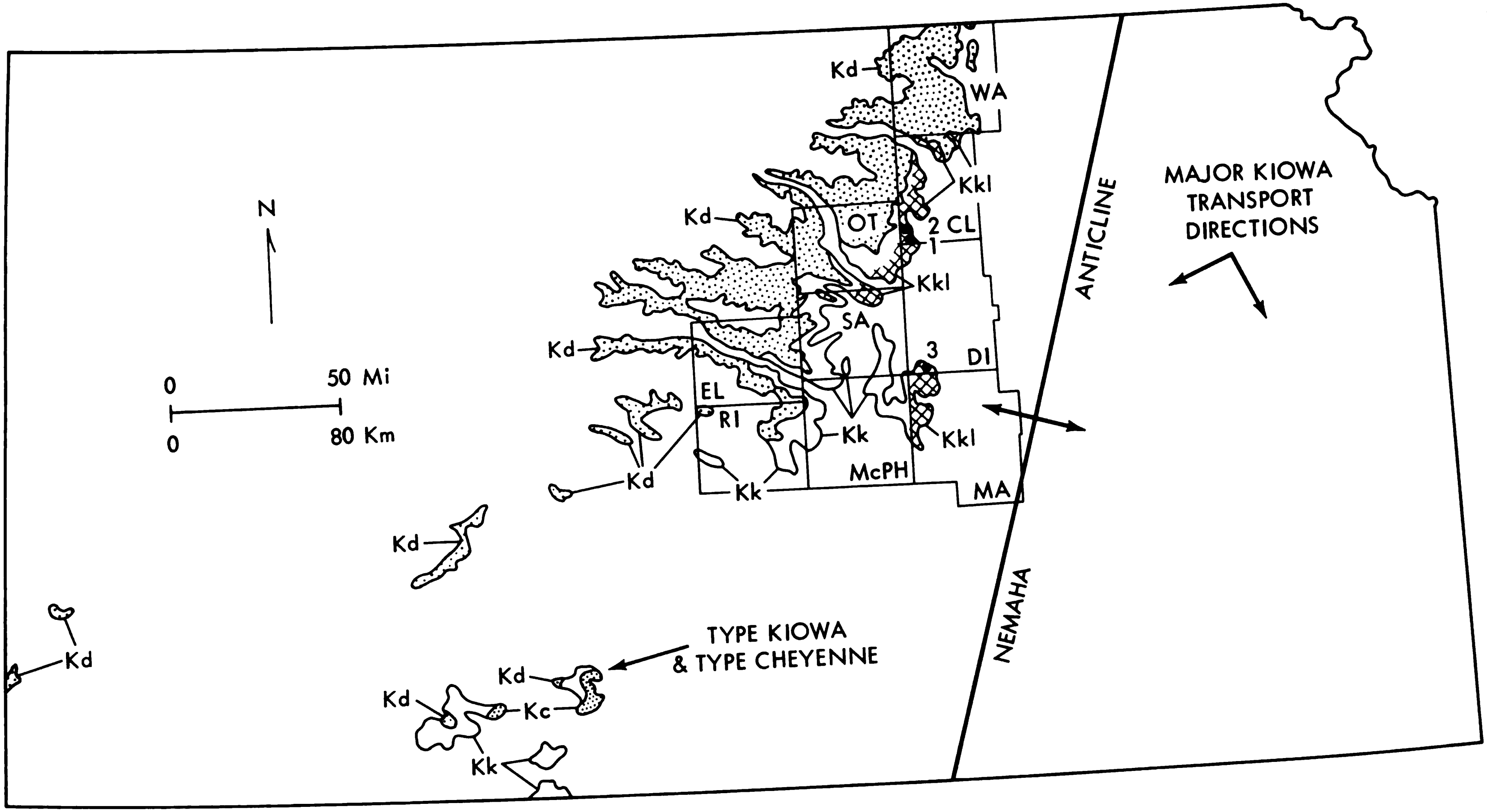



FIGURE 1.—Index map of Kansas showing generalized outcrop belts of Lower Cretaceous Kiowa Formation (Kk, unpatterned), its Longford Member (Kkl, cross-hatched), overlying Dakota Formation (Kd, coarse stipples), and underlying Cheyenne Sandstone (Kc, fine stipples), as well as approximate axial trace of Nemaha anticline. Arrows indicate orientation of major modes of cross-strata dip-bearings measured in Kiowa sandstone: arrow pointing to southwest shows direction of inclination of depositional slope. CL, Clay County; DI, Dickinson County; EL, Ellsworth County; MA, Marion County; McPH, McPherson County; OT, Ottawa County; RI, Rice County; SA, Saline County; WA, Washington County. Numbered dots mark locations of measured sections described in Appendix B and shown in Plate 2.

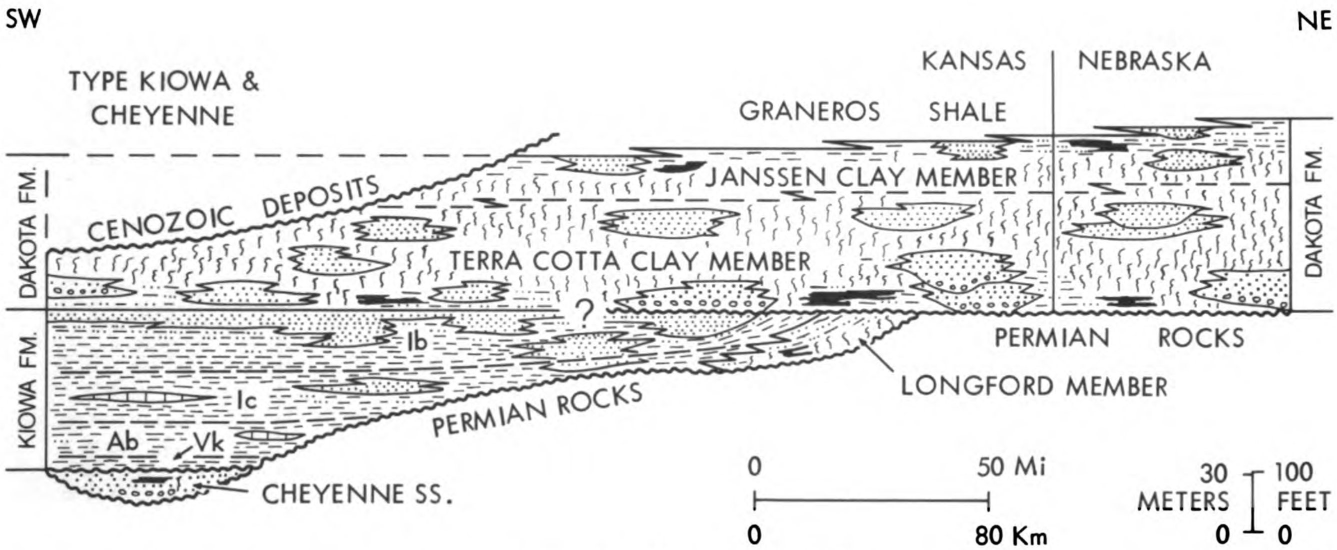



FIGURE 2.—Schematic cross section, showing stratigraphic relations of basal Cretaceous beds cropping out in southern and north-central Kansas. At least locally, Dakota Formation rests disconformably on Kiowa Formation; question mark indicates uncertain extent of disconformable contact. Biostratigraphic zones in Kiowa Formation after Scott (1970b): Vk, *Venezoliceras kiowanum* zone; Ab, *Adkinsites bravoensis* zone; Ic, *Inoceramus comancheanus* zone; Ib, *Inoceramus bellvuensis* zone.

sequence and results in varying degrees of destruction of the sedimentary records generated in coexisting marginal-marine, paralic, and fluvial environments of deposition. Consequently, offshore marine sediments commonly rest disconformably on coastal-plain deposits (Ryer, 1977, p. 183). The Longford Member and associated Kiowa rocks, in contrast, are notable for the absence of either marked or extensive transgressive disconformities that separate them from overlying open-sea deposits, and for the preservation of barrier-island, bay or lagoon, inner shore, estuarine, and fluvial deposits. The completeness of the Longford record also offers insight into the mechanisms by which Kiowa shorelines shifted landward during transgression.

Interpretation of the record of sedimentary environments preserved in Longford rocks, however, is not easy. Except for plant fossils, mainly of uncertain affinities, no environmentally significant fossils, either vertebrate or invertebrate, were found. Moreover, except for the siltstone that marks the top of the member, Longford rocks are not especially well exposed. Consequently, interpretations are based largely on scattered exposures, few of which span appreciable thicknesses of the member. Therefore, vertical and, particularly, lateral sequences, other than rather general ones, are hard to establish. Nonetheless, analysis of the Longford record depends on extensive application of Walther's law of facies succession (Walther, 1894). That principle states, in effect, that the records of depositional environments that existed side-by-side are preserved in vertical sequences of strata as a result of lateral migration of the environments with time. Thus, lateral facies relationships can be deciphered by study of vertical sequences of beds, provided that no significant erosional discontinuities interrupt the vertical successions of strata (Selley, 1976, p. 309-310).

Interpretation of the Longford depositional record also is handicapped because many of the depositional models that have been devised for the study of ancient rocks are based primarily on research in modern environments where relatively thick and complete progradational sequences formed (cf. Klein, 1971, 1972; Mathews, 1974; Selley, 1970). Because Longford sedimentation took place as the Kiowa sea transgressed landward, thick progradational sequences of strata had little opportunity to form. Not only are records of many of the individual depositional environments thinned, but also the sequences of beds in which some of the environments are recorded are upside down compared to the sequences in widely used progradational models.

This report summarizes a reconnaissance study of Longford rocks that stems from a larger study of the Kiowa and Dakota formations of central Kansas (Franks, 1966, 1975). Pertinent details of work methods used in the study of Longford rocks are given in Appendix A, whereas details on work methods used in the study of associated Kiowa sandstone and shale are in the 1966 report. Descriptions of measured sections are given in Appendix B. Nomenclature for formations follows current Kansas Geological Survey practice (Zeller, 1968). The name "mudstone" is used as a field term to refer to nonfissile sedimentary rocks composed primarily of subequal proportions of clay- and silt-sized particles. Usage generally follows that of the first definition in the *Glossary of Geology* (Gary and others, 1972). The term "claystone" is applied to nonfissile rocks composed mainly of clay-sized particles, and the term "clay rock" is used generically to encompass mudstone, claystone, and shale. Terminology for stratification generally follows that of McKee and Weir (1953). Part of the report is devoted to a brief description of the stratigraphic framework represented by the Cheyenne, Kiowa, and Dakota formations in Kansas because of the importance of these rock units to an understanding of Longford stratigraphy and sedimentation. Review of the salient features of the Cheyenne, Kiowa, and Dakota formations also is appropriate because of the similarity of some Longford rocks to rocks in those units.

# STRATIGRAPHIC FRAMEWORK

## THE BASAL CRETACEOUS BEDS OF KANSAS

The basal Cretaceous section in Kansas includes, in ascending order, the Cheyenne, Kiowa, and Dakota formations (Fig. 2). Outcrops of the Cheyenne Sandstone, which probably is early late Albian in age (Scott, 1970b; Berry, 1922), are restricted to the vicinity of its type area in southwestern Kansas (Fig. 1). The Kiowa Formation, which is of late Albian age (Scott, 1970a, 1970b), is exposed in both southwestern and north-central Kansas. The Dakota Formation, which probably spans the Albian-Cenomanian boundary (Eicher, 1965; Franks, 1966, 1975), crops out chiefly in north-central Kansas. The three formations are separated from underlying Permian rocks by a transgressive disconformity along which they show progressive northeastward overlap of older strata (Fig. 2). The Cheyenne, Kiowa, and Dakota formations record an initial transgressive-regressive cycle of Cretaceous sedimentation along the Kansas portions of the Early to mid-Cretaceous seaway that transected western North America (Franks, 1966, 1975; Scott, 1970b). The rocks of the Longford Member of the Kiowa Formation are the easternmost record of continental and paralic sediments deposited during the transgressive phase of the Cheyenne-Kiowa-Dakota cycle.

**Cheyenne Sandstone**—In the type area of the Cheyenne and Kiowa formations (Fig. 1), the Cheyenne Sandstone marks the base of the Cretaceous System. It rests on Upper Permian rocks and has a maximum thickness of about 94 ft (29 m) (Latta, 1948). West of the type region, the Cheyenne Sandstone is absent and the Kiowa Formation rests directly on Permian rocks. No Cheyenne rocks are recognized in scattered Cretaceous outliers south and west of the type area, either in Kansas or in nearby parts of Oklahoma (Franks, 1975; Scott, 1970a, 1970b). The Cheyenne Sandstone thins eastward from its type area, and both the Cheyenne and Kiowa formations disappear northward beneath an extensive blanket of Pleistocene deposits. About 20 mi (32 km) northward from its type locality, the Cheyenne Sandstone apparently pinches out beneath the Kiowa Formation along the western flank of the Pratt anticline in the subsurface of northwestern Pratt County (Layton and Berry, 1973). The formation does not crop out in central Kansas (Franks, 1966, 1975; Plummer and Romary, 1942; Scott, 1970a, 1970b), but Fent (1950, p. 56) suggested that conglomeratic beds of varicolored sandstone and shale (and presumably also white siltstone) exposed along the Permian-Cretaceous unconformity at several places in eastern Rice County (Fig. 1) might be outliers of Cheyenne Sandstone. The beds, however, are basically similar to those seen elsewhere in central Kansas where pebbles of chert and other resistant materials are concentrated along the Permian-Cretaceous unconformity.

The name Cheyenne has been applied to widespread sandstone in the subsurface of western Kansas, but Scott (1970b), using wireline logs, was unable to correlate at least some of that sandstone with type Cheyenne Sandstone. Sequences of nearly white to greenish-gray siltstone, shale, and fine-grained sandstone at the base of the Kiowa Formation in the shallow subsurface to the north and northeast of the type area have been correlated with the Cheyenne Sandstone (Fent, 1950; Latta, 1950; McLaughlin, 1949). Other than position at the base of the Cretaceous System, however, the sequences of silty to shaly and sandy beds may have little in common with rocks exposed in the type area of the Cheyenne Sandstone. The possibilities of overextension and miscorrelation of the formation warrant special consideration in applying the name Cheyenne Sandstone in the subsurface of western Kansas (Franks, 1975).

The abundant sandstone and relatively scarce siltstone and clay rocks in the Cheyenne Sandstone form an upward fining sequence that probably was deposited in fluvial and estuarine environments, largely before the late Albian Kiowa sea invaded southwestern Kansas (Franks, 1975; Scott, 1970b). Silty clay rocks intercalated with sandstone in the lower parts of the formation are light gray, but, in places, they are blotched by reddish-brown hematitic mottles. They resemble clay rocks in both the Dakota Formation and the Longford Member of the Kiowa Formation. Clay rocks higher in the Cheyenne Sandstone tend to be dark gray to brownish gray, and to contain abundant carbonaceous matter. They too resemble clay rocks in parts of the Dakota Formation and the Longford Member of the Kiowa Formation. Cheyenne clay rocks are composed chiefly of smectite and kaolinite. They contrast with the illitic and chloritic clay rocks of the Permian System in Kansas, but, like Longford clay rocks, they are compatible with derivation by reworking of fossil soils such as those that are preserved along the Permian-Cretaceous unconformity in Kansas (Franks, 1975).

**Kiowa Formation**—The Kiowa Formation rests disconformably on the Cheyenne Sandstone in their common type area in southwestern Kansas (Fig. 1) (Franks, 1975; Scott, 1970b). Elsewhere on the outcrop, the formation rests on Permian rocks, but in the subsurface of western Kansas, it overlies widespread sandstone that probably is not coextensive with the Cheyenne Sandstone of the type area (Franks, 1975; Scott, 1970b). In north-central Kansas, the Kiowa Formation overlaps progressively older Lower Permian rocks to the northeast. Along the gently dipping western flank of the Nemaha anticline (Fig. 1), the formation, in turn, is truncated and overlapped by the Dakota Formation about 30 mi (48 km) south of the Kansas-Nebraska border (Fig. 2). As the Kiowa Formation pinches out, the northernmost recognizable Kiowa exposures are of the Longford Member.

The Kiowa Formation records the invasion of the Early Cretaceous epicontinental sea into Kansas, and the numerous fossils studied by Twenhofel (1924) and Scott (1970a, 1970b) testify to the generally marine, if somewhat brackish-water, nature of Kiowa sedimentation. The distribution and grain sizes of Kiowa sandstone, the orientation of cross-strata in the sandstone, the distribution of Longford rocks, and the character and distribution of faunal assemblages all indicate that the Kiowa depositional slope generally was inclined to the southwest (S60°W to S70°W, Fig. 1), and that the Kiowa sea transgressed northeastward across Kansas (Franks, 1966, 1975; Scott, 1970a, 1970b). Variations in thickness of the formation generally accord with these data. The unit is thickest in its type area, where it is about 300 ft (91 m) thick (Latta, 1948). In north-central Kansas, maximum thickness is found near the southern end of the outcrop belt (Fig. 1) and approximates 150 ft (46 m). The formation thins irregularly northward from there until the pinchout between the Dakota Formation and Permian rocks is reached near the Kansas-Nebraska line.

Shale traditionally has been considered to be the kind of rock most characteristic of the Kiowa Formation. Kiowa shale generally is medium to dark gray and weathers olive gray to olive brown. Most of the shale is thinly laminated, fissile, and plastic when wet. Imprints of pelecypods, together with sorted assemblages of fish scales, teeth, and bone fragments, as well as glauconite pellets, are found locally along bedding planes in the shale. Illite, mixed-layer illite-smectite, and smectite are the major clay minerals. Kaolinite also is present and it is more abundant upward in the formation, whereas smectite (chiefly Ca-montmorillonite) is more abundant near the base (Franks, 1966). The mineralogy of typical Kiowa shale contrasts sharply with that of clay rocks in the Longford Member.

Laminae and beds of siltstone and very fine grained sandstone are scattered through sections of Kiowa shale, as are beds of coquinoid limestone or mollusk-shell conglomerate. Many of the siltstone and sandstone beds are fossiliferous or are mottled by burrowing. Concretionary masses and layers of impure siderite (clay-ironstone) and calcareous cone-in-cone structure are common, especially in north-central Kansas (Franks, 1969a, 1969b). Most Kiowa shale, together with the enclosed beds of siltstone, sandstone, and coquinoid limestone, is a product of sedimentation in somewhat brackish, open-sea environments (Scott, 1970a, 1970b; Franks, 1975). Although shale is common in the Kiowa Formation, the type section in southern Kansas is unusual in that it contains so much shale and so little sandstone.

Extensive bodies of dominantly fine- to medium-grained sandstone in the Kiowa Formation can be divided into two genetically significant types of deposits (Franks, 1966, 1975): thick, lenticular deposits as much as 100 ft (30 m) thick, and thin, sheetlike deposits less than 10 ft (3 m) thick. Although sandstone bodies of these types occur throughout the Kiowa Formation in north-central Kansas, they are most common in the upper parts of the unit. They contrast with the thinner beds of very fine grained sandstone and siltstone that are scattered through sections of Kiowa shale. Some of the thick, lenticular deposits of sandstone probably formed as barrier bars and related shoreface accumulations. Some also show sedimentary structures and partial sequences of structures that accord with the migration of tidal inlets and associated spit platforms (Kumar and Sanders, 1974). Apparently, linear clastic shorelines (Selley, 1970) dominated Kiowa sedimentation during maximum transgression of the Kiowa sea into north-central Kansas (Franks, 1975).

Fossil assemblages, sedimentary structures, and ripple-laminated gradations into underlying shale indicate that the sheetlike deposits of sand also accumulated in nearshore realms (Franks, 1966, 1975; Scott, 1970a, 1970b). Widespread sheetlike sandstone that marks the top of the Kiowa Formation is, in part, compatible with deposition as delta-front sheet sands of shoal-water deltaic complexes (Franks, 1975). Their accumulation marked the onset of widespread regressive sedimentation that heralded deposition of the overlying Dakota Formation.

**Dakota Formation**—The Dakota Formation marks the climax of the regressive phase of sedimentation in the transgressive-regressive depositional cycle recorded by Cheyenne, Kiowa, and Dakota rocks. The Dakota Formation, which is about 250 ft (76 m) thick, rests both disconformably and conformably on Kiowa rocks. Evidence of disconformable relationships is seen best where lenticular deposits of fluvial sandstone fill scours that were eroded into underlying Kiowa deposits (Franks, 1966, 1975). Additional evidence of disconformable relationships is found along the pinchout of the Kiowa Formation in north-central Kansas (Figs. 1, 2). As the Kiowa-Dakota contact is followed northward, sandstone marking the top of the Kiowa Formation is truncated, then underlying Kiowa shale and sandstone are truncated, and the last easily recognized manifestation of the Kiowa Formation is the siltstone that marks the top of the Longford Member near the base of the Kiowa Formation. Conformable contacts lack evidence of pre-Dakota weathering and erosion of uppermost Kiowa rocks. Locally, sandstone at the top of the Kiowa Formation grades sharply upward into overlying Dakota clay rocks (Bayne and others, 1971, measured section 5, p. 75-76), or beds and lenses of ripple-marked or other sandstone similar to that in the Kiowa Formation are enclosed by basal Dakota clay rocks (Bayne and others, 1971, p. 16; Franks, 1975, p. 491-497).

The Dakota Formation is noted for its abundant lenticular bodies of sandstone, but kaolinitic clay rocks amount to 60 or 70 percent of the aggregate thickness of the unit. Plummer and Romary (1942) divided the Dakota Formation of central Kansas into two members based on the characteristic colors of the clay rocks. The Terra Cotta Clay Member, which comprises approximately the lower two-thirds of the formation, is characterized by light-gray claystone and mudstone that are splotched by abundant, red, hematitic mottles. The Janssen Clay Member, which comprises approximately the upper third of the formation, includes abundant gray and dark-gray clay rocks as well as lignite and other carbonaceous matter. Gray clay rocks and sparse lignitic layers also occur at numerous localities at the base of the Dakota Formation (Plummer and others, 1963). These gray clay rocks and the associated carbonaceous materials appear to be remnants of delta-plain sediments that were deposited in a complex of shoal-water deltas that formed an important part of the early stages of Dakota sedimentation (Franks, 1975). The red-mottled clay rocks that characterize the bulk of the Terra Cotta Clay Member formed primarily as overbank deposits of streams that flowed southwestward across Kansas as alluvial-plain environments became established. The sediments that formed the gray clay rocks and associated carbonaceous beds, siltstone, and sandstone of the Janssen Clay Member were deposited under the influence of the transgressing, Cenomanian Graneros sea (Franks, 1966, 1975; Siemers, 1971, 1976). Except for their highly kaolinitic nature, and for the related physical properties that depend on it, many of the clay rocks in the Dakota Formation are similar to those in the Longford Member of the Kiowa Formation. The similarity locally poses problems of interpretation, especially near the pinchout of the Kiowa Formation.

## PERMIAN-CRETACEOUS UNCONFORMITY

Some of the characteristics of the unconformity that separates Permian and Cretaceous beds in Kansas (Greene, 1910) are important to a study of the Longford Member. The unconformity is part of a regional, transgressive disconformity that also marks the base of the Longford Member (Fig. 2). Reworking of fossil soils and lag deposits, whose remnants are preserved locally along the unconformity in central Kansas, provided some of the sedimentary materials now incorporated in the Cretaceous beds above the unconformity (Fig. 3). Relief developed on the ancient erosion surface also affected the thickness of Longford sediments that accumulated from place to place.

The Permian-Cretaceous unconformity in north-central Kansas is a mature erosion surface. It commonly shows as much as 50 ft (15 m) of local relief over distances as great as a mile (1.6 km). In some areas, local relief approximates 75 to 100 ft (25 to 30 m). Pebbles and granules of chert, silica-cemented sandstone, quartzite, and "vein" or pegmatitic quartz (Table 1) are concentrated in the basal parts of the Cretaceous units above the unconformity in many places. The pebbles, which generally measure appreciably less than 6 cm in long diameter, occur primarily on, or on the flanks of, topographic highs developed on underlying Permian rocks. The enclosing matrix ranges from sandstone to mudstone, and pebbles in sandstone may be several feet above the actual base of the Cretaceous units. The pebbles and granules are thought to be relicts of a thin lag deposit of gravel

TABLE 1.—Composition of pebbles in fine-grained Longford sandstone in NW ¼ sec. 27, T.8S., R.2E., Clay County, Kansas. Percentages based on a count of 180 pebbles.

| *Pebbles* | *Percentages* |
|---|---|
| CHERT | 69 |
| white, porous | 39 |
| white, dense | 6 |
| black to brown, fine-grained | 14 |
| black to brown, coarse | 10 |
| SILICA-CEMENTED SANDSTONE | 1 |
| QUARTZITE | 15 |
| QUARTZ SCHIST | 1 |
| "VEIN" OR PEGMATITIC QUARTZ | 14 |

FIGURE 3.—Permian-Cretaceous contact at base of Longford Member, near cen. SE¼ sec. 23, T.16S., R.1E., Dickinson County (units 2 and 3, measured section 3, Appendix B and Pl. 2). Pocket knife, about 15 cm long, marks contact. Handle rests against brownish-gray Longford mudstone and claystone (dark gray in photograph); blade penetrates dusky-red (very dark gray in photograph) top of variegated soil developed on normally olive Wellington Formation (Permian) mudstone or shale. Note fragments of dusky-red and light-colored regolith incorporated in basal Longford clay rock; one dusky-red fragment (very dark gray in photograph) is at left edge of knife handle.

that originally was scattered widely over the erosion surface. Reworking of the lag deposits continued as topographic lows were formed, and filled with Cretaceous sediments. The pebbles in the Longford Member and elsewhere at the base of the Cretaceous section in north-central Kansas were derived mainly from Paleozoic sedimentary rocks in the continental interior, but the pebbles of metamorphic or igneous material ultimately may have been derived from the crystalline terrains of the central Appalachians, if they came from the same sources as some of the heavy minerals in Kiowa and Dakota sandstone (Franks, 1966, 1975).

The relict soil profiles preserved at many places along the unconformity generally are less than 4 ft (1.2 m) thick. The upper few feet of normally reddish-brown or olive-gray Permian rocks were extensively bleached or variegated prior to deposition of the overlying Cretaceous sediments (Fig. 3; measured sections 1 and 3, Appendix B and Pl. 2). Illitic and chloritic or vermiculitic Permian siltstone, mudstone, and shale (Swineford, 1955) were converted to clay-mineral assemblages rich in smectite, kaolinite, and, locally, halloysite (Franks, 1966). The fossil soils contain only trace amounts of chloritic or vermiculitic components. The soils commonly show a progressive upward change from smectite-rich assemblages to kaolinite-rich assemblages. The weathering sequence is similar to that described by Altschuler and others (1963) and Altschuler and Dwornik (1964) in which smectite is generated by weathering of mica- and chlorite-like parent minerals, and kaolinite in turn is formed by weathering of smectite.

Keller (1964) observed that smectite (montmorillonite) is generated under conditions of weathering that differ from those under which kaolinite and related minerals form, although both processes may go on side-by-side, or in different parts of the same soil profile. Formation of smectite requires fairly abundant cations in solution in the soil, whereas removal of both silica and cations is essential to the production of kaolinite. Provided that pH in the soil is low, drainage need not be good for generation of kaolinite. Formation of halloysite nodules at least locally in the soil profiles developed on Permian rocks in central Kansas (measured section 3, Appendix B; Franks, 1966) also indicates that the soil was not well drained, and that an acid pH prevailed. The abundance of kaolinite in the upper parts of the soils and localized occurrence of halloysite are highly suggestive of weathering under humid, subtropical conditions (Millot, 1970; Mohr and others, 1972; Parham, 1969, 1970; among others).

Other features also are compatible with development of the fossil soils under conditions of low relief, poor drainage, and humid tropical to subtropical climates. Like the pebbles along the unconformity, the fossil soils are preserved mainly on topographic highs developed on the Permian rocks prior to deposition of the overlying Cretaceous beds, whereas least-altered Permian rocks are found primarily in topographic depressions along the unconformity (cf. measured section 1, Appendix B). The unconformity apparently underwent a period of renewed erosion prior to deposition of the Cretaceous beds, and the unconformity must have shown even less relief at the time the soils were generated than it does now. The general thinness of the soil profiles may stem from repeated erosion during their development, but the thinness and the marked upward changes in mineralogy accord with formation of the soils on relatively impermeable, clayey rocks along a poorly drained erosion surface of low relief (Birkeland, 1974, p. 99; Ollier, 1969, p. 91-92). Similarly, the bleaching and variegation that took place during development of the soils is suggestive of gleying or the dissolution and redistribution of iron oxides under conditions of impeded drainage and fluctuating water content (Birkeland, 1974, p. 118).

When the weathering that generated the fossil soils took place is problematic. The weathering predates the Cretaceous beds above the unconformity. It also predates the inferred episode of erosion that preceded deposition of the Cretaceous sediments. The erosion surface that separates Permian and Cretaceous strata in north-central Kansas, moreover, is not only a transgressive disconformity, but also a composite unconformity (Weller, 1960, p. 396). The single erosion surface in central Kansas is represented by a number of unconformities in the subsurface of western and southwestern Kansas, and in nearby parts of Colorado, New Mexico, and Oklahoma (Baldwin and Muehlberger, 1959, p. 31-51; Fader and others, 1964, p. 15; McLaughlin, 1954, p. 82-96; Merriam, 1963, p. 71-77, Fig. 33; Schoff and Stovall, 1943, p. 43-71; Voegeli and Hershey, 1965, p. 52-58; Zeller, 1968, p. 53-54). There, unconformities separate Permian from Triassic strata, Triassic from Jurassic strata, and Jurassic from Cretaceous strata; other unconformities occur within the Triassic and Jurassic sequences. Casual field observations and a review of the literature suggest that no well-developed fossil soils are present in or beneath pre-Cretaceous Mesozoic beds where they occur above Permian rocks, either in Kansas or in nearby parts of Colorado, New Mexico, or Oklahoma. If true, it may mean that the fossil soils along the Permian-Cretaceous unconformity in Kansas developed primarily during pre-Kiowa or pre-Cheyenne Early Cretaceous time. Nonetheless, the mineralogy of the basal Cretaceous clay rocks seemingly reflects the mineralogy of the fossil soils, which probably were widespread throughout the continental interior (cf. Andrews, 1958, and Parham, 1970). As is true for the soil profiles, many of the Cretaceous clay rocks are composed primarily of smectite and kaolinite, and they contain only trace amounts of chloritic or vermiculitic clays (Franks, 1966). These relationships also apply to the Longford Member of the Kiowa Formation.

# DESCRIPTION OF LONGFORD MEMBER, KIOWA FORMATION

## NAME, TYPE SECTION, DISTRIBUTION, AND THICKNESS

The Longford Member of the Kiowa Formation crops out in two separated areas in north-central Kansas (Fig. 1). The northern area spans parts of Clay, Ottawa, Saline, and northwestern Dickinson counties. The southern area includes parts of western Marion County and small parts of eastern McPherson and southwestern Dickinson counties. Longford rocks were not found in large parts of Saline, McPherson, and Rice counties. The member is named for exposures in the northern area of outcrops, those north of the town of Longford, southwestern Clay County (Pl. 1). The type section (measured section 1, Appendix B and Pls. 1, 2) was described in drainage cutbanks in W ½ SW ¼ sec. 9, T.10S., R.1E., and in roadcuts along the road linking Longford to Oak Hill on the west line SW ¼ sec. 9. The contact with overlying Kiowa rocks is not exposed in the immediate vicinity of Longford, but it is well exposed in sec. 32, T.9S., R.1E., Clay County (measured section 2, Appendix B and Pls. 1, 2). A reference section was described near the cen. SE ¼ sec. 23, T.16S., R.1E., Dickinson County, in the southern area of outcrops where other aspects of the member are expressed (measured section 3, Appendix B, Fig. 1 and Pl. 2). Excellent exposures of the member can be seen in sec. 27, T.9S., R.1E., near the NE corner sec. 1, T.6S., R.1E., and near the center sec. 1, T.6S., R.2E., Clay County (Pl. 1).

The Longford Member ranges in thickness from 0 to about 100 ft (0 to 30 m). Maximum thickness is near Longford and Oak Hill, southwestern Clay County (Pl. 1). The member grades westward into, and intertongues with, more characteristic Kiowa rocks. Longford rocks are facies equivalents of rocks in the upper parts of the *Inoceramus comancheanus* and the lower parts of the *I. bellvuensis* concurrent range zones of Scott (1970b) (Fig. 2). The member thins northward and could not be distinguished in any practical way from basal Dakota rocks north of secs. 33 and 34, T.5S., R.3E., Washington County, along the pinchout of the Kiowa Formation (Fig. 2; Pl. 1). The northward thinning and disappearance of the Kiowa Formation and its Longford Member beneath the Dakota Formation in Clay and Washington counties shows some coincidence with the northward increase in breadth of the gently dipping western limb of the Nemaha anticline (Fig. 1) in north-central Kansas (cf. Merriam, 1963, Fig. 100). In outcrop, the siltstone marking the top of the Longford Member is the northernmost kind of Kiowa rock that can be identified with certainty beneath the Dakota Formation in north-central Kansas.

The distribution of Longford rocks in the subsurface west of the outcrop areas is incompletely known. The member extends beneath large parts of Ottawa County and apparently reaches into the subsurface of northwestern Ellsworth County (Fig. 1) (O. S. Fent, 1977, written communication). The stratigraphic relations between Longford rocks and the sequences of nearly white to greenish-gray siltstone, shale, and sandstone in the subsurface north of the type area of the Cheyenne and Kiowa formations and in western Rice County (Fent, 1950; Latta, 1950; McLaughlin, 1949) are problematic, inasmuch as neither Longford rocks nor those rocks are present at the base of the Kiowa Formation on the outcrop in eastern Rice County or in most of McPherson County (Fig. 1). Despite the pinchout of the Kiowa Formation along the Nemaha anticline in north-central Kansas, Longford rocks, or their analogues, may extend into the subsurface of southern and central Nebraska, as is suggested by the depositional strike inferred for the Kiowa Formation (about N25°W) (Fig. 1) and by the distribution of rock types within the member. As is true in outcrop, tracing the Longford Member in the subsurface will depend largely on being able to recognize the siltstone that marks the top of the member.

## SILTSTONE AT TOP OF LONGFORD MEMBER

The siltstone that marks the top of the Longford Member (Pl. 2; Figs. 4, 5) is a distinctive unit that allows ready differentiation between Longford rocks and the shale and sandstone in higher and laterally equivalent parts of the Kiowa Formation. The siltstone also is a useful stratigraphic reference where the Dakota Formation rests on the Longford Member. The capping siltstone generally is between 3 and 15 ft (0.9 and 4.6 m) thick. Characteristically it is nearly white, but commonly it is light gray or stained light brownish gray, light yellowish gray, or grayish orange by "limonite." In most places, it is well indurated and weathers to thin beds that in turn are composed of thin laminae. The laminae are uniform and horizontal, or wavy, or rippled. Ripple marks include diversely oriented, symmetric and asymmetric transverse forms having amplitudes of 1 to 3 cm and wave-

FIGURE 4.—Thin-bedded to thin-laminated siltstone marking top of Longford Member near type Longford, about 0.15 mi (0.24 km) north of SE corner sec. 8, T.10S., R.1E., Clay County. Roadcut, not actually part of measured section 1 (Appendix B and Pl. 2), almost spans the interval from unit 17 through unit 19 in the measured section. Note dark, discoidal masses of calcite-cemented siltstone at top of exposure. Handle of pick about 2.5 ft (0.8 m) long.

lengths between 3 and 15 cm. Locally, interference patterns are seen. Also present are diversely oriented linguoid ripples as much as 15 cm across. In places, small-scale cross-laminae (including the micro-cross-lamination of Hamblin, 1961, or rib-and-furrow structure of Stokes, 1953) are common. Locally, flaser bedding can be detected in the siltstone. In a few places, even or wavy and ripple-marked laminae form sets of gently inclined cross-strata that depart less than 10 degrees from the horizontal (Fig. 5). Locally, the sets


FIGURE 5.—Thin-bedded to thin-laminated siltstone marking top of Longford Member, SE ¼ sec. 23, T.16S., R.1E., Dickinson County. Siltstone grades downward into dark-gray, plastic claystone in middle ground. Note low-angle cross-strata most evident above and to left of top of pick. Handle of pick about 2.5 ft (0.8 m) long. Numbers refer to units in measured section 3 (Appendix B and Pl. 2).

of cross-strata truncate underlying horizontal stratification.

Well-sorted, subangular to subrounded quartz grains, chiefly medium to coarse silt, approximate 95 percent of the capping siltstone. The rock commonly contains disseminated, rounded to subangular quartz grains that measure as much as 0.2 mm in long diameter and show nearly straight extinction. Generally, the sand-sized grains amount to less than one percent of the rock, but, locally, where they are concentrated, they impart a striking bimodal grain-size distribution to parts of the rock (Fig. 6). One percent or less heavy minerals, chiefly zircon and tourmaline, are present. Sparse nodules of marcasite or pyrite locally enclose scattered fragments of carbonaceous matter. Except for trace amounts of detrital muscovite, interstitial clay makes up most of the remainder of the rock. Where the rock is friable, it is nearly devoid of interstitial clay. Calcite cement is distributed erratically through indurated parts of the siltstone. The cement locally follows bedding, but it also forms discoidal to spherical concretionary masses (Fig. 4). The indurated siltstone tends to form ledges and benches that stand out in relief along gentle slopes developed on overlying Kiowa shale and on underlying Longford clay rocks.

The drilling characteristics of the siltstone at the top of the Longford Member are distinctive and should be helpful in tracing the subsurface distribution of the member even though the rock yields few cuttings when penetrated by a roller bit (O. S. Fent, 1977, written communication). Rather, the rock responds much like a friable, fine-grained sandstone, and the samples contain only sparse chips of indurated, white to light-green siltstone. Decanting the mud fails to reveal sand grains.

Plant material and molds and casts of burrows or trails were the only fossils seen in the siltstone. The plant material occurs chiefly as macerated stem or leaf debris scattered along bedding surfaces, either as carbonized fragments or as imprints. In southern Ottawa County and nearby parts of Saline County (Fig. 1), however, the lower and middle parts of the siltstone locally contain tubular molds and casts resembling features that Hattin (1965) and Siemers (1971) identified as fossil reeds near the top of the Dakota Formation in central Kansas. C. T. Siemers (1976, written communication), however, suggested that some of the Longford structures may be insect burrows, whereas others indeed are reed fossils. The structures in the Longford Member are as much as 38 cm long and 3 mm in diameter. Their orientation is upright or departs less than 15 degrees from the normal to bedding. The fossils occur as molds on weathered surfaces, and as silt- and clay-filled casts. In places, carbonaceous matter outlines or is entrapped in the casts. Some of the casts contain iron sulfide or "limonite," and some are surrounded by iron-oxide oxidation haloes that are as much as 6 cm in diameter. Bedding is not disturbed by the structures, nor is it apparent in them.

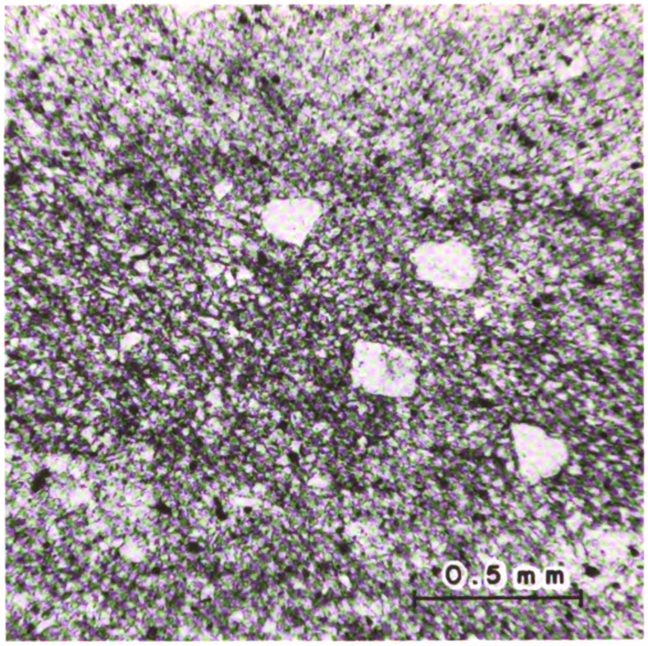



FIGURE 6.—Photomicrograph of siltstone in upper part of Longford Member, center W line SW ¼ sec. 9, T.10S., R.1E., Clay County (unit 13, measured section 1, Appendix B and Pl. 2). Carbonaceous lamina (dark gray) contains subrounded quartz grains as much as 0.2 mm in long diameter. The rock also contains appreciable interstitial clay. Plane-polarized light.

Molds and casts of branching stems or rootlets are preserved in the capping siltstone in parts of Dickinson County where the base of the siltstone is within a few feet of lignite in the lower parts of the member. The walls of the structures are smooth, relatively straight to arcuate, and branching. The structures are 2 mm or less in diameter, but they are as much as 7 cm long. The molds and their silty casts tend to follow bedding, but they also penetrate it at steep angles. Trace fossils that may be insect trails (Siemers, 1976, written communication) commonly are associated with the fossil stems or rootlets. They are smooth-walled to beaded, gently arcuate bedding-plane casts that are 1 to 2 mm in diameter and as much as 5 cm long.

Other trace fossils in the capping siltstone include burrows of uncertain affinities. One is a *Planolites*-like structure (Häntzschel, 1962; Frey and Howard, 1970) that consists of circular molds and casts that both transect and tend to follow the wavy bedding of the rock. The molds and their siltstone casts are about 1 cm in diameter and as much as 10 cm long. A second type of burrow is tubular and subvertical. Some are rod-shaped; others are Y-shaped. They measure from 0.2 to 1.5 cm in diameter and are as much as 20 cm long, although most are less than 15 cm long. Most of the structures are smooth walled and resemble the various rod-shaped and tubular burrows described by Frey and Howard (1970) and Siemers (1971). Some, however, have a segmented appearance similar to that of *Arthrophycus* (Frey and Howard, 1970), but the segmentation may be an artifact of bedding.

An upward progression of trace fossils and sedimentary structures can be deciphered locally in southwestern Dickinson County (Fig. 1). Wavy laminated strata containing molds and casts of branching stems or rootlets, as well as insect trails, occur low in the capping siltstone in places. Elsewhere, the lower parts of the unit show only even, horizontal laminae along which carbonaceous plant debris is scattered. In the middle and upper parts of the siltstone, rod- and Y-shaped burrows occur widely and are associated with current and oscillation ripple marks.

The contact between the siltstone and overlying Kiowa shale is gradational (measured sections 2 and 3, Appendix B and Pl. 2). The basal few centimeters or tens of centimeters of shale above the Longford Member commonly are light to dark gray, smectitic and kaolinitic, and silty. Discontinuous laminae of clay-rich siltstone decrease in abundance upward in the interval, and the silty interval grades upward to more typical, plastic, fissile, olive-gray Kiowa shale containing discoidal concretions of impure siderite (clay ironstone) that are strung out along bedding planes. In many places, the siltstone that marks the top of the Longford Member is separated from typical Kiowa shale by as much as 6 ft (2 m) of plastic, light-gray shale having poor fissility. The shale is composed of variable proportions of smectite, kaolinite, and illitic clay. The clay-mineral assemblage is intermediate between that of typical Kiowa shale and that of Longford rocks. The shale commonly is silty and much of it contains numerous nodules of marcasite or pyrite as well as variable amounts of carbonaceous matter. Weathered slopes developed on intervals of pyritic shale yield gypsum "sunbursts" and show abundant "limonite" stain. In some parts of southeastern Ottawa County and northern Saline County, the capping siltstone is absent and Kiowa shale rests directly on clay rocks belonging to the lower parts of the member.

Thick, lenticular deposits of fine- to medium-grained Kiowa sandstone, rather than typical, olive-gray Kiowa shale, also rest on the light-gray silty shale described above. In a few places in northwestern Dickinson County and western Clay County (e.g., due south and due west of Longford), thick, lenticular deposits of Kiowa sandstone may rest directly on the siltstone at the top of the Longford Member, but poor exposures of the lower parts of the sandstone lenses prevent direct observation of the contacts.

Where the upper parts of the Kiowa Formation are truncated by the Dakota Formation in Clay and Washington counties (Pl. 1), Dakota rocks rest unconformably on the Longford Member (Fig. 2). Nonetheless, the capping siltstone still allows recognition of the top of the Longford Member in many places, even though the lowermost parts of the Dakota Formation contain appreciable silt, much of it apparently reworked from Longford rocks. However, where pre-Dakota erosion removed the siltstone that marks the top of the member, separation of Longford rocks from the overlying Dakota Formation generally is impractical owing to the similarity of the clay rocks in the two units.

The boundary between the siltstone at the top of Longford Member and the argillaceous rocks that form the bulk of the underlying parts of the member is gradational in most places. The siltstone becomes enriched in clay near the boundary; the clay rocks become laminated in their upper parts; and laminae and beds of siltstone become more abundant upward. The transition commonly takes place in a stratigraphic interval that is 3 to 6 ft (1 to 2 m) thick (measured section 3, Appendix B and Pl. 2). Elsewhere, sequences of poorly indurated siltstone, or siltstone showing little sign of bedding, intervene between the clay rocks and the capping siltstone (measured section 1, Appendix B and Pl. 2).

## LOWER PART OF LONGFORD MEMBER

The assemblage of Longford rocks beneath the capping siltstone is highly varied, both in lithology and thickness. Local variations in thickness are controlled chiefly by relief along the Permian-Cretaceous unconformity, but regional variations in thickness also occur. Greatest thickness, as much as 80 ft (24 m), is in western Clay County in the vicinity of Longford and Oak Hill (Plate 1). The rocks thin southward and westward from there, and, in parts of Marion County, the capping siltstone appears to rest directly on weathered Permian strata.

Longford rocks beneath the capping siltstone mostly are clay rocks and siltstone, but sparse layers of lignite and beds and lenses of sandstone also are present. Light-gray clay rocks blotched by abundant red, hematitic mottles are found chiefly in the northern parts of the Longford outcrop belt (Clay and Ottawa counties, Fig. 1), whereas gray clay rocks and carbonaceous beds are most abundant in the southern parts of the outcrop belt (Dickinson and Marion counties, Fig. 1). The scattered exposures of the lower parts of the member in eastern Ottawa County and north-central Saline County, southwestward from the type area, also suggest an analogous westward decrease in the abundance of red-mottled clay rocks, and an increase in the abundance of gray clay rocks and carbonaceous materials. Most of the sandstone exposed in the lower parts of the member is in the northern parts of the outcrop belt, chiefly in Clay County.

**Clay rocks and carbonaceous beds**—The clay rocks in the lower part of the Longford Member range from dark-gray and brownish-black, plastic claystone and thin-laminated, brownish-gray, carbonaceous shale to light-gray, light brownish-gray, and red-mottled mudstone (Figs. 5, 7; measured sections 1 and 3, Appendix B). Some of the mudstone grades imperceptibly into siltstone. In places, where the lower parts of the member are relatively silty in Saline, Ottawa, and Clay counties (Fig. 1), the rocks enclose lenses or tongues of fissile gray to olive-gray shale that locally contains concretions of impure siderite (clay ironstone) and is typical of other parts of the Kiowa Formation. Some of the dark clay rocks and brownish-gray carbonaceous shale enclose or grade laterally into beds of lignite, especially in southwestern Dickinson County and northwestern Marion County (Fig. 1).

The clay rocks form discontinuous beds and lenses generally less than 10 ft (3 m) thick. The writer suspects that no single bed or lense can be traced laterally for more than a few hundred feet (about 100 m). The beds and lenses mostly show gradational or sharply gradational contacts with overlying and underlying units (measured sections 1 and 3, Appendix B and Pl. 2; Figs. 5, 7). Some of the dark-gray and brownish-gray clay rocks that contain abundant carbonaceous matter are thinly laminated and fissile, but bedding within most of the clay rocks is poorly developed and indistinct. In some clay rocks, indistinct bedding is evident on thickness scales of one or more feet (0.3 m or more) because of variations in the amount of silt that the rocks contain. As the proportion of silt in some of the massive mudstones increases, however, the rocks may show indistinct lamination and thin bedding similar to that developed in the capping siltstone. In some red-mottled clay rocks, indistinct layering is imparted by vertical variations in the size and abundance of the red mottles.

Abundant small-scale, randomly oriented, slickensided fractures characterize much of the massive claystone and mudstone. On weathering, the massive clay rocks break into angular and conchoidal chips measuring about 1 cm in long dimension. As weathering continues, however, clay rocks containing appreciable smectite develop puffy, nonresistant slopes much broken by shrinkage cracks (Fig. 7). The puffy slopes developed on Longford clay rocks set them apart from the more resistant slopes formed on kaolinitic, but otherwise similar, rocks in the Dakota Formation.

X-ray diffraction studies show that Longford clay rocks contain mainly kaolinite and smectite in the fraction finer than 2 $\mu$ (Fig. 8). Illite tends to be a minor component, except in the lenses or tongues of more typical Kiowa shale enclosed by the member. Vermiculite or chlorite occurs widely in Longford clay rocks, but generally in amounts that barely exceed the limits of detection by diffraction methods, primarily as mixed-layer structures with smectite (Fig. 8). No consistent relation was found between the clay minerals in the rocks and their color: gray and dark-gray clay rocks can be as kaolinitic or as smectitic as red-mottled or light-gray rocks, although red-mottled clay rocks tend to be more kaolinitic than the others (Pl. 2). Clay rocks rich in smectite tend to occur near the base of the member (measured section 1, Pl. 2), but exceptions are manifold (measured section 3, Pl. 2). Variations in the proportions of kaolinite and smectite may depend partly on the proportions of silt in the clay rocks, or on the siltiness of the enclosing rocks (units 10 and 15, measured section 1, Pl. 2). Detectable chloritic or vermiculitic components, however, are most common in the lower parts of the member. In contrast, chloritic or vermiculitic clay minerals were not detected either in overlying and facies-equivalent Kiowa rocks or in the bulk of the Dakota Formation in north-central Kansas (Franks, 1966).

A kaolinite-smectite mixed-layer clay was found in Longford rocks just above the Permian-Cretaceous unconformity in southwestern Dickinson County. X-ray diffraction traces of the mixed-layer clay are shown in Figure 9. The clay is of particular interest because of the apparent scarcity of interstratified kaolinite-smectite complexes in the rock record (Schultz and others, 1971; Wiewiora, 1971). The mixed-layer clay occurs near the base of the brownish-gray and red-mottled clay rocks of unit 3, measured section 3 (Appendix B and Pl. 2), where numerous fragments and pods of claystone reworked from underlying Permian rocks also are present (Fig. 3). The upper parts of unit 3 contain abundant smectite and lesser amounts of kaolinite, but they lack major amounts of the kaolinite-smectite mixed-layer clay.

The red-mottled clay rocks in the lower parts of the Longford Member offer a striking contrast to the generally drab or gray shale of the Kiowa Formation. Where the red-mottled clay rocks contain abundant kaolinite and little smectite, however, they closely resemble red-mottled clay rocks in the Dakota Formation. The red mottles in Longford clay rocks are

**Figure 7.—Red-mottled mudstone near base of Longford Member, near cen. W line SW ¼ sec. 9, T.10S., R.1E., Clay County (unit 5, measured section 1, Appendix B and Plate 2). Hematite in the mottles (medium gray near head of pick) has been redistributed by weathering. Puffy and cracked nature of slope indicates the smectitic nature of the rock. Handle of pick about 2.5 ft (0.8 m) long.**

produced by irregularly shaped concentrations of red, purplish-red, and reddish-brown hematite that stand out against the pale-gray or pale greenish-gray tones of the remainder of the rock (Fig. 7). Some of the hematite is present as stain that permeates the clay matrix of the rocks to form isolated, irregularly shaped patches measuring 1 or 2 cm across; some hematite has a reticulate pattern and fills minute, randomly oriented cracks; and some coats the surfaces of diversely oriented, small-scale, slickensided fractures. The patches, fracture fillings, and fracture coatings coalesce to produce irregularly shaped mottles as much as 10 cm long. In places, the mottles in turn coalesce to produce irregularly shaped vertical streaks

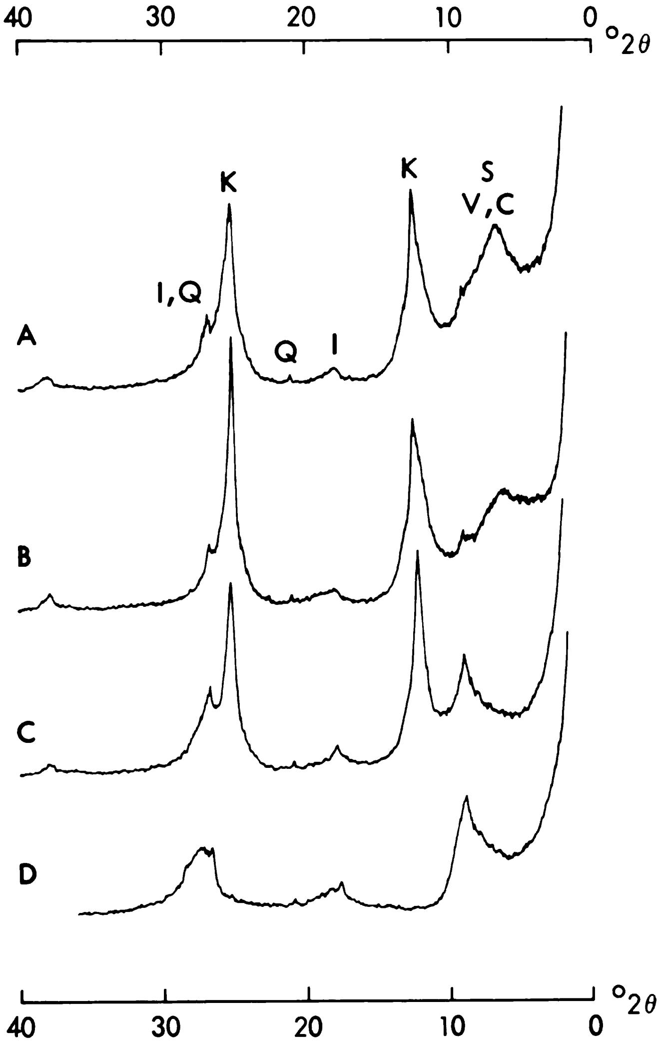



FIGURE 8.—Diffractometer patterns of fraction finer than 2 μ from mudstone at base of Longford Member, Kiowa Formation, SE ¼ SW ¼ SW ¼ sec. 9, T.10S., R.1E., Clay County (near measured section 1, Appendix B and Pl. 2). Scale-factor setting 8, time-constant setting 4, nickel-filtered copper radiation. A) air-dried; B) glycerated; C) heated to 450°C; D) heated at 575°C for one-half hour. Pattern A shows abundant kaolinite (K), lesser amounts of illite (I), and small amounts of quartz (Q). Smectite or mixed-layer peak near 6.4°2Θ (13.8A) (S, V, C) shows sign of a possible "superlattice" diffraction near 2.8°2Θ (31 or 32A). In pattern B, smectite or mixed-layer diffraction shows only partial expansion toward 18A and is centered near 5.2°2Θ (17.4A). Pattern C shows incomplete collapse of 13.8-A peak toward 10A and consequent skewing of illite 001 maximum past 6.5°2Θ. Pattern D shows little change in the skewed 10-A diffraction. The presence of small amounts of vermiculite or chlorite interlayered with smectite can be inferred.

as much as 2 ft (0.6 m) long. Red mottles also occur in dark-gray and brownish-gray clay rocks, but the mottles tend to be less prominent and less abundant than they are in the light-colored clay rocks. Holocene weathering has redistributed the hematite in the mottles so that the surfaces of some exposures are stained and streaked grayish-red and red, as are joints and shrinkage cracks.

The pale-gray and light greenish-gray interareas between the red mottles in Longford clay rocks are composed of the same clay minerals that are present in the mottles. In some exposures, the interareas also contain trace amounts of disseminated flecks of carbonaceous matter that was not detected in the mottles. Blebs of marcasite or pyrite localized by the carbonaceous matter were seen in some exposures. Spherules of siderite, generally measuring less than 1 mm in diameter, are common in both the red mottles and the interareas between them in Dakota clay rocks (Plummer and Romary, 1942; Franks, 1966), but siderite spherules were noted in only one exposure of red-mottled Longford mudstone (NE ¼ NW ¼ sec. 17, T.6S., R.2E., Clay County).

Thin sections reveal that the red-mottled clay rocks contain variable amounts of silt-sized quartz grains; sparse, partly altered, silt-sized grains of feldspar (less than 2 percent of the silt fraction); scarce, sand-sized quartz grains; and comminuted mica flakes embedded in a clay matrix. The red-mottled mudstone pictured in Figure 10 shows domain and plexoidal clay fabrics that are characteristic of massive clay rocks, but it lacks the relatively uniform orientation of clay flakes and aggregates parallel to bedding that is shown by many shaly rocks. The domain and plexoidal fabrics are produced by variously oriented, birefringent streaks and patches that are composed of aggregated clay and comminuted mica flakes having similar optical orientation. The birefringent streaks and patches measure as much as 0.3 mm long. They enclose areas measuring as much as 0.2 mm across that are composed of smaller, irregularly oriented clay domains that impart a flecked appearance to the thin section (Fig. 10-B, -C). Elongate domains of optically continuous clay locally form linear to arcuate, birefringent streaks as much as 3 mm long and 0.05 mm wide. Some of these streaks have concentrations of hematite stain extending along their central parts (Fig. 10-C). They appear to be oriented aggregates of clay that formed along slickensided fractures seen in hand samples of the rock. In addition, domains of oriented clay partly envelop and form "clay skins" around some of the detrital quartz grains (Fig. 11).

The clay-rock textures noted above are similar to those produced by flocculation of clays (Meade, 1964), but they also have much in common with fabrics developed in soils (Brewer, 1964). Table 2 is a reconsideration of the fabrics in light of the terminology developed by Brewer (1964). Table 2 is offered, not

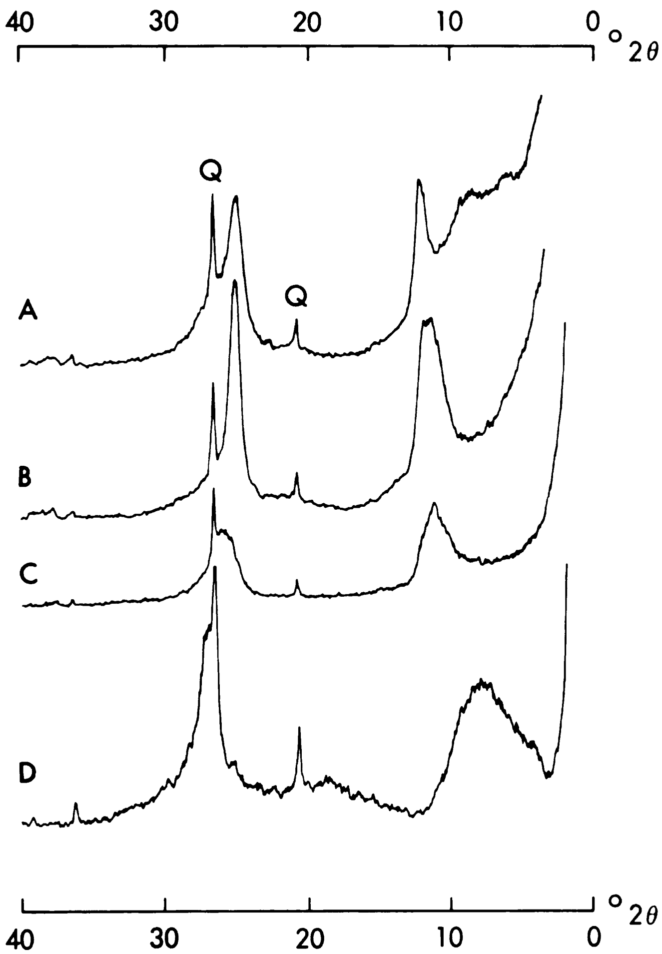



FIGURE 9.—Diffractometer traces (nickel-filtered copper radiation) of fraction finer than 2 μ from mixed-layer kaolinite-smectite clay, base of Longford Member, Kiowa Formation, SE ¼ sec. 23, T.16S., R.1E., Dickinson County (unit 3, measured section 3, Appendix B and Pl. 2). A) air-dried, scale-factor setting 4, time-constant setting 4; B) glycerated, scale-factor setting 4, time-constant setting 4; C) heated to 450°C, scale-factor setting 8, time-constant setting 4; D) heated at 575°C for one-half hour, scale-factor setting 2, time-constant setting 8. Pattern A shows prominent quartz peaks (Q) and prominent, broad diffractions at 12.2° and 25.1°2θ (7.25 and 3.54A). Diffraction at 12.2°2θ is sharply skewed to 11.3°2θ (7.8A) and mixed-layer maxima are centered near 6° and 8.5°2θ (14.7 and 10.4A). Patterns run using ¼° slits (not figured) reveal a "superlattice" diffraction centered near 3.6°2θ (24 to 25A). In pattern B, an inflection point may show near 5.2°2θ; peak at 25.1°2θ is sharpened; peak at 12.2°2θ is broadened, has a distinct shoulder at 12.0°2θ, has a maximum at 11.2°2θ (7.4 to 7.9A), and is sharply skewed to 9.6°2θ. Patterns run using ¼° slits (not figured) show broad "superlattice" maxima centered near 1.7° to 1.8°2θ (about 50A) and near 2.5°2θ (25A). They also verified the presence of an inflection point near 5.2°2θ (17A). The strength of the peaks at 12.2° and 25.1°2θ in pattern A, combined with the loss of intensity and low-angle shift of the major diffractions in patterns C and D, indicate that kaolinite is a major component in the clay. Kaolinite may also be present as a discrete mineral species. Low-angle shift of the "superlattice" maxima on glyceration indicates the presence of an expansible component, probably smectite, as does the inflection point at 17A. Shifting of the peaks on heating can be explained both by destruction of kaolinite and by collapse of the expansible component to a 10-A spacing.

TABLE 2.—Clay fabrics in red-mottled Longford mudstone, SW ¼ NW ¼ sec. 4, T.12S., R.1W., Ottawa County, and soil-matrix ("s-matrix") and void-fabric terminology of Brewer (1964).

| MICROFABRICS | BREWER'S TERMINOLOGY[1] |
|---|---|
| Irregularly oriented, small-scale, birefringent aggregates of clay that impart a flecked appearance to thin sections at high magnifications (80 to 400X). Fig. 10-B, 10-C. | Asepic fabric, p. 309 |
| Birefringent aggregates of clay showing preferred orientation, commonly as streaky or linear domains, or sets of domains, that delimit or enclose other regions of smaller, irregularly oriented clay aggregates (asepic regions). Fig. 10-A. | Masepic fabric, p. 313. |
| Birefringent aggregates of clay as above, but as two or more sets of linear or streak-like domains intersecting at various angles, and enclosing regions of smaller, irregularly oriented aggregates (asepic regions). Fig. 10-B. | Bimasepic (trimasepic, etc.) fabric, p. 313. |
| Optically oriented clay aggregates along the margins of silt or sand grains, but forming part of linear or streaklike domains that extend through clay-rich parts of the rock. Fig. 10-A, 10-B. | Skelsepic fabric (neostrians), p. 104, 294. |
| Linear, optically oriented aggregates of clay along variously oriented fractures. | Skew-plane cutans or skew-plane argillans, p. 210. |
| As above, but marked by accumulations of iron oxide. Fig. 10-C. | Skew-plane ferriargillan, p. 210. |
| Optically oriented, thin aggregates of clay enveloping grains of silt or sand ("clay skins"). Fig. 11. | Grain cutans or grain argillans, p. 209. |

[1] Soil materials finer than 30 microns are referred to as "plasma." Aggregates of oriented particles are termed "plasma separations." Hence, the use of the suffixes "sepic" and "masepic" in many of Brewer's coined terms.

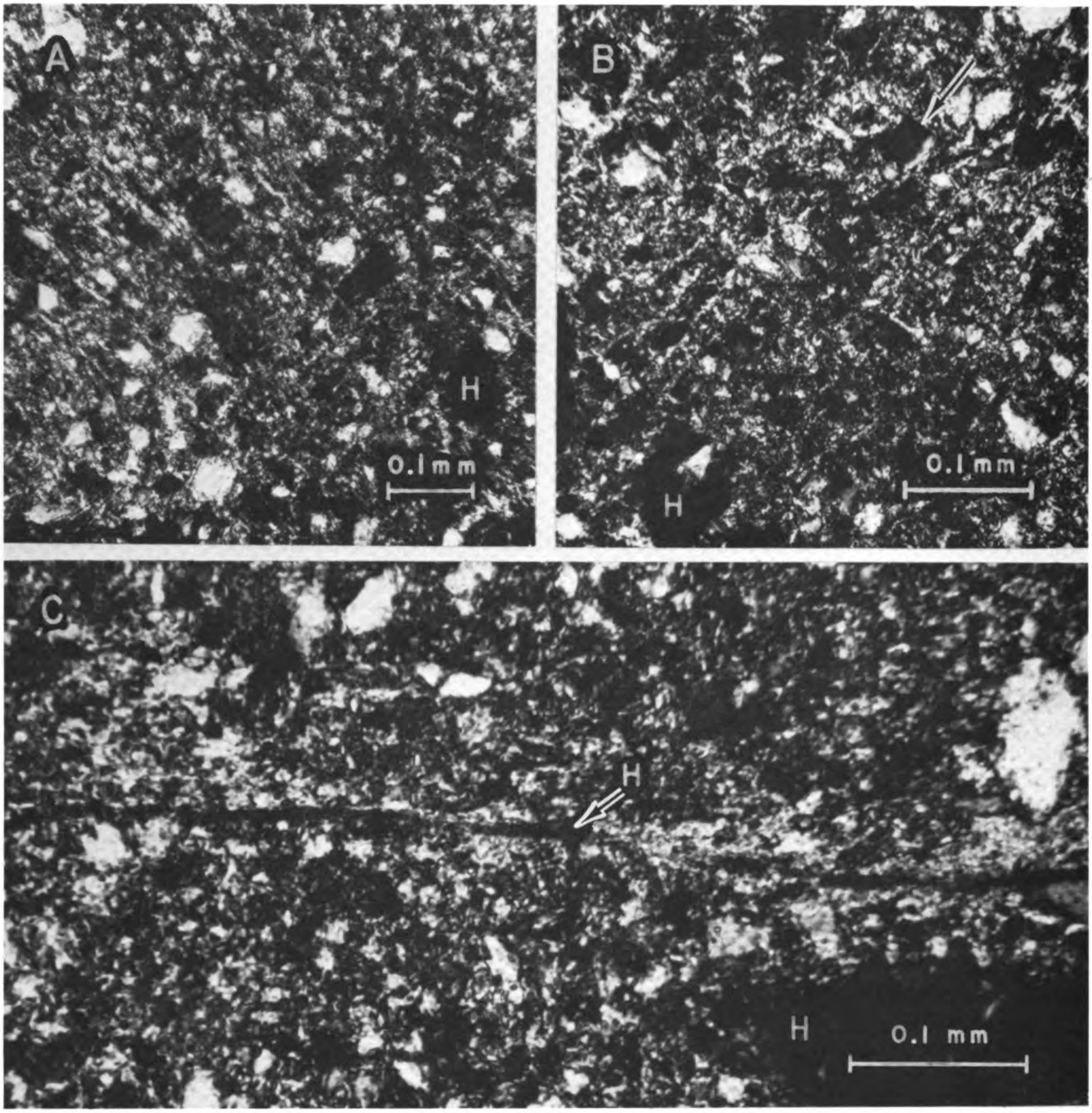



FIGURE 10.—Photomicrographs (crossed polarizers) of thin section of red-mottled Longford mudstone, SW¼ NW¼ sec. 4, T.12S., R.1W., Ottawa County. X-ray diffraction studies show that the clay fraction is composed largely of kaolinite, but contains appreciable illite, sparse smectite, and detectable chlorite or vermiculite. Some irregularly shaped black areas are concentrations of hematite (H) sieved by silt grains and comminuted mica flakes; other black areas are silt particles at extinction. Angular to subrounded silt-sized quartz (mainly light gray and white) abounds. A) Domain fabric imparted by streaklike aggregates (light gray to white) of optically oriented clay and comminuted mica extends from lower right to upper left; a second, less pronounced domain of preferred orientation is at right angles to that trend. Streaks of oriented clay following both trends locally coincide with margins of coarse silt grains. B) Plexoidal fabric imparted by birefringent streaks of oriented clay trending from lower right to upper left and from lower left to upper right, nearly at right angles. Except for silt particles, areas between intersecting streaks are composed largely of irregularly oriented, smaller clay aggregates that impart speckled appearance to photograph. Arrow indicates silt grain enlarged in Figure 11. C) Linear, oriented clay aggregate bounding fracture extends from left to right. Film of hematite (H with arrow, black along center line of streak) marks the fracture. Note flecked appearance of photograph imparted by irregularly oriented, minute domains of clay away from fracture-controlled birefringent streak.

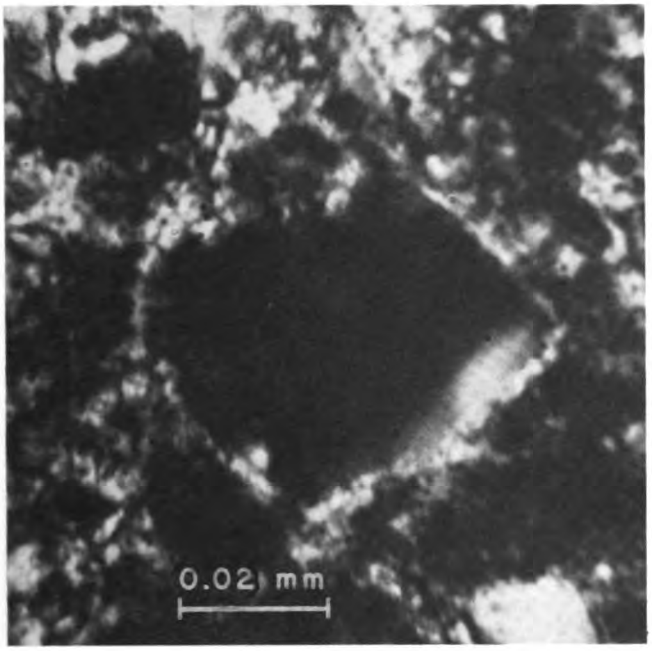



FIGURE 11.—Enlargement (crossed polarizers) of silt grain indicated by arrow in Figure 10-B. Oriented aggregate of clay (light gray to white) forms thin "clay skin" that nearly envelopes grain. Oriented clay along upper and lower right, nearly straight margins of grain contributes to the plexoidal fabric apparent in Figure 10-B.

for the sake of the terminology itself, but because of the importance attached to clay fabrics in the microscopic examination of soils, as opposed to the lack of importance attached to the subject in sedimentary petrology. Evaluation of the fabrics using terminology from soil science, however, necessitates the proviso that similar fabrics may be produced by different processes, and that soil-like fabrics in sedimentary rocks may or may not indicate the operation of soil-forming processes in the genesis of the rocks (Teruggi and Andreis, 1971). Nonetheless, thin-section studies suggest that most massive clay rocks in the Longford Member, red-mottled or not, show fabrics similar to those described above and in Table 2.

The amount of carbonaceous matter in Longford clay rocks ranges widely. Light-gray and red-mottled clay rocks contain either no carbonaceous matter or only sparse disseminated flecks and fragments of carbon. Dark-colored and brownish-gray clay rocks, on the other hand, may contain abundant carbon, not only as disseminated flecks, but also as carbonized films and recognizable plant debris. Where the clay rocks contain abundant carbon, they may also contain appreciable marcasite or pyrite, much of it localized by the carbonaceous matter. The iron sulfide occurs as scattered nodules in the rocks or as disseminated, nearly microscopic, polycrystalline aggregates. Much of it also occurs as films on, and replacements of, carbonized woody tissue. Weathering of the iron sulfides leads to the development of abundant "limonite" and jarosite (?) stain, as well as the growth of gypsum crystals and "sunbursts" in surface exposures of the clay rocks. As the amount of carbonized plant debris increases, the clay rocks grade into lignite seams.

Layers of lignite, which are most abundant in the southern parts of the Longford outcrop belt in southwestern Dickinson County and northwestern Marion County (Fig. 1), are intercalated with and grade both vertically and laterally into dark-gray and brownish-gray clay rocks. Schoewe (1952, p. 108) reported a 22-inch (0.56 m) lignite layer in SE ¼ sec. 24, T.16S., R.1E., Dickinson County, but most layers are less than 10 cm thick. Most of the lignite is composed of transported and macerated, carbonized plant debris that is embedded in a clayey to silty matrix. The lignite commonly shows thin lamination, much of it wavy or contorted. The carbonized plant debris occurs as thin films that follow the lamination of the rock, and as fragments of carbonized wood as much as 1 cm thick. Many of the carbonized wood fragments have rounded edges and appear to have been abraded. Few of the carbonized films and wood fragments measure more than 10 cm long, but the writer did find one broken, carbonized log measuring about 7 cm in diameter and nearly 15 cm long. The carbonized plant material commonly contains replacement patches and coatings of marcasite or pyrite. Weathering of the iron sulfide yields abundant "limonite" or jarosite (?) stain in the lignite seams.

The lignite layers commonly overlie gray, carbonaceous clay rocks that show little or no sign of bedding. In parts of southwestern Dickinson County, the underlying clay rocks contain carbonized remains and molds of rootlets measuring as much as 2 mm in diameter and as much as 5 cm long. The clay rocks grade sharply into the overlying lignite as they become laminated in their upper parts and carbonaceous films increase in abundance upward. The thickest lignite layers in southwestern Dickinson County occur near the top of the member. They are separated from the cap siltstone by less than 2 ft. (0.6 m) of gray, wavy-laminated siltstone into which they grade. Their stratigraphic position approximates that of the nearly black claystone (unit 4) in measured section 3, Plate 2.

One piece of silicified gymnosperm wood (L. W. Macior, 1976, written communication) was found in a sandy zone just above a lignite seam near the base of the member, near NE cor. sec. 18, T.17S., R.2E., Marion County. The silicified wood measured about 10 cm in diameter and nearly 50 cm long. It lacks marked seasonal banding and indicates an equable climate.

**Siltstone**—The character of siltstone in the lower parts of the Longford Member varies widely from place to place. Some is similar to the siltstone that marks the top of the member. It forms resistant layers composed of well-sorted, medium to coarse silt grains and contains but little interstitial clay; sedimentary structures range from even, horizontal lamination and low-angle cross-stratification to transverse ripple marks and micro-cross-stratification; and it contains only scattered carbonaceous matter and plant debris. Other siltstone beds or lenses also are well indurated, but they show little sign of internal bedding. Some siltstone contains appreciable clay and grades upward or downward into either mudstone or shaly clay rocks. Colors range mainly from nearly white to yellowish brown. Like the siltstone at the top of the member, that in the lower parts locally contains scattered sand-sized grains of quartz. In thin section, much of it resembles the siltstone shown in Figure 6.

Unusual siltstone crops out near the base of the Longford Member in northeastern Clay County (Fig. 12). The siltstone is nearly white, but it is marked by numerous, irregularly shaped, subvertical streaks and mottles of red to purplish-red hematite that are as much as 3 ft (0.9 m) long. Slopes developed on the siltstone commonly have miniature hoodoos on which burr-like aggregates of hematite weathered from the siltstone form protective caps. Upper contacts with overlying light-gray, unmottled siltstone or with red-mottled mudstone are gradational. No contacts with underlying rocks were seen. The siltstone ranges from well to poorly indurated. Bedding is not apparent except where lenses of fine- to medium-grained sandstone are enclosed in the rock. The lenses are as much as 1.5 ft (0.45 m) thick. The sandstone shows the same coloring as the siltstone and is poorly sorted (curve D, Fig. 13). In places, the siltstone also contains abundant disseminated sand grains and locally verges on being sandstone (curve E, Fig. 13). A variant of the red-streaked siltstone crops out at the common corner of sections 9, 10, 15, and 16, T.6S., R.2E., Clay County. There, a siltstone layer having horizontal hematitic streaks rather than subvertical streaks underlies siltstone marking the top of the member. The layer is about 4 ft (1.2 m) above vertically streaked siltstone, and the horizontal streaking appears to be controlled by indistinct bedding.

**Sandstone**—Sandstone is not abundant in the lower parts of the Longford Member. Most of it occurs as pale yellowish-gray to yellowish-brown beds and lenses that are less than 10 ft (3 m) thick. The color

FIGURE 12.—Siltstone near base of Longford Member, NE cor. sec. 1, T.6S., R.1E., Clay County. Hematitic red mottles are exceptionally coarse and form irregular vertical streaks (as below hammer head and along upper right margin). This red-mottled siltstone is about 28 ft (8.5 m) below laminated, resistant siltstone marking the top of the member.

depends mainly on the concentration of iron-oxide cement, which in turn correlates well with the topographic position of exposures and the amount of iron-oxide case hardening induced by Pleistocene or Holocene weathering. Most Longford sandstone is at or near the base of the member, primarily in Clay and Ottawa counties (Fig. 1), but some seems to be higher in the section, not far below the siltstone that marks the top of the member. The most prominent lens of sandstone is as much as 20 ft (6 m) thick and forms the bulk of the Longford outliers in sections 16 and 21, T.9S., R.2E., Clay County (Pl. 1). In a few places, lenticular, upward fining beds of sandstone less than 1.5 ft (0.45 m) thick are enclosed by red-mottled mudstone and siltstone. The upper boundaries of the beds are gradational, but their lower boundaries may be erosional in places. The upper and lower boundaries of most sandstone lenses, however, are not exposed. The lens of sandstone in sections 16 and 21, T.9S., R.2E., Clay County, is an exception. It has a sharp, erosional contact with underlying Longford clay rocks.

Stratification in Longford sandstone lenses ranges widely in character. Some sandstone is indistinctly stratified, whereas stratification in other sandstone has been obliterated by iron-oxide diffusion structures of surprising complexity and size. Low-angle cross-strata were noted in poorly exposed, local lenses of fine-grained sandstone at the base of the member. Small- and medium-scale wedge-planar and trough sets of high-angle cross-strata are common in the prominent sandstone outlier in sections 16 and 21, T.9S., R.2E., Clay County, even though bedding is much obscured by iron-oxide cement and diffusion structures. Twenty-two cross-strata dip-bearings measured at that locality gave a vector resultant oriented S44°W. The standard deviation and consistency ratio (Reiche, 1938) computed for the dip-bearing measurements were 22° and 0.929 respectively. These few measurements are in approximate accord with the southwesterly orientation of the depositional slope of the Kiowa Formation (S60°W to S70°W) inferred from wider studies of cross-stratification (Fig. 1; Franks, 1966, 1975).

Prominent sandstone also caps an outlier of Longford rocks in NW ¼ sec. 27, T.8S., R.2E., Clay County (Pl. 1). The stratigraphic position of the sandstone, however, is uncertain owing to the lack of exposures of the rocks immediately beneath the sandstone. The sandstone either is near the top of the Longford Member, or it is near the base of the higher parts of the Kiowa Formation. The thickness of the Longford Member in areas to the west and southwest of section 27, however, suggests that the sandstone may well be within the upper parts of the member. Abundant iron-oxide diffusion structures obscure much of the bedding, but tabular- and wedge-planar sets of high-angle cross-strata allowed measurement of 20 dip-bearings. The vector resultant computed for the measurements is S37°W. The standard deviation and consistency ratio of the measurements are 19° and 0.957 respectively.

Longford sandstone ranges from very fine grained to coarse grained and conglomeratic. Most conglomeratic sandstone is at or near the base of the member where pebbles and granules of chert, quartz, and other resistant materials occur along the Permian-Cretaceous unconformity (Table 1). The pebbles are well rounded and measure as much as 6 cm in long diameter, but most are less than 3 cm long. Although much of the conglomeratic sandstone occurs at the base of the member, several feet of clay rocks locally separate conglomeratic sandstone from underlying Permian rocks, as in the outlier of sandstone in sections 16 and 21, T.9S., R.2E., Clay County (Pl. 1). Sandstone higher in the member is not conglomeratic.

Examination of three thin sections of Longford sandstone indicates that it tends to be mineralogically mature, but texturally submature (Folk, 1968). It is similar to other Kiowa sandstone, which, however, tends to be both mineralogically and texturally mature (Franks, 1966, 1975). The grains in Longford sandstone range from angular to well rounded (Powers, 1953), but most are subangular to rounded. Quartz grains generally amount to 90 percent or more of the sandstone. Few of the grains are composite grains (polycrystalline quartz and chert). Some quartz grains show partial, poorly developed quartz overgrowths. Where abundant iron oxide cements the rocks, the margins of the quartz grains commonly are etched. Detrital feldspar, chiefly potash feldspar partly altered to clay, amounts to less than 3 percent of the sand fraction. Scarce grains of heavy minerals (zircon, tourmaline, and staurolite) accord with the characteristic heavy mineral assemblages in the Kiowa Formation (Franks, 1966, 1975). Detrital micas are a minor component in Longford sandstone. Sandstone near the base of the member commonly contains fragments of penecontemporaneously reworked mudstone and siltstone that measure as much as 4 mm long.

Iron-oxide cement and clay matrix may amount to as much as 20 percent of some sandstone samples. Silt and clay particles, however, generally do not exceed 10 percent of much Longford sandstone, which tends to be relatively friable. The sandstone in the outlier in sections 16 and 21, T.9S., R.2E., Clay County (Pl. 1), was quarried for concretionary masses of calcite-cemented rock that was used as road metal. The same sandstone body locally contains abundant barite sand-crystal rosettes. The calcite and barite cement are like that described for other Kiowa sandstone by Swineford (1947).

A comparison of the grain-size and sorting characteristics of Longford and other Kiowa sandstone may aid in interpretations of provenance and depositional environments of Longford rocks. Sieve analyses of two samples of sandstone collected near the base of the Longford Member (one from a lens in red-mottled siltstone, curve D, Fig. 13; and one from the sandstone outlier in sections 16 and 21, T.9S., R.2E., Clay County, curve C, Fig. 13) show size and sorting characteristics that lie within the range of statistical parameters calculated for samples of Kiowa sandstone not from the Longford Member (Table 3). Data for a third sandstone sample (curve B, Fig. 14) are included

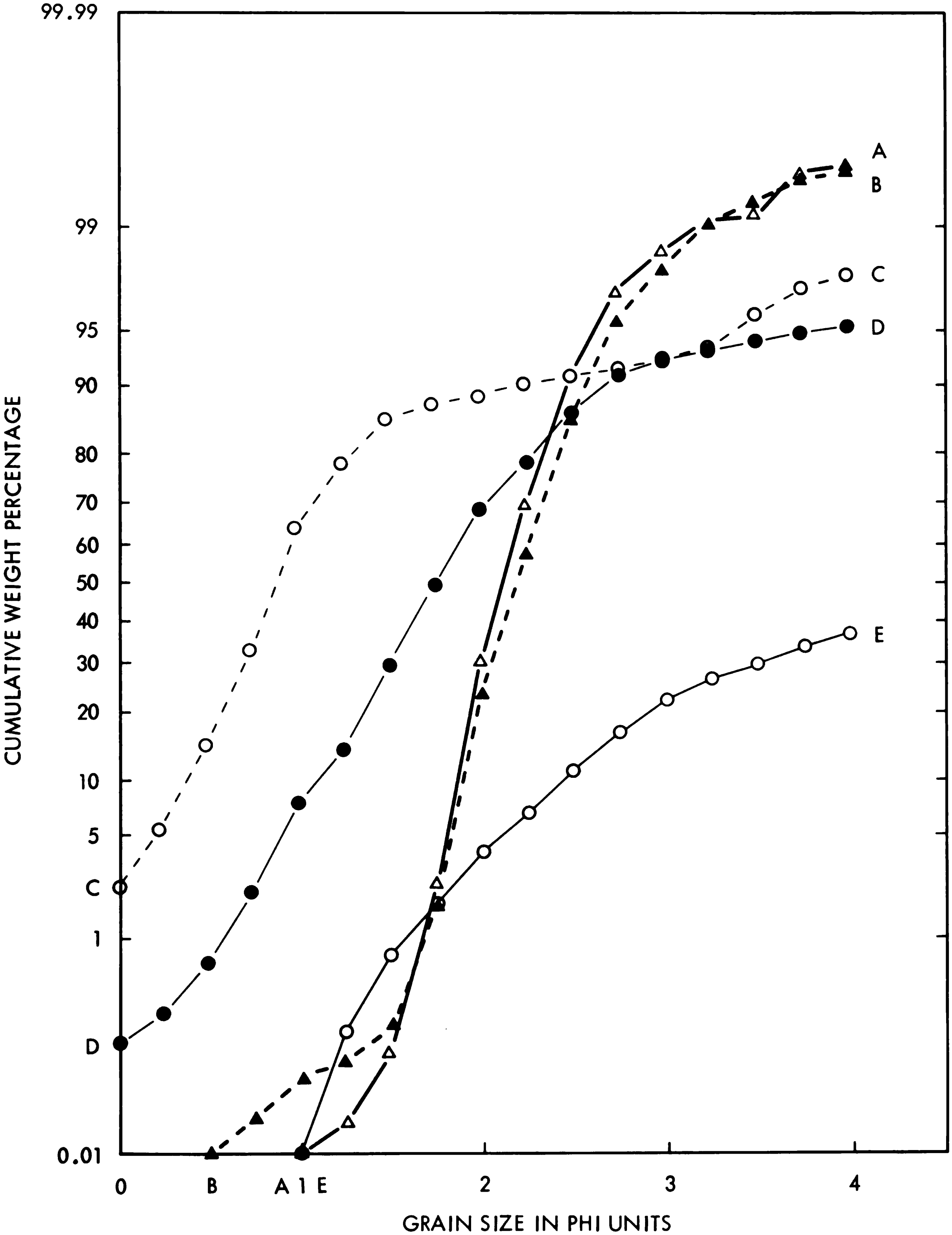



FIGURE 13.—Log-probability plots of grain-size distributions of selected samples of Kiowa sandstone above Longford Member (triangles) and of Longford sandstone and siltstone (circles). A) Kiowa sandstone, W ½ SW ¼ sec. 3, T.11S., R.1E., Dickinson County. B) Kiowa sandstone, cen. N line sec. 25, T.9S., R.1E., Clay County. C) Longford sandstone, E ½ NW ¼ sec. 21, T.9S., R.2E., Clay County. D) Longford sandstone, NE cor. sec. 1, T.6S., R.1E., Clay County. E) sandy Longford siltstone, NE cor. sec. 1, T.6S., R.1E., Clay County.

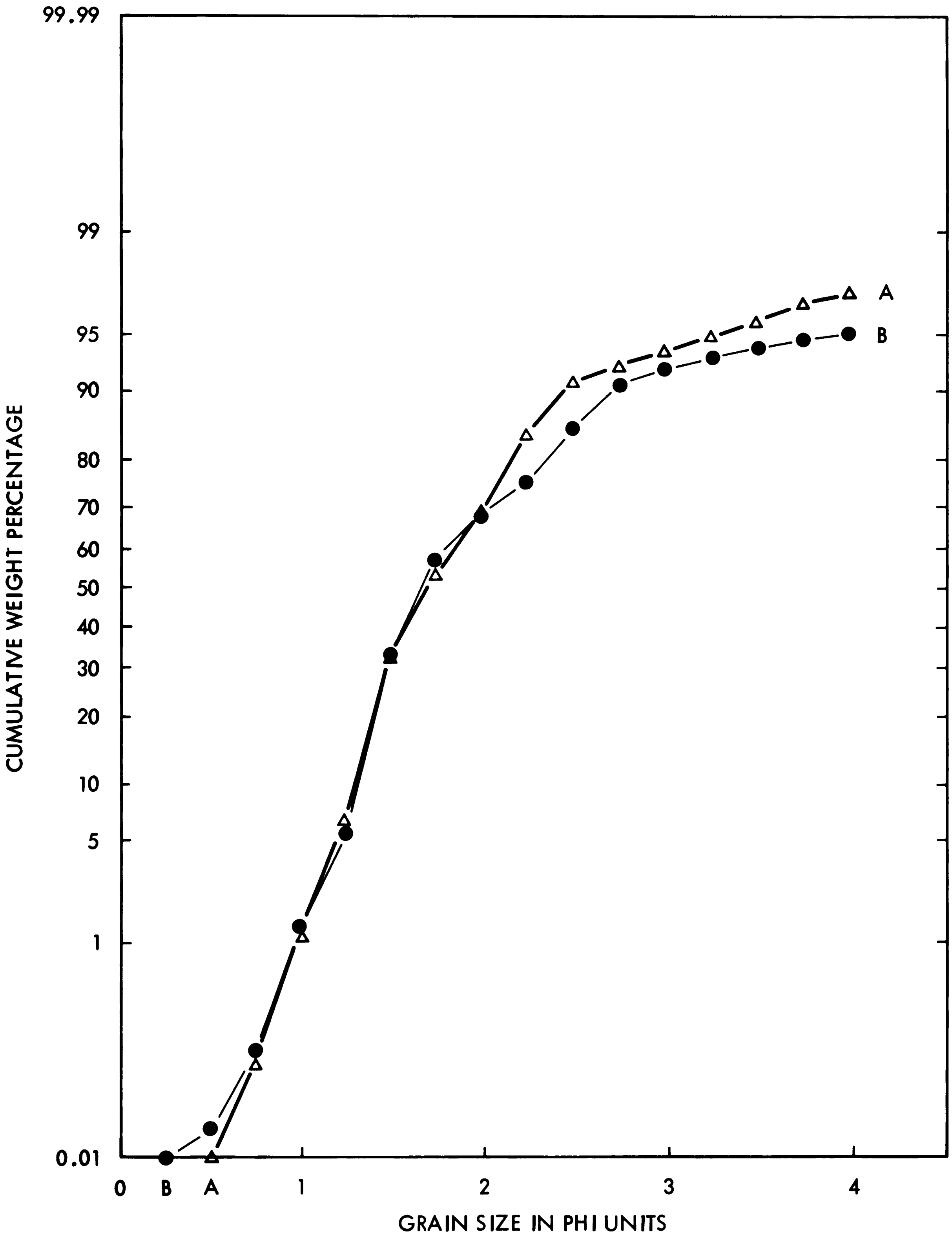



FIGURE 14.—Log-probability plots of grain-size distributions of selected samples of Kiowa sandstone (triangles) above Longford Member and of Longford (?) sandstone (circles) collected near top of member. A) Kiowa sandstone, SE ¼ SW ¼ sec. 17, T.10S., R.1E., Clay County. B) Longford (?) sandstone, NW ¼ sec. 27, T.8S., R.2E., Clay County.

TABLE 3.—Results of sieve analyses of two samples of Longford sandstone and one sample of Longford (?) sandstone compared with results from 23 samples from other parts of the Kiowa Formation. Parameters from Inman (1952) and Folk and Ward (1957). Data for Kiowa sandstone samples from Franks (1966).

| | | Median diameter (mm) | Phi median diameter | Phi graphic mean | Phi deviation measure (sorting) | Phi inclusive graphic standard deviation | Phi skewness measure | Phi inclusive graphic skewness | Phi graphic kurtosis | Normalized phi graphic kurtosis |
|---|---|---|---|---|---|---|---|---|---|---|
| Longford Samples | Curve C, Fig. 13 | 0.54 | 0.89 | 0.96 | 0.46 | 0.72 | +0.22 | +0.51 | 2.40 | 0.71 |
| | Curve D, Fig. 13 | 0.29 | 1.77 | 1.84 | 0.57 | 0.78 | +0.18 | +0.57 | 1.80 | 0.64 |
| Longford (?) | Curve B, Fig. 14 | 0.31 | 1.67 | 1.84 | 0.55 | 0.69 | +0.47 | +0.89 | 1.44 | 0.58 |
| 23 Kiowa Samples | Mean | 0.20 | 2.30 | 2.34 | 0.37 | 0.42 | +0.13 | +0.22 | 1.35 | 0.56 |
| | Standard Deviation | ...... | 0.53 | 0.53 | 0.13 | 0.17 | 0.13 | 0.14 | 0.46 | 0.06 |
| | Range | 0.54, 0.10 | 0.90, 3.32 | 0.96, 3.29 | 0.22, 0.68 | 0.22, 0.78 | −0.11, +0.48 | −0.05, +0.60 | 0.87, 2.90 | 0.48, 0.70 |

in Table 3, but the sandstone is assigned questionably to the Longford Member. The sample was taken from the previously described sandstone (p. 24) that caps an outlier of Longford rocks in sec. 27, T.8S., R.2E., Clay County (Pl. 1). Its size and sorting characteristics also are within the range calculated for other Kiowa sandstone (Table 3). The two Longford samples (Table 3) are only moderately sorted (Folk, 1968), whereas the Longford (?) sample is moderately well sorted. The grain-size distributions of all three samples are strongly skewed toward the fine end. They are leptokurtic, indicating that the tails of the distributions are less well sorted than the central parts. The data in Table 3, together with size-analysis curves for Kiowa sandstone collected less than 30 ft (10 m) above the Longford Member (curves A and B, Fig. 13; curve A, Fig. 14), suggest that much Longford sandstone is coarser grained, less well sorted, more positively skewed, and more leptokurtic than most Kiowa sandstone.

# DEPOSITIONAL ENVIRONMENTS OF LONGFORD ROCKS

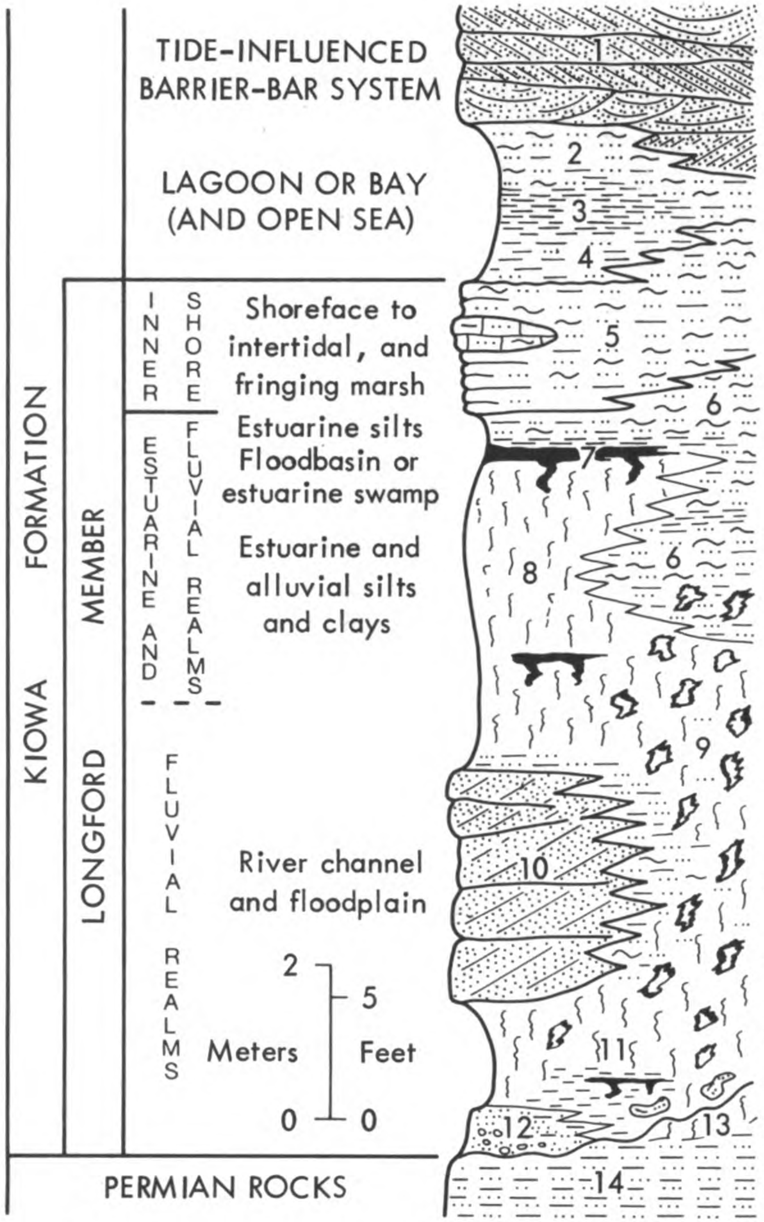
## GENERAL SETTING

Environments of deposition inferred for the Longford Member and associated Kiowa rocks are shown in Figure 15, a schematic and generalized composite section that is based on numerous exposures, no one of which spans the complete sequence of rocks represented in the figure. Longford sediments were deposited as corresponding Kiowa sediments belonging to parts of the *Inoceramus comancheanus* and *I. bellvuensis* zones of Scott (1970a, 1970b) accumulated in open-sea and sea-margin environments (Scott, 1970b; Franks, 1975), and as the epicontinental Kiowa sea spread northeastward across central Kansas (Fig. 2). Longford sediments were deposited mainly landward of the transgressing sea, before the onset of regressive conditions that heralded deposition of the nonmarine sediments of the Dakota Formation (Fig. 2). Accordingly, inferences about Longford depositional environments depend greatly on the framework of transgressive sedimentation established by facies-equivalent Kiowa rocks. Because Longford sediments accumulated near the shores of the Kiowa sea, two aspects of Kiowa sedimentation become especially important to analysis of Longford depositional environments: one is the nature of Kiowa shorelines; the other is the extent to which tidal forces influenced both Kiowa and Longford sedimentation. These two aspects of Kiowa shorelines will be reviewed in the pages that follow. Then, the major sedimentary environments indicated in Figure 15 will be taken up.

**Linear clastic Kiowa shorelines**—The Kiowa sea flooded the gentle topography that had been established on Permian bedrock. Although an embayed coast having drowned river valleys or estuaries can be inferred readily, barrier-bar or linear clastic shorelines (Selley, 1970) also were an important aspect of Kiowa sedimentation in north-central Kansas (Franks, 1966, 1975). Linear clastic shorelines are characterized by two high-energy zones, one of which is appreciably weaker than the other (Selley, 1970, p. 95). The highest energy zone is along the barrier-bar system where waves and tidal action may be intense; the weaker of the two high-energy zones is on coastal plains and along the inner shores of lagoons and bays formed behind the barrier system. There the intensity of waves and tides is reduced because of limited fetch and restricted connections to the open sea, and fluvial currents are reduced in intensity as streams approach the coast. The bulk of Longford sediments was deposited in the second, weaker, high-energy zone landward from Kiowa barrier systems: along the inner shores of lagoons or bays, in drowned valleys or estuaries, and in river valleys (Fig. 15).

Several lines of evidence indicate that linear clastic shorelines characterized Kiowa sedimentation in north-central Kansas, and that thick, lenticular deposits of Kiowa sandstone (Franks, 1966, 1975) formed barrier systems. Upward coarsening gradations from typical Kiowa shale through sequences of interlaminated shale, siltstone, and sandstone, into the thick, lenticular deposits of sandstone (Fig. 15) accord with the depositional records of barrier bars (Davies and others, 1971). Flaser bedding, transverse ripple marks, interference ripple patterns, and U-shaped burrows within the interlaminated parts of the gradations also are in keeping with the interpretation. One articulated specimen of the mactrid clam *Flaventia* was found in a conglomeratic lens in one of the thick, lenticular deposits of sandstone (Franks, 1966, 1975). The diverse kinds and orientations of high-angle cross-strata within the thick, lenticular deposits and the presence of seaward-dipping, large-scale, low-angle cross-strata in the upper parts of some of the deposits also support the barrier-bar interpretation. Some of the high-angle cross-strata dip southwestward, parallel to the depositional slope of the Kiowa Formation, but much of the high-angle cross-stratification dips to the south and southeast, parallel to the length of many of the thick, lenticular deposits and parallel to the depositional strike of the Kiowa Formation (Fig. 1). The southward dipping cross-strata probably reflect longshore transport of sand in the shoreface zone. The low-angle cross-strata formed in beach-face sands.

Log-probability plots of grain-size distributions of samples from thick, lenticular deposits of Kiowa sandstone also point to deposition of the sand in barrier systems. The plots also contrast sharply with those from Longford sandstone. Curves A and B, Figure 13, are representative of the log-probability plots of size-analysis data from a large number of samples taken from thick, lenticular deposits of Kiowa sandstone. According to Visher (1969), the curves can be subdivided into nearly straight-line segments that reflect those parts of the grain-size distributions that were deposited as components of suspension, saltation, and traction grain populations. Washing of fines from the sandy sediment also influences segmentation of the




1. SS. F-M. M- to S-scale wedge- to tab-planar and trough cross-strata sets; dips to SW, S, SE; L-scale low-angle cross-strata locally; dips to SW. Granules of quartz and clay pebbles locally concentrated on scour surfaces. 0-30 ft (0-9 m).
2. Interlam ST, SS, and SH. LtGy-YwGy. Horizontally stratified; ripple lam and flaser bedding common. 0-10 ft (0-3 m).
3. SH. M-DkGy; weathers OlGy. Thin-lam; fissile, plastic. Illitic and smectitic. Impure siderite concs locally. 0-30 ft (0-9 m).
4. SH. Lt-DkGy. Commonly Sty. Lam to thin-lam; poor fissility; plastic. Smectitic, kaolinitic, and illitic; pyrite or marcasite common. 1-8 ft (0.3-2.4 m).
5. Capping ST. M-C silt. VLtGy-GyOr. Mostly horizontally stratified; gently inclined sets of strata locally; ripple lam and S-scale cross-strata common; oscillation and current ripple marks, especially near middle and top of unit. Sp *Planolites*-like and vertical tubular burrows; imprints of and Cbnized plant debris common; reed (?) molds and casts locally; branching stems or rootlets and insect trails common where unit is within a few feet of lignite. Grades into next below. 3-15 ft (1-4.6 m).
6. ST. VLtGy-GyOr. Commonly clayey and grading down to MS; commonly friable. Thin-bedded to thin-lam; wavy lam common; Sp S-scale cross-strata and transverse ripple marks; bedding commonly indistinct. Sp lenses or beds of Sty SH and F SS lacking sign of internal bedding. Cbnized macerated stem and leaf debris locally common; also Cbnized wood frags. Locally blotched by R hematitic mottles. 0-20 ft (0-6 m).
7. Plastic CS and lignite. BrBk. Grade laterally into each other. CS kaolinitic and smectitic. Lignite thin-lam; composed mainly of Cbnized wood macerels in Sty matrix. Grades into next below. 0-2 ft (0-0.6 m).
8. MS and CS. BrGy-Gy. Kaolinitic and smectitic. Slickensided fracs and Cbn flecks throughout; Cbnized rootlets common near top. Locally encloses seams of lignite and Cbnaceous CS. 0-15 ft (0-4.6 m).
9. MS and CS. R-mottled. Kaolinitic and smectitic. Blocky structure; slickensided fracs. Sp siderite spherules and Cbn flecks only locally present. May grade laterally or vertically to ST or BrGy clay rocks. 0.25 ft (0-7.6 m).
10. SS. M-C. Lenticular; M-scale tab- and wedge-planar cross-strata sets; dips to SW. 0-25 ft (0-7.6 m).
11. MS. BrGy-Gy. Kaolinitic and smectitic; Sp RBr hematitic mottles. Locally laminated and enclosing seams and lenses of lignite and Bk plastic CS. Locally contains contorted clasts of weathered Permian rocks. Similar to 8 above. 0-15 ft (0-4.6 m).
12. SS to ST. VF-silt. SH partings common; S-scale cross-strata and ripple lam. May contain A macerated Cbnized plant debris. Locally contains granules and pebbles of chert, quartzite, etc. 0-5 ft (0-1.5 m).
13. Fossil soil at top of Permian rocks. Kaolinitic to smectitic. Variegated to pale GnGy. 0-5 ft (0-1.5 m).
14. ST, Sty SH, and MS. RBr-OlGy and GnGy. Illitic and chloritic or vermiculitic.

**FIGURE 15.—Schematic, highly idealized, composite sequence of Longford and associated Permian and Kiowa rocks, showing inferred environments of deposition. Any single Longford rock type, or combination of rock types, beneath the capping siltstone (unit 5) can be omitted from the diagram to represent local sections of Longford rocks, or sequences that actually can be deciphered from a series of exposures. Specifically omitted from the diagram are lenses or tongues of Kiowa shale enclosed by Longford rocks beneath the capping siltstone. Graphic scale provides only a reasonable measure of the orders of magnitude of thicknesses that might be found in the field. Abbreviations explained in Table 4.**

TABLE 4.—Abbreviations used in Figure 15.

| | | | |
|---|---|---|---|
| A | abundant | Lt | light |
| Bk | black | M | medium |
| Br | brown | MS | mudstone |
| C | coarse | N | north |
| Cbn | carbon | Ol | olive |
| Cbnaceous | carbonaceous | Or | orange |
| Cbnized | carbonized | R | red |
| concs | concretions | S | small, south |
| CS | claystone | SE | southeast |
| Dk | dark | SH | shale |
| F | fine | Sp | sparse |
| fracs | fractures | SS | sandstone |
| frags | fragments | ST | siltstone |
| Gn | green | Sty | silty |
| Gy | gray | SW | southwest |
| interlam | interlaminated | tab | tabular |
| L | large | VF | very fine |
| lam | laminae, laminated | Yw | yellow |

curves (Stapor and Tanner, 1975). The fluid-dynamic regimes associated with particular depositional environments are thought to impart characteristic shapes to the grain-size curves despite such factors as source-area controls, inheritance of basic grain populations from pre-existing sediment, and post-depositional changes in the sediment (Visher, 1969; Kolmer, 1973; Stapor and Tanner, 1975). Admittedly, the curves must be evaluated with some care and cannot serve as the only basis for analysis of depositional environments inasmuch as post-depositional growth of interstitial clays and cements can have profound affects on their shapes (Wilson and Pittman, 1977). Nonetheless, Visher's empirical method has been employed with some success and may have greater reliability than other techniques of environmental analysis that depend on grain-size characteristics of sandstone (Siemers, 1976; Amaral and Pryor, 1977).

In curves A and B (Fig. 13), fine-grained saltation populations form steeply sloping, well-sorted, central parts of the distributions between 1.5 and 2.75 Ø-units. The saltation populations amount to more than 95 percent of the distributions. Suspension populations are represented by the gently sloping, poorly sorted parts of the curves starting at about 2.75 Ø-units. Small, medium- to coarse-grained traction populations can be inferred from the segments between 0.5 and 1.5 Ø-units.

Environmental interpretations based on the shapes of curves A and B (Fig. 13) are somewhat ambiguous because the curves have attributes of more than one of the nearshore depositional environments examined by Visher (1969), Visher and Howard (1974), and Walton and Goodell (1972). The large, well-sorted saltation populations, combined with the small traction populations truncated near 1.5 Ø-units, may indicate that the sand accumulated in dune ridges adjacent to Kiowa beaches (Visher, 1969, p. 1083). On the other hand, the well-defined suspension populations associated with a 2.75 to 3 Ø-unit break in slope suggest wave-zone sedimentation (Visher, 1969, p. 1083, 1087; Figs. 7, 9-B), but the curves lack the well-developed, coarse-grained traction populations generally found in sublittoral or shoreface, wave-zone sediments (Visher, 1969; Visher and Howard, 1974; Sanders and Kumar, 1975). Cross-stratification in the sandstone at both sample locales, however, is similar to that in other thick, lenticular deposits of Kiowa sandstone, and it is consistent with longshore transport and sublittoral or shoreface sedimentation of the sand (Franks, 1966, 1975). It includes trough and wedge-planar sets of high-angle cross-strata whose vector resultants are oriented S35°E (curve A) and S9°E (curve B), approximately parallel to the depositional strike of the Kiowa Formation (Fig. 1). The absence of large, coarse-grained traction populations in the curves may only mean that nearshore realms in the Kiowa sea generally lacked an abundant supply of coarse sediment (cf. Table 3 and Franks, 1966, 1975).

**Evidence of tidal action**—Whether or not tides affected Kiowa sedimentation also is important to the interpretation of Longford depositional environments, especially if intertidal sedimentation is invoked to account for parts of the capping siltstone (Fig. 15). The same may be true if estuarine environments are called upon (cf. Pritchard, 1967; Schubel and Pritchard, 1971). The extent to which tides influenced sedimentation in ancient epicontinental seas like the Kiowa sea has been questioned (Shaw, 1964), but tidal effects may have been marked (Ryer, 1976; Swift, 1969, p. DS-7–DS-14), especially in epicontinental seas that had "open connections to the oceans" (Curray, 1964, p. 196). The Early Cretaceous Kiowa sea had broad connections to the oceans in the general region that now forms the Texas Gulf Coast (Williams and Stelck, 1975), and evidence that progradational tidal deposits formed along the western borders of the Kiowa sea has been reported from Colorado (MacKenzie, 1972, 1975; Weimer and Land, 1972). Franks (1975) cited evidence from southwestern Kansas that tides also influenced estuarine deposition of parts of the Cheyenne Sandstone as the Kiowa sea transgressed into that area.

Longford rocks, however, contain little definitive evidence of tidal deposits, even though tides apparently affected the Kiowa sea. The rocks lack upward fining, progradational sequences of tidal deposits such as those that Klein (1971, 1972) proposed as models for estimating ancient tidal ranges. Nor do the rocks show many of the sedimentary features, or sequences of features, described from modern, tide-controlled environments (Ginsburg, 1975; Klein, 1976; Reineck and Singh, 1973; Reineck, 1972; Terwindt, 1971; Van Straaten, 1961; among others). For understandable reasons, studies of modern tidal environments have been concentrated along coasts where the tidal range is at least as great as 1 m, and commonly much greater, and where progradation of supratidal, intertidal, and subtidal deposits is active. Few studies have dealt with the records of tidal sedimentation produced during marine transgression, particularly in regions of small tidal range. Indeed, some studies (Fischer, 1961; Kraft, 1971; Ryer, 1977; Swift, 1968) indicate that records of tidal-flat sedimentation may largely be destroyed during transgression. That abundant evidence of tidal sedimentation should be lacking in Longford rocks, therefore, is not surprising, particularly if rates of supply of sediment did not allow extensive progadation of tidal deposits.

The best evidence that tides influenced Kiowa shorelines, and presumably, therefore, Longford sedimentation, is found in the thick, lenticular deposits of Kiowa sandstone that formed the barrier systems behind which Longford sediments accumulated. Locally, exposures of cross-stratified sandstone show elements of the bedding and grain-size sequences that are produced along barrier coasts by the migration of tidal inlets (Kumar and Sanders, 1974). Indications of tidal-channel, spit-platform, and beach deposits can be recognized. Reactivation surfaces (Klein, 1970) separate sets of cross-strata in places, and the surfaces of scour channels are marked by concentrations of shale pebbles as well as granules of quartz and chert (Franks, 1966, p. 75, 82). Sparse, shallow scour channels are filled by conglomeratic lenses, which, at one locality, contained a single specimen of the mactrid clam *Flaventia*. In places, scour channels at the bases of thick, lenticular deposits of sandstone are filled mainly by shale-pebble conglomerate, the imbricated pebbles of which were derived by reworking of shale and siltstone underlying the sandstone bodies. All of the above features are consistent with the action of tidal currents along barrier coasts. They imply that tides could well have influenced inner shore sediments of the Longford Member.

Additional evidence that tides influenced nearshore Kiowa sedimentation is seen in the grain-size distribution curves in Figure 14. Both curves are significantly different from curves A and B, Figure 13, even though all of the samples came from thick, lenticular deposits of sandstone. Curve A, Figure 14, is from standstone in the lower parts of the Kiowa Formation, not far above the contact with the Longford Member. Curve B represents sandstone that is either in the upper parts of the Longford Member or in the lower part of the overlying Kiowa Formation. The sandstone caps an outlier of the Longford Member (sec. 27, T.8S., R.2E., Clay County, Pl. 1), but the exposures do not allow precise determination of the stratigraphic position of the sandstone. Both curves in Figure 14 are marked by the absence of traction populations, by moderately well sorted saltation populations extending from 0.5 to about 1.75 Ø-units, and by well-defined, poorly sorted suspension populations finer than 2.5 or 2.75 Ø-units. The suspension populations amount to nearly 10 percent of the samples. The saltation populations are less well sorted than those in curves A and B, Figure 13, but they are better sorted than the saltation populations in curves C and D. In contrast to the curves in Figure 13, the breaks between the suspension and saltation populations are not sharp. Rather, the breaks in Figure 14 consist of straight-line segments extending from about 1.75 to 2.5 or 2.75 Ø-units. The segments may stem from the mixing of fundamentally different grain-size distributions, or from the presence of a population of grains that behaved as part of a saltation population under one set of hydraulic conditions, and as part of a suspension population under higher energy conditions.

Curves A and B, Figure 14, have shapes that accord with deposition in tide-regulated or tide-influenced environments. Especially important are the gently sloping segments between 1.5 to 1.75 and 2.5 to 2.75 Ø-units that connect the saltation and suspension populations. Grain-size curves showing such segments have been reported for sands from both modern and ancient tidal channels, tidal flats, tide-controlled estuaries, and tide-regulated marine shelves (Sindowski, 1957; Visher, 1969; Visher and Howard, 1974; Walton and Goodell, 1972). Especially striking is the similarity of curves A and B to some curves from sand waves in the tide-controlled Altamaha River estuary (Visher and Howard, 1974, Fig. 18) and to suites of curves from Bajocian tidal-flat and tidal-channel sediments, as well as some curves from modern tidal-flat sands from the Frisian coast (Sindowski, 1957, p. 257, Fig. 33; p. 259, Figs. 37, 38). Curves A and B also are similar to a curve plotted by Visher and Howard (1974, Fig. 23) to illustrate the effects of fractionation on sediment that might finally be trapped in the troughs of sand waves in the mouth of a tidal inlet. Interestingly, the sandstones represented by the curves in Figure 14 not only are cross-stratified, but vector resultants computed for each locality are oriented about S40°W, approximately down the depositional slope of the Kiowa Formation, and at a large angle to cross-strata in the shoreface sands described in the preceding section (curves A and B, Fig. 13).

## FLUVIAL AND ESTUARINE ASSOCIATIONS

Most sandstone, the red-mottled siltstone, the red-mottled clay rocks, and some brownish-gray and gray clay rocks below the siltstone that marks the top of the Longford Member are products of fluvial sedimentation. Most of the sandstone probably originated as channel deposits in streams that flowed along the gentle valleys eroded into Permian bedrock. The red-mottled siltstone and clay rocks, and many of the brownish-gray clay rocks, formed as overbank deposits of those streams. The lenticular beds of lignite and nearly black claystone low in the member are interpreted as products of deposition in marshy floodbasins, whereas those near the top of the member may have accumulated as either floodbasin or estuarine-marsh deposits. Much of the wavy-laminated to poorly bedded, light-gray siltstone and some of the gray to brownish-gray clay rocks may have formed as estuarine sediments (Fig. 15).

**Sandstone**—Log-probability plots of grain-size distributions afford evidence that sandstone in the lower part of the Longford Member originated as fluvial deposits. Curves C and D (Fig. 13) contrast sharply with those from Kiowa sandstone that apparently formed as wave-zone or tide-influenced deposits (curves A and B, Fig. 13; Fig. 14). Curves C and D are marked by poorly sorted suspension populations amounting to 10 percent or more of the samples and by better sorted saltation populations coarser than 2.75 Ø-units. The shapes of the curves compare well with curves from modern fluvial sands (Visher, 1969, 1972). Southwesterly dipping, wedge-planar and trough cross-stratification in the sandstone outlier in T.9S., R.2E., Clay County (Pl. 1), is compatible with, but not diagnostic of, fluvial sedimentation. The sandstone does not show the kinds of upward fining sequences that characterize point-bar deposits or deposits formed by braided streams (Allen, 1965a; Harms and Fahnestock, 1965; Smith, 1970; Visher, 1965, 1972). Either the upper parts of point-bar sequences were not preserved, or the sand may have been deposited in the shifting channel of an unbraided, southwesterly flowing stream. Scour-fill contacts at the base of the sandstone and fragments of penecontemporaneously reworked clay rocks and siltstone incorporated in the sandstone are consistent with fluvial sedimentation. Lenses and layers of sandstone intercalated with red-mottled siltstone and clay rocks also suggest fluvial sedimentation (curve D, Fig. 13).

The Longford (?) sandstone that caps the outlier in sec. 27, T.8S., R.2E., Clay County (Pl. 1), was discussed in the section on evidence of tidal action. The precise stratigraphic position of the sandstone is not known, but it is near the top of the Longford Member or in the lower parts of overlying Kiowa strata. The sand may have accumulated in a tide-regulated part of a Kiowa barrier system, or conceivably in a subtidal channel close to the inner shore of a lagoon or bay, or as shoals near the mouth of an estuary, where ebb-oriented deposits may be common (Tucker, 1973). Whichever idea may be correct, each accords with the grain-size distribution, the southwesterly dipping cross-strata, and the location of the lens near the contact between the Longford Member and overlying parts of the Kiowa Formation. If the sandstone indeed is within the upper parts of the Longford Member, it indicates a complex interplay of estuarine and nearshore facies, not unlike that described along the South Carolina coast by Colquhoun and others (1972). The same is true for lenses of typical Kiowa shale locally enclosed by the Longford Member.

**Red-mottled clay rocks and red-mottled siltstone**—Red-mottled Longford claystone, mudstone, and siltstone (Pl. 2; Figs. 7, 12, 15) are inferred to be products of overbank sedimentation by the streams that deposited lenticular Longford sands as they flowed southwestward down broad, gentle valleys eroded into Permian bedrock. The red-mottled clay rocks and siltstone, however, lack the well-developed bedding and lamination of many modern overbank deposits (Allen, 1965b, p. 145-155). Instead, they form massive, lenticular bodies and are marked by small-scale slickensided fractures and by microscopic domain and plexoidal fabrics (Figs. 10, 11; Table 2). The lack of small-scale bedding features, the massive character, the slickensided fractures, and the microscopic fabrics can be interpreted in two ways: (1) as products of flocculation and dewatering (Meade, 1964) of overbank sediments (Franks, 1966, 1975), or (2) as products of soil-forming processes. The "clay skins" along the boundaries of detrital quartz grains (Fig. 11) and along fractures in the clay rocks (Fig. 10-C) strongly suggest a pedogenic origin for the fabrics (Brewer, 1964), and soil-forming processes do lead to the development of relatively homogeneous silts and clays from levee, floodplain, and floodbasin sediments (Allen, 1965b; Edelman and Van der Voorde, 1963).

Grain-size distributions provide added insight into the environments of deposition of some of the red-mottled siltstone. Curve E, Figure 13, is from the poorly sorted, coarse-grained, red-mottled siltstone shown in Figure 12. The siltstone encloses layers of fluvial sandstone represented by curve D (Fig. 13), and curve E can also be examined according to the methods of Visher (1969). The curve shows a weakly defined, poorly sorted saltation population that merges with a large, poorly sorted suspension population. The shape of the curve is markedly similar to the shapes of some curves for materials deposited by turbidity currents (Visher, 1969, p. 1100-1103, Figs. 20-B, 20-D). Curve E could well reflect deposition of suspended sediment as levee or other overbank accumulations of the fluvial system that deposited the sands represented by curve D (Allen, 1965b, p. 145). Alternatively, the shape of the curve might be attributed to post-depositional homogenization of overbank sediments by soil-forming processes, and the admixing of sand, silt, and clay from originally better sorted sediments. Either way, the grain-size distribution of the siltstone is compatible with overbank fluvial sedimentation.

The origins of the red, hematitic mottles in Longford clay rocks and siltstone (Figs. 7, 12) also are pertinent to considerations of Longford sedimentary environments. A genetic relationship between the red mottling and the fabrics shown by the clay rocks is suggested by the common concentration of hematite stain and coatings along slickensided and other small-scale fractures in and near the red mottles in the clay rocks (Fig. 10-C). Moreover, the mottles show striking similarity, both in gross aspect and in detail, to so-called soft plinthite or hematitic mottles that abound in a variety of lateritic tropical soils (Soil Survey Staff, 1960, p. 62; Mohr and others, 1972, p. 190). As is true for the fabrics shown by the clay rocks, the red mottles may also stem from soil-forming processes.

Franks (1966, 1975) suggested that the red mottles in Longford and Dakota clay rocks formed by way of early diagenetic, partial reduction and redistribution of iron oxides in sediments that initially were stained red or brown by iron oxides. The presence of reworked clasts of red-mottled claystone and mudstone in Dakota clay-pebble conglomerate (Franks, 1966, p. 211) and the presence of red mottles in subsurface Dakota rocks (Swineford and Williams, 1945) were cited as evidence that the mottles developed only shortly after deposition of the sediments. Evidence for partial reduction and redistribution of iron oxides in both Longford and Dakota rocks includes the light-gray to greenish-gray interareas between the mottles, the local presence of spherulitic siderite in the rocks, and the persistence of flecks of carbonaceous matter and trace amounts of pyrite in the light-gray to greenish-gray parts of some of the mottled rocks. An early diagenetic origin for the red mottles also is consistent with their generation by soil-forming processes operating in unconsolidated alluvial deposits.

Development of the red mottles in Longford clay rocks and siltstone most likely occurred during the initial stages of soil formation in alluvial sediments, during so-called soil ripening (Pons and Zonneveld, 1965), but red mottles also are characteristic of the B zones of groundwater laterites (Mohr and others, 1972). Soil ripening entails dewatering of sediments; oxidation and removal of organic matter; reduction, redistribution, and oxidation of iron compounds under the influence of fluctuating water tables (gley formation); consequent color mottling; and biological and physical homogenization of mineralogically and texturally immature soils before well-developed soil profiles can form. Color mottles are characteristic of gleyed hydromorphic soils developed in alluvial sediments (Buringh, 1970, p. 71; Edelman and Van der Voorde, 1963, p. 259-260), and gleyed soils are common in the floodplains, deltas, and estuaries of modern tropical rivers. One such river, along which mottled alluvial soils are well documented, is the Amazon in Brazil (Sombroek, 1966; Sioli, 1966). A shallow, fluctuating water table in low-lying tropical regions leads to the formation of immature gley soils, whereas groundwater laterites form in somewhat elevated materials where analogous reduction and oxidation of iron compounds are controlled by perched water tables, or by lateral flow of groundwater, to yield so-called pseudogley soils (Schlichting, 1973; Kovda, 1973; Mohr and others, 1972, p. 218-219, 229).

The red-mottled clay rocks of the Longford Member, as has been noted, closely resemble red-mottled clay rocks of the Dakota Formation, for which a fluvial origin generally has been accepted (Twenhofel, 1924; Tester, 1931; Plummer and Romary, 1942; Franks, 1975; Siemers, 1976). Thorp and Reed (1949), moreover, suggested that the red-mottled clay rocks of the Dakota Formation are similar to the red-mottled parts of lateritic soils, particularly the mottled zones of groundwater laterites that were described in the Amazon region of Brazil by Marbut (Marbut and Manifold, 1926?; Marbut, 1932?). They also likened "cellular, slaglike 'ironstone'" in the Dakota Formation to laterite duricrusts (the classical hardened laterite of Buchanan, 1807, or the hard plinthite of the Soil Survey Staff, 1960). Franks (1966, p. 163-165; 1975) noted, however, that materials interpreted as laterite duricrusts by Thorp and Reed were not products of penecontemporaneous lateritic weathering of Dakota sediments. Rather, geologically young (Pleistocene or Holocene) weathering of surface exposures of Dakota clay-pebble conglomerate and sandstone gave rise to the cellular and irregular, concretionary iron-oxide structures. Even so, Thorp and Reed (1949) may have been closer to the mark than Franks (1966, 1975) allowed inasmuch as soil-forming processes can account for the development of red-mottled Dakota and Longford rocks, and it might be tempting to assign a groundwater laterite origin to the red-mottled clay rocks and siltstone of the Longford Member.

The red mottles in Longford rocks do show a marked resemblance to mottled soft plinthite that characterizes the B horizons of groundwater laterites and other latosols of tropical regions (Mohr and others, 1972, Pt. II, Chpt. 1). Important differences, however, seemingly exist between red-mottled Longford rocks and groundwater laterites, including those developed in the unconsolidated Tertiary and Quaternary terrace deposits along the Amazon River valley (Sombroek, 1966).

(1) Longford clay rocks lack vertical changes in color and texture that correspond to the A, B, and C horizons of groundwater laterites (Mohr and others, 1972, p. 198-232; Sombroek, 1966, p. 100-111). Particularly important is the absence from the Longford Member of layers of concretionary, hard plinthite in the upper parts of red-mottled clay sections and the absence of pallid zones beneath the red-mottled rocks.

(2) The clay-mineral assemblages in red-mottled Longford rocks are less mature than those in groundwater laterites, although both are kaolinitic. Even though groundwater laterite profiles developed in the terrace deposits along the Amazon River are not as intensely weathered as most of the laterite profiles described by Mohr and others (1972), their clay fractions contain no illite, smectite, chlorite, or other 2:1 clay minerals, and they do contain gibbsite (Sombroek, 1966, p. 109). In contrast, neither red-mottled Longford rocks nor soils developed on the modern floodplain of the Amazon River (Sombroek, 1966, p. 155-157) contain gibbsite, whereas they do contain 2:1 clay minerals. Indeed, although red-mottled Longford rocks tend to be more kaolinitic than other Longford clay rocks (Pl. 2), they commonly do contain as much illite, smectite, and chlorite or vermiculite as the unmottled rocks.

(3) Red-mottled Longford rocks probably formed in low-lying floodplains, whereas the terrains in which groundwater laterites form must have some relief, even if the soils are imperfectly drained (Mohr and others, 1972; Sivarajasingham and others, 1962; Sombroek, 1966).

Fluctuating groundwater tables, whether perched or not, seem to be essential to the generation of groundwater laterite profiles, and the range of fluctuation of water tables in Longford sediments most likely was insufficient to permit the required degree of weathering. Because Longford fluvial sediments accumulated as the Kiowa sea transgressed central Kansas, they probably were not raised appreciably above the general level of the main streams. No doubt, they were water logged much of the time, and were buried by new sediment before extensive weathering could take place. Whether or not the original coloration of the mottles was red, brown, or some shade of orange is problematic, but aging of hydrated iron oxides to hematite (Fischer and Schwertmann, 1975; Van Houten, 1968) could account for their present-day redness. The properties and occurrence of red-mottled Longford clay rocks and siltstone compare most favorably with their generation as gleyed alluvial soils and products of soil ripening, rather than as products of relatively intense weathering and the formation of groundwater laterites.

**Lignite and dark claystone**—Lenses and seams of lignite and nearly black, highly plastic claystone commonly are underlain by gray clay rocks in which carbonized rootlets are preserved (cf. top of unit 4, measured section 3, Appendix B and Pl. 2; Fig. 5). Such rocks occur both high and low in the Longford Member. Lignite and dark, plastic claystone in the lower parts of the member are judged to represent floodbasin deposits of Longford fluvial systems. Floodbasin sediments commonly are penetrated by abundant rootlets; may contain abundant, transported plant debris and other organic matter; and may be marked by homogenized soils rich in organic matter (Allen, 1965a, p. 152; Edelman and Van der Voorde, 1963, p. 261). The fragmental, abraded nature of the carbonized woody debris in Longford lignite, together with the wavy bedding shown by carbonaceous films and the silty to clayey matrix, indicates that the rock is of detrital origin. Rootlets preserved in underlying gray and carbonaceous clay rocks accord with deposition of the plant debris in low-lying floodbasins of streams. The detrital lignites and associated gray clay rocks near the top of the member, however, could represent deposits that accumulated in the marshy regions of estuaries (Gorsline, 1967). The detrital character of the lignite could stem from storms that drove woody debris into low-lying terrains along the shores of estuaries, or, more likely, the plant debris could have accumulated as flood deposits on delta plains that formed at the heads of estuaries where streams entered the drowned parts of their valleys.

Dark, nearly black, plastic, kaolinitic and smectitic claystone, into which some of the lignite beds grade laterally, shows abundant slickensided fractures, churned fabrics, and downward projections of the dark clay into underlying gray clay rocks. The fractures, churned fabrics, and projections are features that are characteristic of so-called vertisols (Soil Survey Staff, 1960, p. 124). Vertisols develop deep cracks during dry spells. Surface material sloughs into the cracks and becomes thoroughly churned into the soil during repeated wetting and drying. Although vertisols are not necessarily characteristic of tropical alluvial settings, they do develop in them (Buringh, 1970, p. 74-75, 79-87; Mohr and others, 1972, p. 308-339), and would form in floodbasins where smectite-bearing clays rich in organic matter underwent surface drying between floods while the subsoil remained wet. Such soils also develop in marshy estuarine settings that are subjected to periods of wetting and drying.

**Other clay rocks and siltstone**—The abundant gray and brownish-gray clay rocks in the Longford Member probably are products of sedimentation in a variety of environmental settings. Many of them contain appreciable carbonaceous matter and some are marked by sparse reddish-brown mottles. Many of them also show slickensided fractures and lack obvious signs of bedding. Such clay rocks, like the more intensely mottled rocks, probably represent floodbasin or other overbank deposits of streams. Others that lack color mottling and obvious signs of internal bedding may represent either stream deposits or sediments that accumulated in estuarine bays. Those that show good lamination and are intercalated with light-gray siltstone near the top of the member most likely stem from deposition on or near estuarine bay-head deltas. The scarce, local lenses or tongues of typical Kiowa shale enclosed within the siltier parts of the member in parts of Saline, Clay, and Ottawa counties may indicate either marine incursions or temporary shifts of lagoonal realms into more estuarine settings during transgression of the Kiowa sea.

Like the associated clay rocks, siltstone near the top of the Longford Member, but below the capping siltstone, probably stems from deposition of silty sediment in different environmental settings (units 12 through 16, measured section 1; unit 1, measured section 2; unit 5, measured section 3, Pl. 2). Some of the silt probably was deposited as overbank stream sediments, but much of it is inferred to have deposited in estuarine realms (unit 6, Fig. 15). Inferrence of estuarine depositional environments is based on a variety of considerations: (1) on the knowledge that the Kiowa sea transgressed a mature erosion surface developed on Permian bedrock, and the likelihood that drowned valleys influenced by tides developed during the transgression; (2) on the fact that modern estuaries are traps for abundant silty sediment (Folger, 1972a, 1972b; Gorsline, 1967); and (3) on the stratigraphic position of the siltstone between fluvial sediments below and shoreline to open-sea sediments above (Tucker, 1973). Many of the siltstone beds and lenses show features that are consistent with estuarine sedimentation (Emery and others, 1957; Gorsline, 1967; Terwindt, 1971): locally, bedding surfaces show transverse ripple marks indicative of wave action; in places, ripple laminae, flaser bedding, and low-angle cross-strata are common; much of the siltstone is well sorted and nearly free of admixed clay; and carbonized plant debris and imprints of plant debris are common on bedding surfaces.

Where, within Longford estuarine systems, the silt was deposited is problematic inasmuch as the abundance and distribution of silty sediments in modern estuaries depends on a variety of factors: circulation patterns within the estuary, rates of supply and kinds of river-borne sediments, the extent to which sediment is transported into the estuary from the open sea, and tidal ranges within and without the estuary (Folger, 1972a, 1972b; Gorsline, 1967; Ryan and Goodell, 1972). Stratigraphic proximity of some of the siltstone beds to underlying fluvial deposits suggests that some Longford silt accumulated on or near deltas at the heads of estuaries, perhaps in a fashion analogous to silt near the head of Mobile Bay, Alabama (Ryan and Goodell, 1972, Fig. 17). Some of the coarse-grained silt that contains little interstitial clay, however, may have been deposited near estuary shores. Widespread, upward gradation of the silty rocks into the siltstone that caps the Longford Member seemingly marks a change from estuarine to lagoon or bay sedimentation, and a change from sedimentation dominated by fluvial processes to sedimentation dominated by coastal processes (Fig. 15).

## INNER SHORE ENVIRONMENTS

The widespread, well-sorted, coarse-grained siltstone that marks the top of the Longford Member (unit 5, Fig. 15) shows a number of features that indicate deposition of the silt along the landward, inner shores of an embayed or lagoonal coast that developed behind protective barriers formed by thick, lenticular deposits of Kiowa sand. Chief among those features is the consistent stratigraphic position of the siltstone above highly varied lower Longford rocks and below Kiowa shale and thick, lenticular deposits of Kiowa sandstone (Pl. 2, Fig. 15). The presence of siltstone rather than sandstone in that stratigraphic position emphasizes that the depositional environment was one of low energy compared to the barrier system in which facies-equivalent, lenticular deposits of Kiowa sand accumulated. The good sorting of the siltstone, compared to most other Longford rocks and to the silty and clayey rocks that mark the base of the overlying Kiowa section (measured sections 2 and 3, Appendix B and Pl. 2), certainly accords with deposition of the silt along the inner shores of lagoon or bay systems. Even though fetch and tidal range were limited, wave action and currents were strong enough or persistent enough to bring about effective washing of clays from the silty sediment (Emery and others, 1957; Folger, 1972a, 1972b; Kraft, 1971; Koefeld and Gorsline, 1963; Nichols, 1964; Ryan and Goodell, 1972; Shepard and Moore, 1960). The common presence of transverse ripple marks and interference patterns on bedding surfaces shows that wave action influenced deposition of much of the silt, perhaps in shoreface or subtidal zones of that inner shore. The preservation of scattered low-angle cross-strata is suggestive of local swash-zone or beach-face sedimentation for some of the silt (Reineck and Singh, 1973, p. 301-304; Van Straaten, 1959, p. 204), although the low-angle cross-strata could stem from other causes. Small-scale cross-stratification produced by the migration of current ripples shows that weak currents also operated in the depositional system. In combination with asymmetric, transverse ripple marks and flaser bedding, the current structures imply deposition of some of the silt in tide-influenced, inner shore settings (Reineck, 1972; Reineck and Singh, 1973; Reineck and Wunderlich, 1968; Van Straaten, 1959).

Plants contribute appreciably to the sediments of tidal flats and marshes that fringe many bays and lagoons (Phleger, 1969). The carbonaceous matter and imprints of plant debris in the siltstone at the top of the Longford Member also accord, therefore, with deposition of some of the silt in intertidal to supratidal zones, even though the sediments are not as clayey as those of many modern tidal flats and salt marshes (Reineck and Singh, 1973). Repeated gentle wave agitation, however, can and does remove much clay from intertidal sediments at some locales (Van Straaten, 1961). The reed fossils preserved locally in the lower and middle parts of the capping siltstone are especially suggestive of intertidal or supratidal sedimentation. Likewise, the branching plant stems and rootlets that occur locally near the base of the siltstone impart textures that resemble those of modern, silty to sandy tidal and salt-marsh deposits studied by Bouma (1963). The presence of probable insect burrows and trails (Siemers, 1976, written communication) near the base of the siltstone certainly implies subaerial exposure of some of the silt during part of its depositional history.

The idea of tidal-flat sedimentation for part of the Longford cap siltstone might be even more convincing were there clear evidence of scour-and-fill structures produced by tidal creeks and by tidal channels in lower tidal-flat realms (Van Straaten, 1952, 1961). The size and abundance of tidal creeks, and presumably channels in lower tidal-flat realms, depend, however, on both tidal range and the breadth of tidal flats and marshes (Van Straaten, 1959, p. 198), as well as on the particle size (and permeability?) of the tidal flat sediment (Evans, 1965). In regions of low tidal amplitude where tidal flats and marshes are narrow, few, if any, tidal creeks and channels are apt to form.

The relative scarcity of burrows in the siltstone, especially in the lower parts of the unit, and the absence of macroinvertebrate fossils might also be disconcerting. Scarcity of burrows, however, is not unusual in intertidal deposits where frequent agitation of the sediment takes place (Van Straaten, 1959, p. 199-200). The absence of shelled fossils may reflect the generally brackish character of Kiowa sedimentary environments and the northeastward decrease in salinity detected by Scott (1970a, 1970b). Lagoons and bays bordering the Kiowa sea may well have been too brackish to support extensive populations of shelled invertebrates, and any shell material washed onto or indigenous to tidal flats may have been destroyed by postdepositional dissolution.

The gross sequence of sedimentary structures, trace fossils, and plant fossils that can be established in the siltstone from place to place is compatible with gradual submergence of inner shore deposits of lagoons or bays. Deposition of horizontally stratified to wavy-laminated, largely supratidal silts locally containing abundant plant debris, rootlets, reeds, and insect trails apparently gave way to deposition of intertidal sediments. The intertidal silts show both even stratification and current ripples and locally contain reed fossils, as well as low-angle cross-strata. The intertidal realm was succeeded by deposition of shoreface or subtidal silts showing transverse ripple marks and small-scale cross-strata and locally containing abundant rod-shaped burrows. The inner shore succession in turn is overlain by deposits that probably accumulated in deeper parts of lagoons or bays.

## LAGOON OR BAY TO OPEN-SEA REALMS

Dominantly light-gray, silty to clayey shale (unit 3, measured section 2; units 7, 8, 9, measured section 3; Appendix B and Pl. 2) overlies the siltstone that marks the top of the Longford Member and is in sharply gradational contact with the siltstone. The shaly beds mark the base of the overlying parts of the Kiowa Formation. They are inferred to be products of lagoon or bay sedimentation that took place behind barriers represented by thick, lenticular deposits of Kiowa sandstone (unit 4, Fig. 15). The inference is based largely on the stratigraphic position of the beds between the capping siltstone below and characteristic Kiowa shale or thick, lenticular deposits of sandstone above. Pyritic nodules, together with "limonite" stain and gypsum aggregates produced by recent oxidation and weathering of iron sulfide, indicate that reducing conditions prevailed within the sediment. The lamination of the shaly beds does not accord with the churned and homogenized nature of lagoon or bay deposits in areas of abundant benthic life (Shepherd and Moore, 1960), but it does accord with the stratification in lagoon or bay deposits formed where benthic organisms are scarce owing to low salinity or other factors (Van Straaten, 1959). As was noted in the discussion of Longford inner shore deposits, the salinity of the Kiowa sea decreased northeastward. The water in Kiowa bays or lagoons may well have been brackish or nearly fresh and may account for the absence of shelled invertebrate fossils from the rocks.

Appreciable thicknesses of light-gray to medium-gray, silty, poorly fissile shale also occur near the base of the Kiowa Formation along the western fringes of the Longford outcrop belt in parts of Ottawa, Saline, and McPherson counties. These beds, which locally overlie or enclose brownish-gray clay rocks and shale containing appreciable carbonaceous matter, may be facies equivalents of Longford rocks. They may represent lagoon or bay deposits that formed behind Kiowa barrier bars as fluvial and estuarine Longford sediments accumulated to the north and east.

Typical, olive-gray to drab Kiowa shale overlies the light-gray shaly beds above the Longford cap siltstone in many places (unit 4, measured section 2; unit 10, measured section 3; Appendix B and Pl. 2). At least locally, typical Kiowa shale intervenes between the light-gray shaly beds and thick, lenticular deposits of Kiowa sandstone. Typical Kiowa shale is thought to represent open-sea or open-sound sediments. The shale might represent deposits that formed in the central parts of lagoons or bays, but its similarity, both physical and mineralogical, to the bulk of the shale in the Kiowa Formation suggests that the muds were deposited under conditions of more open circulation than those that might have obtained in restricted bays or lagoons (Franks, 1966, 1975; Scott, 1970a, 1970b). That the shale overlies lagoon or bay deposits and locally underlies sandstone that represents parts of barrier systems implies at least local complex shifting of shorelines and incursion of the sea into marginal-marine realms. Both the shale and the thick, lenticular deposits of Kiowa sandstone above the Longford Member indicate continued northeastward transgression of the Kiowa sea.

# MECHANISMS OF KIOWA TRANSGRESSION IMPLIED BY LONGFORD MEMBER

The abundance of thick, lenticular deposits of Kiowa sandstone in facies-equivalent parts of the Kiowa Formation in Ottawa, Saline, and McPherson counties, and elsewhere in central Kansas (Franks, 1966, 1975; Bayne and others, 1971; Scott, 1970b), indicates that Longford sediments were deposited along a lagoonal or embayed barrier coast as the Kiowa sea flooded over the gentle topography developed on Permian bedrock along the gently dipping western flank of the Nemaha anticline (Fig. 1). In many places, typical Kiowa shale overlies the inferred lagoon or bay deposits, which in turn overlie the inner shore deposits represented by the siltstone that marks the top of the Longford Member (Pl. 2, Fig. 15). Elsewhere, thick, lenticular deposits of Kiowa sandstone overlie the lagoon or bay deposits and the inner shore siltstone (e.g., in parts of southwestern Clay County, northwestern Dickinson County, and northeastern Saline County). Even though it has not been studied in detail, the apparently scattered distribution of Kiowa shale and sandstone above the fluvial to inner shore sequences of Longford Member, combined with the absence of obvious transgressive disconformities, offers insight into the mechanisms by which the Kiowa shoreline shifted landward as the sea invaded north-central Kansas (Fig. 16).

Papers by Fischer (1961), Schwartz (1967), and Swift (1968) include descriptions of more-or-less classical concepts of transgressive stratigraphy whereby shoreface erosion, combined with washover of sand from barrier islands and growth of tidal deltas into lagoons, drives barrier bars landward (Fig. 16-A). As noted by Swift (1968), shoreface erosion or ravinement can be complemented by erosion induced by migration of tidal inlets as barrier islands are extended by longshore movement of sediment. Ravinement leads to partial to complete destruction of lagoon or bay deposits as barrier islands migrate over them. Figure 16-A shows only partial destruction of the lagoonal record. One or more transgressive disconformities are introduced into the transgressive sedimentary record. Tidal-inlet sands or open-sea sands reworked from the barrier system commonly form a transgressive sheet sand that disconformably overlies the eroded remnants of lagoon or bay deposits, or whatever remains of the land surface on which the lagoon or bay sediments accumulated. The scheme does not match the sequences of Longford beds preserved in north-central Kansas.

Ryer (1977) developed a skillful modification of the transgressive model outlined above in order to account for the distribution of open-sea, barrier-island, lagoon, and fluvial sediments in strata that were deposited along the western margins of the Late Cretaceous sea of the Western Interior (Fig. 16-B). Ryer noted the diverse balances that can obtain between rates of transgression and rates of supply of sediment to barrier shorelines, and the importance of progradational periods during times of overall transgression. As above, transgression leads to ravinement and to development of transgressive disconformities that separate open-sea deposits from whatever materials the sea transgressed. Deep erosion on the shoreface and in migrating tidal inlets can eliminate all record of lagoon or bay deposits. Periods of progradation, however, lead to outbuilding of barrier-island, lagoon, and fluvial deposits (Fig. 16-B-2), which become truncated in turn by the next transgressive pulse as the supply of sediment is reduced (Fig. 16-B-3). Instead of a continuous transgressive sheet sand being deposited, discontinuous bodies of barrier-island sand are preserved. Depending on the degree of detail with which field work is carried out, the discontinuous, progradational barrier sands could be mistaken for a single, transgressive sheet sand. Moreover, careful synthesis of field data should reveal that fluvial and lagoon deposits overlie the barrier sands, which in turn overlie open-sea sediments.

The apparent completeness of the Longford record in many places and the widespread occurrence of inferred lagoon or bay deposits above the capping Longford siltstone accord with a model of in-place growth, eventual submergence, and step-wise, landward shifting of barrier islands during transgression (Fig. 16-C). Sanders and Kumar (1975) cited evidence for in-place drowning and landward shift of barrier systems along the Atlantic coast of Long Island, New York, and a similar mode of transgression accounts for the distribution of Late Cretaceous swamp, barrier-island, and open-sea deposits in the San Juan basin of New Mexico (Hollenshead and Pritchard, 1961). The process readily accounts for the apparently scattered distribution of open-sea deposits of Kiowa shale and thick, lenticular deposits of Kiowa barrier sandstone above the inner shore deposits of siltstone that cap the Longford Member. It also is consistent with the thick, lenticular deposits of Kiowa sandstone in facies-equivalent parts of the Kiowa Formation, and with the relatively thick accumulations of silty shale and

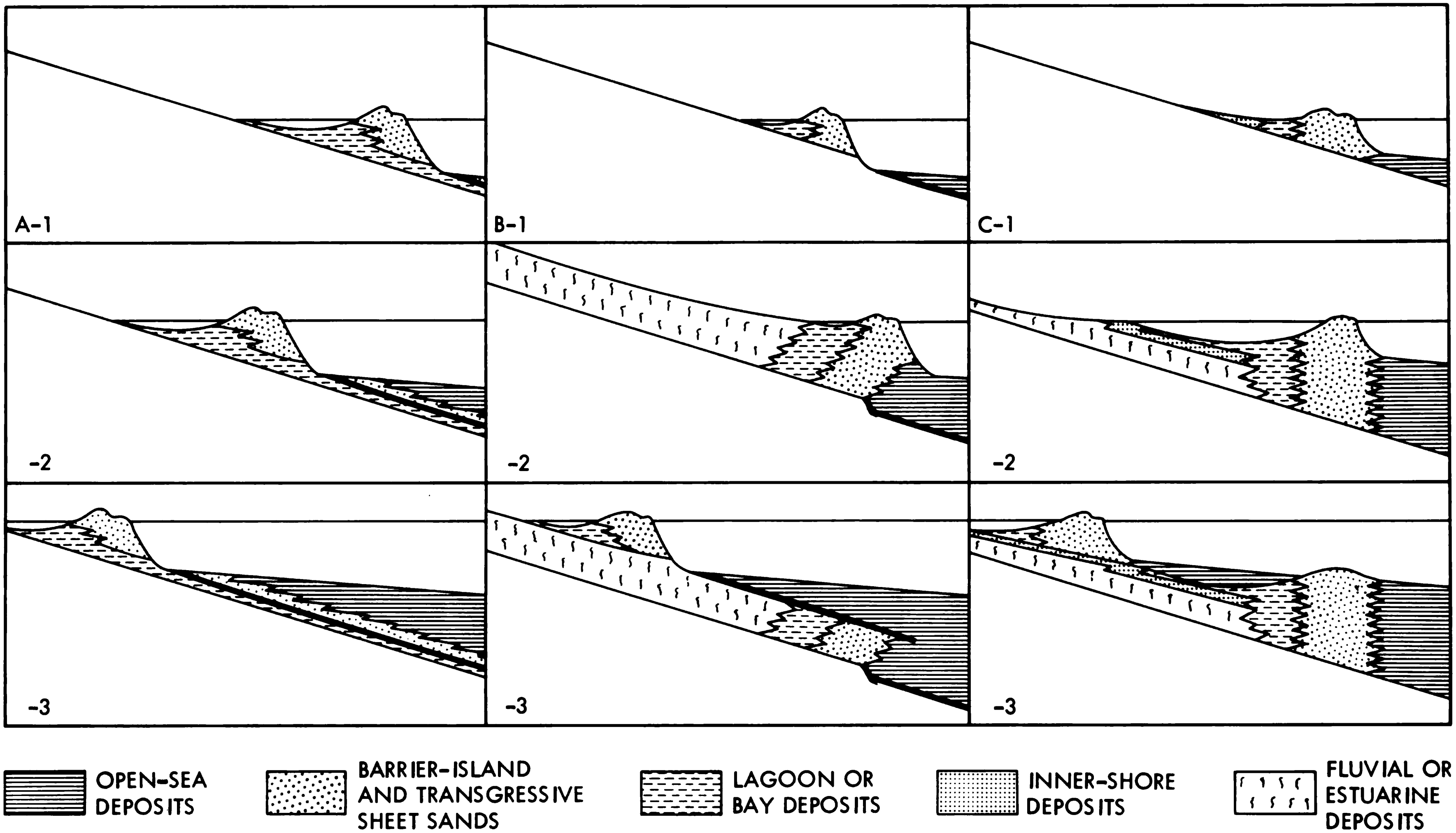



FIGURE 16.—Models of transgressive records along barrier coasts. A) Transgression and shoreface erosion, commonly complemented by erosion induced by migration of tidal inlets, resulting in partial destruction of lagoon sediments. Erosion results in transgressive disconformities; nearshore deposition may lead to development of transgressive sheet sands (after Fischer, 1961, and Swift, 1968). B) Transgression and shoreface erosion, perhaps accompanied by erosion induced by migration of tidal inlets, resulting in destruction of barrier-bar and lagoon records, but modified by progradation of barrier-island, lagoon, and fluvial sediments (after Ryer, 1977). C) Transgression accompanied by in-place growth and eventual submergence of barrier islands, leading to stepwise, landward shift of barrier-island shores and preservation of fluvial, estuarine, and lagoon or bay records. Model C applies to Longford records developed in valleys in Permian bedrock.

brownish-gray clay rocks and shale that were noted near the western limits of Longford exposures. The silty shale and brownish-gray clay rocks and shale may well represent lagoon or bay sediments that were deposited behind barrier systems while Longford fluvial and estuarine sediments were deposited in broad valleys to the east and northeast (Fig. 16-C-2).

A sufficient supply of sand must have been provided to the barrier system to counterbalance the rise in sea level and to permit upward growth of the barriers (Swift, 1968; Sanders and Kumar, 1975). The abundant, southward dipping cross-strata in thick, lenticular deposits of Kiowa sandstone accord with longshore transport of the sand (Fig. 1) (Franks, 1966, 1975; Bayne and others, 1971), and with upward growth of the barriers (Fig. 16-C-2). Closer to the barrier system, relatively thick sequences of estuarine and lagoon sediments probably formed. As sea level rose, the locus of fluvial and estuarine sedimentation shifted up valley, and inner shore deposits represented by the siltstone at the top of the Longford Member migrated landward. Eventually, the supply of sand to Kiowa barrier islands became modified, perhaps in response to the increasing width of the lagoon or bay systems, so that upward growth of the barriers could not continue. The barriers became drowned and new barrier systems were generated close to the inner shores of former bays and lagoons as surf zones shifted landward (Fig. 16-C-3). The stepwise, landward shift of the barrier systems resulted in widespread formation of open-sea deposits of Kiowa shale over lagoon or bay deposits that overlie the inner-shore siltstone, and in more localized deposition of barrier sands above lagoon and inner-shore sediments.

The thickest and most complete Longford records formed along the deeper valleys eroded into Permian bedrock on the gentle western flank of the Nemaha anticline (Fig. 1). One such valley must have been located in southwestern Clay County near type Longford. Figure 16-C can be construed as representing the sedimentary framework along such a deep valley. Elsewhere, inner shore silts were deposited directly on drainage divides as sea level rose, as in parts of northwestern Marion County (Fig. 1). The situation is analogous to the preservation of Holocene fluvial, marsh-fringe, salt-marsh, lagoonal-shore, and lagoon sediments beneath barrier-island sands in drowned valleys developed on Pleistocene coastal sediments along the Delaware coast (Kraft, 1971, p. 2152-2153). There, as in Kansas, pre-transgression topography exerted a profound control on the preservation of fluvial to lagoon sediments. The pre-transgression topography developed on Permian bedrock may also have influenced patterns of longshore transport of sand, and ultimately may have been responsible for the landward shift of barrier islands, many of which may have developed originally as spits.

# PROVENANCE OF LONGFORD SEDIMENTS AND CLAY-MINERAL DISTRIBUTIONS

Heavy minerals noted in thin sections of Longford sandstone (zircon, tourmaline, staurolite) suggest that the heavy-mineral suite in Longford rocks does not differ appreciably from the heavy-mineral suites in other Kiowa sandstone (Franks, 1966, 1975). The heavy minerals accord with derivation of much of the sand by reworking of Paleozoic sedimentary rocks in the continental interior to the east and northeast of Kansas. The staurolite, however, indicates that some of the sand may have been derived from the crystalline terrains of the central Appalachian Mountains, either directly, or by reworking of Upper Permian or Mesozoic sedimentary rocks now eroded from the continental interior (Franks, 1966, 1975).

Much of the silt in the Longford Member probably was derived from nearby Paleozoic source rocks, chiefly silty Permian or Pennsylvanian bedrock. The scarce, sand-sized quartz grains set in a matrix of silt-sized quartz in much Longford siltstone (Fig. 6) impart a texture that is suggestive of much siltstone that Swineford (1955) described from the Upper and Middle Permian of southern Kansas (cf. Swineford's Plate 19-D, p. 128). The sand-sized grains in both the Longford siltstone and the Permian rocks are rounded and show nearly straight extinction. Most of the siltstone in the upper parts of the Longford Member probably formed by reworking and washing of Longford sediments close to or along the inner shores of estuaries, bays, and lagoons, but the ultimate source for much of the silt doubtless included nearby Pennsylvanian and Permian rocks. The heavy minerals (chiefly zircon and tourmaline) noted in thin sections of the siltstone are compatible with heavy minerals in Pennsylvanian and Permian rocks (Potter and Pryor, 1961; Swineford, 1955). The absence or near absence of feldspar, which is present in Permian siltstone, can be attributed to weathering of Permian rocks prior to reworking.

The abundant kaolinite and smectite in Longford clay rocks (Pl. 2), together with the small amounts of chlorite or vermiculite detected in many of them (Fig. 8), points to derivation of much of the clay from soils that formed on dominantly illitic and chloritic or vermiculitic Paleozoic bedrock before the Kiowa sea invaded central Kansas. X-ray diffraction studies of Longford clays, including samples from a number of scattered exposures and three measured sections not shown in Plate 2, indicate that the proportions of kaolinite and smectite do not vary systematically with the kinds of Longford clay rocks. Soil-forming processes that affected Longford flood-plain sediments were not responsible for consistent changes in mineralogy in the detrital clays, and weathering of the alluvial soils was inadequate to bring about complete destruction of 2:1 clay minerals (Pl. 2).

An irregular upward increase in the proportions of kaolinite to smectite and other clay minerals is apparent in the Longford Member (Pl. 2). The upward change may be related as much to the upward increase in the abundance of siltstone in the member as it is to variations brought on either by penecontemporaneous soil-forming processes or by differences in the detrital clay assemblages supplied to Longford depositional realms. The abundance of kaolinite in silty Longford rocks or in clay rocks intercalated with more permeable silty beds could reflect postdepositional, intrastratal alteration or recent weathering of smectite to kaolinite (Glass, 1958; Glass and others, 1956; Potter and Glass, 1958; Wilson and Pittman, 1977). Alternatively, the relative abundance of kaolinite associated with siltstone might stem from sorting brought on by differential transport and sedimentation of differently sized kaolinite and smectite settling aggregates (Edzwald and O'Melia, 1975). Evidence discussed in the following paragraphs, however, suggests that differential transport and sedimentation may not have been an important factor in the development of Longford clay-mineral assemblages.

Relatively consistent variations in clay-mineral assemblages are found among Longford rocks, Kiowa lagoon or bay deposits above the capping Longford siltstone, and typical, open-sea Kiowa shale. Illite and "degraded illite" are scarce components in Longford clay rocks (Pl. 2), but they are major components in most Kiowa shale (Franks, 1966). The proportions of kaolinite and smectite in Kiowa shale range widely, but kaolinite generally is most abundant near the top of the formation, whereas smectite generally is most abundant near the base of the unit (Franks, 1966). X-ray diffraction traces from light-gray shales above the Longford Member show illite diffractions that are less intense than those obtained from most Kiowa shale, and kaolinite diffractions that are more intense than those obtained from most Kiowa shale in the lower parts of the formation. Smectite diffractions may be as intense as or less intense than those shown by Kiowa shale. Thus, the clay assemblages in the lagoon or bay deposits are intermediate between those of typical Kiowa shale and Longford clay rocks. Some indication of the trend is shown by unit 9, measured section 3, and unit 4, measured section 2, Plate 2. The lateral variations in Longford and Kiowa clay-mineral assemblages are analogous to those summarized by Parham (1966). Kaolinite is most abundant shoreward; illite is most abundant basinward; and smectite may be present in both realms, although it commonly increases in abundance basinward.

The intermediate character of the clay assemblages in the lagoon or bay deposits may stem from a variety of causes, but two of them seem more feasible than others: (1) the clay assemblages may reflect some mechanism of differential transport and sedimentation similar to that described by Edzwald and O'Melia (1975); or (2) typical Kiowa shale may have derived the bulk of its clay minerals from sources different from those that supplied Longford clay minerals, and the lagoon or bay deposits may represent a blending of different detrital clay suites. Taking up the first possibility, Edzwald and O'Melia (1975) noted that the bottom sediments in the Pamlico River estuary, North Carolina, show a downstream increase in the proportions of illite to kaolinite. Experiments by Edzwald and O'Melia (1975) also indicate that illite should be transported farther seaward and into more saline realms than either kaolinite or smectite before it forms settling aggregates of sufficient size to bring about deposition. Kaolinite, on the other hand, tends to settle out in realms of low salinity, and smectite in realms of intermediate salinity. The separation is imperfect, however, and could lead to variations in mineral composition like those found between Longford and open-sea Kiowa sediments. Much illite could have bypassed Longford depositional realms and been differentially transported to and deposited in open-sea Kiowa realms.

Alternatively, the intermediate character of clay-mineral assemblages in the lagoon or bay shales above the capping siltstone might reflect the blending of clays from different source areas. Weaver (1958a, 1958b) noted that clay-mineral assemblages in many sedimentary settings can be explained as well by source-area controls as they can by diagenetic changes or other mechanisms. Several studies of clay minerals in marine, marginal-marine, estuarine, and fluvial settings along the Atlantic and Gulf coasts of the United States support Weaver's conclusion (Griffin, 1962; Hathaway, 1972; Pevear, 1972; among others). The same may be true for the Kiowa Formation. For example, layers of Ca-montmorillonite (Ca-bentonite) in Kiowa shale in southwestern Kansas suggest the possibility that much of the smectite in Kiowa shale was derived, either directly or indirectly, from western volcanic sources. Much of the illite in Kiowa shale could have been swept into the Kiowa sea from distant, unknown source areas. Both the illite and smectite could have been transported southward along the coast in north-central Kansas by long-shore currents, evidence for which is seen in the abundant southward dipping cross-strata in Kiowa sandstone (Fig. 1; Franks, 1966, 1975). Much of the illite and smectite in the silty, light-gray shales, therefore, may have been washed from the open sea into Kiowa lagoons (Meade, 1969), and deposited there along with the relatively abundant kaolinite and smectite supplied by Longford streams. The lack of abundant illite in most Longford clay rocks accords with the idea (Pl. 2), even though weathering penecontemporaneous with deposition of Longford alluvial sediments could have destroyed much illite.

If significant differential transport and sedimentation of illitic clays did take place in Longford fluvial and estuarine realms, some of the overbank sediments laid down by Longford streams should contain appreciable illite that was not destroyed by weathering, because floodplains or floodbasins are places where effective differential transport and sedimentation of kaolinite, smectite, and illite probably cannot operate. Many of the overbank deposits do contain illite (Pl. 2), but the amounts may not be sufficient to account for the abundance of illitic clays in Kiowa lagoon, bay, or open-sea sediments. Blending of clays from different source regions may best account for the major variations in the clay-mineral assemblages in Longford and Kiowa rocks.

# SUMMARY AND CONCLUSIONS

The Longford Member is a mappable unit and a distinctive facies of the late Albian Kiowa Formation in north-central Kansas. The top of the member is marked by a conspicuous siltstone, whereas the lower parts consist of varied assemblages of kaolinitic and smectitic clay rocks, siltstone, sandstone, and lignite. The unit is named for exposures near Longford, southwestern Clay County. Brightly colored, red-mottled clay rocks in the lower parts of the member and the nearly white siltstone that caps the member contrast with olive-gray shale and brown sandstone in facies-equivalent and overlying parts of the Kiowa Formation. Longford rocks rest with transgressive disconformity on Lower Permian rocks and are overlain conformably by higher parts of the Kiowa Formation. As the Kiowa Formation thins and pinches out northward between Permian rocks below and the Dakota Formation above, the rocks of the Longford Member also are truncated by the Dakota Formation. The northward thinning and disappearance of the Kiowa Formation and its Longford Member coincide with the northward increase in breadth of the gently dipping western limb of the Nemaha anticline. The siltstone marking the top of the member is the northernmost kind of Kiowa rock that can be identified in any practical way as the formation pinches out beneath red-mottled clay rocks and sandstone of the Dakota Formation near the Clay County-Washington County boundary, about 30 mi (48 km) south of the Kansas-Nebraska border (Figs. 1, 2; Pl. 1).

Longford rocks have been confused with rocks belonging to the Dakota Formation in central Kansas. Both units contain red-mottled clay rocks, gray to brownish-gray clay rocks, lignite, light-colored siltstone showing reed or trace fossils, and sandstone. The stratigraphic position of the Longford Member at the base of the Kiowa Formation, in combination with the characteristic sequence of rock types, allows easy recognition of the member in most places. From top to bottom, the characteristic sequence of strata is: Kiowa shale or sandstone overlies silty or shaly beds that grade sharply downward into the nearly white siltstone that marks the top of the member; the siltstone in turn grades downward into varied assemblages of clay rocks, siltstone, sandstone, and lignite; these assemblages in turn rest unconformably on Permian rocks (Pl. 2; Fig. 15). Most, but not all, lignite in the Dakota Formation is near the top of the unit. Consequently, stratigraphic position beneath the capping siltstone is the distinguishing feature of Longford lignite. The same is true for Longford sandstone, much of which resembles other sandstone in the Kiowa and Dakota formations. Longford clay rocks commonly contain appreciable smectite. As a result, their weathering characteristics differ from those shown by most Dakota clay rocks. Slopes developed on exposed Longford clay rocks tend to be puffy and to have abundant shrinkage cracks (Fig. 7), in contrast to the relatively smooth, resistant surfaces formed on kaolinitic Dakota clay rocks. The siltstone that marks the top of the Longford Member could be confused with nearly white Dakota siltstone that also contains reed or trace fossils, except that such Dakota siltstone is near the top of that formation. Where the capping Longford siltstone lacks reed or trace fossils, however, it can be mistaken for siltstone near the base of the Dakota Formation. Along the pinchout of the Kiowa Formation, therefore, where the Dakota Formation rests on and truncates the siltstone marking the top of the Longford Member, attempts to distinguish between Longford and Dakota rocks commonly become impractical. Consequently, some of the rocks mapped as part of the Dakota Formation north of the Kiowa pinchout shown in Plate 1 probably are coextensive with Longford rocks.

The distribution of Longford rocks in the subsurface west of the outcrop belt is incompletely known. Longford rocks are present in the subsurface of Ottawa County, and probably also in the subsurface of northwestern Ellsworth County (Fig. 1) (O. S. Fent, 1977, written communication). Despite the northward pinchout of the Kiowa Formation along the Nemaha anticline in Kansas (Fig. 1, Pl. 1), the depositional strike of the Kiowa Formation (about N25°W) implies that equivalents of Longford rocks may be present in the subsurface of southern and central Nebraska. Longford rocks, accordingly, may be important to the unravelling of Cretaceous stratigraphy there. Longford rocks, especially the red-mottled and brownish-gray to gray clay rocks, should not be equated with similar rocks in the Albian Cheyenne Sandstone of southern Kansas, or with similar rocks in the Albian (?) to Cenomanian Dakota Formation of central Kansas and southern Nebraska. Longford rocks are facies equivalents of Kiowa shale and sandstone in Scott's (1970a, 1970b) *Inoceramus bellvuensis* and *I. comancheanus* concurrent-range zones (Fig. 2). The stratigraphic relations, if any, between the Longford Member and widespread sandstone and sequences of nearly white to greenish-gray siltstone, shale, and sandstone at the base of the Cretaceous System in the subsurface of western Kansas need to be examined.

The thickness of the Longford Member ranges from 0 to about 100 ft (0 to 30 m). The thickness, especially of the lower parts, is controlled mainly by topography developed on underlying Permian rocks prior to deposition of the Cretaceous beds. In places, the siltstone that marks the top of the member rests directly on Permian rocks, but elsewhere the lower parts of the member are as much as 80 ft (24 m) thick. Topography developed on Permian rocks before transgression of the Kiowa sea not only influenced the kinds of Longford sediment that accumulated from place to place, but they also exerted profound control over Longford environments of deposition.

Longford sediments were deposited in fluvial, estuarine, and inner shore realms behind barrier systems that formed linear clastic Kiowa shorelines. Light-gray, commonly silty Kiowa shale overlying the siltstone that marks the top of the Longford Member is inferred to be the product of deposition of muds in bays or lagoons, whereas the nearly white, thin-bedded to thin-laminated, well-sorted siltstone is judged to represent deposits that accumulated along the landward, inner shores of the lagoons or bays (Fig. 15). Evidence that tides affected those inner shores mostly is indirect and depends on indications that tides influenced deposition of the thick, lenticular deposits of sand that formed the barrier systems. The siltstone does not show the kinds of upward fining, progradational sequences that characterize many tidal-flat deposits (Klein, 1971, 1972). The siltstone formed under conditions of general transgression, and opportunities for deposition of diagnostic, progradational sequences were limited. Current ripple marks, flaser bedding, reed fossils, plant rootlets, and probable insect trails preserved in the siltstone, however, are consistent with intertidal to supratidal deposition of much of the silt. Locally detected upward changes from siltstone containing reed fossils or plant rootlets into siltstone showing low-angle cross-strata, oscillation and current ripple marks, and rod-shaped burrows indicate submergence of the inner shore deposits as Kiowa lagoons or bays expanded or shifted landward.

Relatively scarce sandstone and abundant red-mottled clay rocks in the lower parts of the member represent fluvial deposits that accumulated in the broad valleys that were eroded into Permian bedrock before the Kiowa sea transgressed north-central Kansas. Log-probability plots of grain-size distributions in the sandstone have shapes that compare well with those of fluvial sands studied by Visher (1969). The orientation of cross-strata (vector resultant of S44°W) in one lens in Clay County (sections 16 and 21, T.9S., R.2E., Pl. 1) also accords with fluvial transportation of the sand down the Kiowa depositional slope. The scour-fill contact at the base of the lens and the incorporation of fragments of penecontemporaneously reworked clay rocks in the sandstone also are consistent with fluvial processes. Intercalation of other sandstone with red-mottled clay rocks also suggests fluvial sedimentation of the sand.

Red-mottled clay rocks and less common red-mottled siltstone in the lower parts of the member stem from overbank deposition of clay and silt by Longford streams. Slickensided fractures, microscopic evidence of clay skins, and other evidence of soil-matrix fabrics (Brewer, 1964) support the idea. The red, hematitic mottles are attributed to early diagenetic reduction, redistribution, re-oxidation, and dehydration of iron oxides in the floodplain sediments, which may have been stained red or brown initially. The mottled rocks are products of gleying and formation of so-called soft plinthite in low-lying alluvial soils in which fluctuating groundwater tables lay close to the floodplain surfaces. The red-mottled rocks resemble groundwater laterites, but they have more in common with immature alluvial soils that have not undergone extensive weathering. Like immature alluvial soils along the Amazon River (Sombroek, 1966), the red-mottled Longford rocks do not show evidence of well-developed soil profiles, even though they are extensively mottled; they lack layers of concretionary hard plinthite near the tops of the mottled lenses; although they contain abundant kaolinite, the red-mottled rocks also contain appreciable 2:1 clay minerals, chiefly smectite and lesser amounts of illite; and they do not contain gibbsite. The red-mottled rocks seem to be products of initial soil-forming processes or "soil ripening" on floodplains (Pons and Zonneveld, 1965).

Other Longford sediments may have accumulated in fluvial or estuarine settings. Lenses and seams of lignite are composed of detrital plant debris that may have been deposited in more than one environmental realm. Those low in the member probably formed as floodbasin deposits, whereas those near the base of the siltstone that caps the member could have formed either as floodbasin deposits or on the plains of estuary-head deltas. The same is true for lenses of nearly black, highly plastic, smectitic claystone. The claystone shows slickensided fractures, churned fabrics, and downward projections into underlying clay rocks. It likely formed as a tropical vertisol (Soil Survey Staff, 1960), a common soil in floodbasins or estuarine settings where there is repeated wetting and drying of the soils. The abundant gray and brownish-gray mudstone and claystone, much of it containing appreciable carbonaceous matter, likewise may have formed from deposits that accumulated in fluvial or estuarine settings. Because the Kiowa sea flooded the gentle topography and broad valleys developed on Permian bedrock, products of estuarine sedimentation are to be expected among the rocks underlying, or not far below, the siltstone marking the top of the Longford Member (Fig. 15).

Much of the siltstone below that marking the top of the member probably stems from estuarine sedimentation, as is indicated by its position between fluvial deposits below and inner shore deposits above. Many of the siltstone beds show oscillatory and current-ripple structures and contain abundant carbonaceous plant debris or imprints of plant debris, features that accord with estuarine deposition of silts, perhaps near or on bayhead deltas.

The climate that prevailed during deposition of Longford sediments was equable, humid, and tropical. The lack of marked seasonal banding in the piece of silicified gymnosperm wood that was found near the base of the member in Marion County indicates the equable nature of the climate. The similarity of the red-mottled clay rocks and the dark, highly plastic, smectitic claystone to modern, alluvial tropical soils is suggestive of the humidity and warmth of the climate. These inferences reinforce interpretations about Kiowa climates by Franks (1966, 1975) and Scott (1970a, 1970b).

Heavy minerals noted in thin sections of Longford sandstone also are found in other Kiowa sandstone (Franks, 1966, 1975). The tourmaline and zircon in the heavy-mineral suites indicate derivation of the sands by reworking of Paleozoic sedimentary materials in the continental interior to the east of Kansas. Staurolite in the sandstone, however, suggests that some of the sand was derived from the crystalline terrains of the central Appalachian Mountains, either directly, or indirectly by reworking of Upper Permian or younger rocks now eroded from the continental interior (Franks, 1966, 1975).

The abundant kaolinite and smectite in Longford clay rocks probably were derived by reworking of fossil soils similar to those preserved locally along the Permian-Cretaceous unconformity in central Kansas. Similar soils probably were widespread on Paleozoic rocks in the continental interior during Early Cretaceous time. Those in Kansas likely developed under conditions of tropical weathering and poor drainage of the illitic and chloritic parent rocks. The small amounts of illite and mixed-layer chlorite or vermiculite in Longford rocks are consistent with derivation of those minerals by erosion of weathered Paleozoic source rocks in the continental interior.

Irregular upward increases in the proportions of kaolinite to smectite in Longford clay-mineral assemblages may depend on upward increases in the abundance of siltstone in the member. Post-depositional alteration of smectite or illite to kaolinite in the more permeable silty beds may be the chief factor (Glass, 1958; Glass and others, 1956; Potter and Glass, 1958; Wilson and Pittman, 1977), but the data are inconclusive. Alternatively, the estuarine environments in which much of the silt was deposited may have favored the formation of relatively coarse kaolinite settling aggregates, thus leading to differential transport of smectite and illite to more saline realms (Edzwald and O'Melia, 1975).

Lateral variations in the relative amounts of illite, smectite, and kaolinite in Longford and facies-equivalent Kiowa rocks generally correspond to the observations summarized by Parham (1966). Illite tends to be most abundant in sediments deposited in offshore realms, whereas kaolinite is most abundant in sediments deposited in nonmarine realms. Smectite (montmorillonite) may show distributions similar to those of illite, or it may be most abundant in nearshore or nonmarine sediments. The lateral variations in clay-mineral assemblages in Longford and Kiowa rocks can be interpreted as reflecting different source areas for much of the illite and smectite in Kiowa shale and much of the kaolinite and smectite in Longford rocks (Weaver, 1968a, 1968b), or as a response to differential transport and sedimentation of the clay minerals in fluvial, estuarine, lagoon, and marine environments (Edzwald and O'Melia, 1975). Definitive data are not at hand, but the small amount of illite in most Longford fluvial overbank deposits implies that differences in source areas may best account for the lateral variations. Floodplains and floodbasins would not be places where differential settling tendencies of differently sized clay aggregates could bring about effective sorting and separation of clay-mineral species.

The apparently scattered distribution of Kiowa shale and thick, lenticular deposits of Kiowa sandstone above the Longford Member affords insight into the mechanisms by which the Kiowa sea flooded onto the gentle topography developed on the western flank of the Nemaha anticline in north-central Kansas. Transgression must have been accompanied by in-place growth and eventual submergence of Kiowa barrier systems. Submergence resulted in step-wise, landward shifting of surf zones (Sanders and Kumar,



1975) without effective ravinement (shore-face and tidal-inlet erosion) of barrier-island, lagoon or bay, estuarine, and fluvial deposits (Fig. 16-C). The landward shift of surf zones resulted in construction of new barriers close to the former inner shores of lagoons or bays, and in renewal of the process. Submergence of Kiowa barrier systems and landward shifting of surf zones readily accounts for the sequences of open-sea Kiowa shale that overlie lagoonal shales and the capping Longford siltstone in many places, and for the presence of thick, lenticular deposits of Kiowa barrier sandstone above lagoon or bay and inner shore deposits in other places. It also accounts for the excellent preservation of thick sections of Longford alluvial and estuarine deposits along valleys eroded into Permian bedrock.

## SUGGESTIONS FOR FURTHER STUDY

This study of Longford rocks indicates a number of projects that could add to the understanding of Cheyenne, Kiowa, and Dakota stratigraphy and sedimentation. The subsurface distribution of Longford rocks north and west of the outcrop belt is of especial interest, as would be more accurate determination of the facies relationships between Kiowa shale and sandstone and Longford rocks. Similarly, the stratigraphic relations between Longford rocks and strata that have been called Cheyenne Sandstone in the subsurface of western Kansas warrant careful analysis. Detailed petrologic and stratigraphic studies of thick, lenticular deposits of Kiowa sandstone should further understanding of the factors that controlled the development of Kiowa barrier systems. Conceivably, increased knowledge of the sandstone could lead to exploitation of similar sandstone in the subsurface of western Kansas and southern Nebraska as traps for oil and gas.

Detailed study of clay-mineral assemblages in Kiowa and Longford rocks might permit assessment of the extent to which soil-forming processes controlled Longford clay-mineral assemblages, and the extent to which differences in Longford and Kiowa assemblages reflect differential transport and sedimentation or differences in source materials. Detailed study of plant fossils in Longford and facies-equivalent Kiowa rocks might aid interpretations of Longford and Kiowa depositional environments, as well as contribute to the growing body of knowledge on angiosperm evolution.

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# APPENDIX A

## WORK METHODS AND TERMINOLOGY

Field work pertinent to this report was done as a part of a larger study of Kiowa and Dakota rocks (Franks, 1966, 1975) in 1962, 1963, and 1965. Additional field work was done in 1975 and 1976. Reconnaissance mapping of the Permian-Cretaceous boundary and formational contacts shown in Plate 1 was done in the field using vertical aerial photographs. The upper contact of the Longford Member is based partly on field mapping, but it also was inferred from field notes, aerial photographs, and topographic maps. Accordingly, that contact is approximate at best, but it is judged to be reasonably accurate at the scale of the map. Surficial deposits shown in Plate 1 were mapped only locally, and their boundaries are adapted mainly from the work of Walters and Bayne (1959). The bedrock-surficial deposit contacts are intended to indicate those areas where abundant Quaternary cover generally masks bedrock.

Clay-rock samples taken for X-ray diffraction studies were dispersed in distilled water, and Stokes-law settling was used to obtain the fraction finer than two microns. Smear mounts on glass petrographic slides were prepared using the method of Gibbs (1965). X-ray diffraction traces were made using a proportional counting diffractometer and nickel-filtered copper radiation at a scan rate of $1°2\theta$ per minute. One-degree and one-half degree slit systems generally were employed, and the instrument was operated at 35 KV and 18 ma. The terms "illite" and "kaolinite" are used here as generally accepted. Any 14-Angstrom component that expanded on treatment with glycerol to give a diffraction of 18 A, and collapsed to about 10 A on heating to 450°C, was classed as smectite. The name "chlorite" refers to any 14-Angstrom component whose basal spacing did not expand on glyceration and did not collapse below 13.8 A on heating at 575°C for one-half hour. The name "vermiculite" refers to 14-Angstrom components that showed variable expansion toward 18 A on glyceration and variable collapse below 13.8 A on heat treatment. Randomly interstratified mixed-layer clays whose diffraction maxima showed as skew shoulders on the low-angle side of the illite 001 diffraction, or as humps between that diffraction and 14 A, generally had illite assigned as one component. Depending on the degree of expansion and collapse, other components, such as smectite, chlorite, or vermiculite, were assigned to the mixed-layer structures.

Size analyses of sandstone samples were made after disaggregating the samples using a ceramic mortar and a rubber-tipped pestle. One-hundred gram samples were sieved for a total of 30 minutes on a Ro-Tap, using standard screens sized at 1/4-Phi intervals.

# APPENDIX B

## DESCRIPTIONS OF MEASURED SECTIONS

1. Type section for Longford Member, Kiowa Formation, measured from creek bottom about 0.25 mile south cen. W line sec. 9, T.10S., R.1E. (about 200 ft south of bridge) to top of roadcut about 0.15 mile north SW cor. sec. 9, T.10S., R1E., Clay County, Kansas. Measured by Paul C. Franks.

| Longford Member, Kiowa Formation: | Thickness (feet) | (meters) |
|---|---|---|
| Top of hill. | | |
| 19. Siltstone, very light brownish-gray; weathers pinkish gray to pale yellowish orange. Thin wavy laminae and even laminae weathering to sets 0.1 to 0.5 ft thick. Contains sparse mica flakes; sparse limonitic stain; sparse pyrite nodules. Abundant concretionary masses of calcite cement as much as 5 ft thick and 10 ft long near top; concretions stand out in relief and hold abundant disseminated pyrite. Abundant carbon as flecks and films. Grades sharply into next below ...... Exposed, | 8.8 | 2.67 |
| 18. Siltstone, light-gray to light brownish-gray; weathers very pale orange to grayish orange. Indistinct thin wavy and thin even laminae; abundant carbon flecks, films, and fragments on bedding surfaces; sparse mica flakes; carbonized wood commonly replaced by pyrite; argillaceous. Grades into next below | 3.7 | 1.13 |
| 17. Siltstone, pale grayish-orange. Bedding largely masked by limonitic stain; chalky texture, but hard and weathers to form prominent ledge. Irregularly distributed calcite cement; basal bed 0.1 to 0.2 ft thick and cemented by calcite stands out in relief | 4.5 | 1.37 |
| 16. Siltstone, very light gray; sparse grayish-orange "limonite" stain along bedding surfaces; weathers yellowish gray to very pale orange. Thin indistinct wavy laminae; weathers to beds 0.1 to 1. ft thick. Hard, but does not stand out in relief. Sparse interstitial clay. Scour-fill contact with next below | 3.1 | 0.95 |
| 15. Shale, medium-gray to medium light-gray; sparse limonitic stain; weathers very light gray to yellowish gray. Kaolinitic but contains some smectite. Thin-laminated to laminated; poor fissility; silty. Thickens southward into carbonaceous, silty mudstone. Grades sharply into next below | 0.7 | 0.21 |
| 14. Siltstone, light-gray; weathers yellowish gray. Bedding indistinct; abundant interstitial clay, sparse mica flakes ...... Exposed, | 2.9 | 0.88 |
| Covered interval | 7.9 | 2.40 |
| 13. Siltstone, light-gray to very light gray; weathers very light gray to yellowish orange and yellowish gray. Thin-laminated and ripple laminated; bedding inclined 5 or 6° north. Variable amounts of interstitial clay; where interstitial clay is abundant, contains contorted pods of less argillaceous siltstone measuring 1 cm thick and up to 3 cm long. Carbon as leaf and stem imprints on bedding planes locally abundant; local calcareous cement forms discoidal concretions as much as 0.2 ft thick; concretions contain scattered pyrite nodules | 14.5 | 4.40 |
| 12. Sandstone and siltstone, light-gray to very light gray; abundant yellowish-orange limonitic stain. Very fine grained to silt-sized; friable; sparse interstitial clay; sparse grains of pink quartzite and flakes of mica. No obvious bedding. Grades sharply into next below | 1.3 | 0.40 |
| 11. Sandstone, very light gray; mainly weathered dark yellowish orange to light brown and dusky brown; abundant limonitic and manganese-oxide stain. Fine- to coarse-grained; mainly medium-grained; thin-bedded to wavy laminated; contains sparse medium gray mudstone seams and pellets; sparse interstitial clay; abundant pyrite nodules; friable. Scour-fill contact with next below | 4.0 | 1.22 |
| 10. Mudstone, light-gray with brownish overtones grading down to light brownish-gray. Dominantly kaolinitic but contains some smectite and sparse chloritic or vermiculitic mixed-layer clay. Abundant limonitic stain on fracture surfaces in siltiest parts; conchoidal fracture in upper parts; blocky fracture in lower parts. Sparse light-gray siltstone laminae near base. Bedding mainly indistinct. Grades into next below | 4.6 | 1.40 |
| 9. Mudstone, very light gray to light yellowish-gray; abundant purplish-red to moderate reddish-brown mottles. Dominantly kaolinitic but contains sparse illite, smectite, and chloritic or vermiculitic mixed-layer clay. Plastic; blocky fracture. Grades sharply into next below | 4.8 | 1.46 |
| 8. Siltstone, light-gray to very light gray with brown overtones. Bedding not obvious; locally calcareous; carbon fragments towards base; abundant limonitic stain. Grades into next below | 1.7 | 0.52 |
| 7. Mudstone, light-gray with brown overtones and moderate reddish-brown to reddish-brown mottles. Largely kaolinitic but contains smectite and perhaps mixed-layer vermiculite-smectite. Less silty toward base; abundant carbon fragments in basal 1 to 2 ft. Grades sharply into next below | 11.1 | 3.36 |
| 6. Lignitic shale, pale brown. Thin wavy laminae; abundant carbon as flecks, films, and carbonized wood fragments; abundant jarosite stain. Grades laterally into mudstone containing isolated fragments of carbonized wood. Grades sharply into next below | 0.5 | 0.15 |
| 5. Siltstone grading down to mudstone. Light-gray grading down to light-gray with reddish-brown mottles. Clay fraction composed | | |

| | Thickness (feet) | (meters) |
|---|---|---|
| mainly of smectite and illite. Abundant grayish-yellow to yellowish-orange limonitic stain in upper 1 ft. Blocky to conchoidal fracture | 12.2 | 3.71 |
| 4. Mudstone, light-gray grading down to light brownish-gray. Clay fraction composed mainly of illite and smectite but contains sparse kaolinite and perhaps chlorite or vermiculite. Very silty towards base; bedding indistinct; blocky fracture; abundant carbon in basal 3 ft as flecks, films, fragments, and imprints of stems and leaves; abundant jarosite stain on randomly oriented fractures | 8.8 | 2.67 |
| 3. Mudstone, medium-gray with brown overtones and sparse moderate-red mottles; abundant "limonite" and jarosite stain along fracture surfaces. Clay fraction composed almost completely of smectite but contains sparse illite and vermiculite or chlorite. Nonplastic, blocky fracture; sparse carbon flecks, films, and imprints of plant debris. Base Longford Member, Kiowa Formation | 3.0 | 0.91 |
| Total thickness Longford Member, Kiowa Formation, measured | 98.1 | 29.91 |
| Unconformity. | | |
| Wellington Formation: | | |
| 2. Mudstone, reddish-brown. Thin-laminated to laminated but poor fissility; blocky fracture. Clay fraction composed mainly of illite but contains appreciable smectite and chlorite. Scattered laminae bleached yellowish-gray in top 0.5 ft. Abundant limonitic or hematitic cement in top 0.5 ft. Grades sharply into next below | 3.4 | 1.03 |
| 1. Mudstone, light-gray to moderate greenish-gray; weathers dusky yellow. Thin-laminated but poor fissility; blocky fracture. Clay fraction composed mainly of illite but contains appreciable smectite and chlorite. Base covered ... Exposed, | 1.5 | 0.46 |
| Total thickness Wellington Formation measured | 4.9 | 1.49 |

Note: The Permian-Cretaceous contact and the base of the Longford Member, Kiowa Formation, also are exposed in a gully near cen. W ½ NW ¼ sec. 16, T.10S., R.1E. where the Permian-Cretaceous contact is about 18 ft higher than in the section described above. The contact corresponds approximately to the middle of unit 5. The topmost Permian mudstone is intensely weathered and variegated. It contains abundant kaolinite, illite, and smectite as well as minor amounts of chloritic or vermiculitic mixed-layer clay, and is overlain by a carbonaceous mudstone and lignitic sequence as well as by gray mudstone showing abundant red mottles similar to material described under unit 7 above. Sparse chert and quartzite pebbles weather out of the basal carbonaceous Longford mudstone.

## 2. Section measured from Longford Member, Kiowa Formation, into overlying Kiowa rocks near cen. W ½ sec. 32, T.9S., R.1E., Clay County, Kansas. Section complements type section of Longford Member in sec. 9, T.10S., R.1E., Clay County, Kansas. Measured by Paul C. Franks.

| | Thickness (feet) | (meters) |
|---|---|---|
| Kiowa Formation: | | |
| Covered. | | |
| 4. Shale, light-gray to light olive-gray; weathers light gray to moderate olive brown. Composed mainly of smectite but contains abundant illite and kaolinite. Thin-laminated, plastic. Sparse jarosite stain; sparse layers of discoidal concretions of impure siderite. Grades sharply into next below. Top covered ... Exposed, | 4.0 | 1.21 |
| 3. Shale, very light gray. Dominantly kaolinitic but contains abundant illite and smectite. Silty, plastic; laminated to thin-laminated, poor fissility; sparse limonitic stain along lamination. Grades sharply and irregularly into next below | 1.1 | 0.34 |
| Total thickness upper part Kiowa Formation measured | 5.1 | 1.55 |
| Longford Member, Kiowa Formation: | | |
| 2. Siltstone, very light gray with brown overtones; weathers yellowish gray to white. Sparse "limonite" stain; hard; ripple and even thin laminae; micro-cross-lamination in sets as much as 0.2 ft thick. Local calcite cement forms concretions as much as 3 ft in diameter; abundant limonitic cement in top 0.1 ft. Symmetric transverse ripple marks with wave lengths as great as 0.2 ft common on bedding surfaces. Grades irregularly into next below | 5.2 | 1.58 |
| 1. Siltstone, very light brownish-gray with brownish-gray carbonaceous thin laminae; weathers light grayish orange. Laminated to thin-laminated; ripple and wavy laminae. Sparse "limonite" stain; argillaceous ... Exposed, | 2.0 | 0.61 |
| Total thickness Longford Member, Kiowa Formation, measured | 7.2 | 2.19 |
| Total thickness Kiowa Formation, measured | 12.3 | 3.74 |

## 3. Section of Longford Member, Kiowa Formation, measured along gully near cen. SE ¼ sec. 23, T.16S., R.1E., Dickinson County, Kansas. Measured by Paul C. Franks.

| | Thickness (feet) | (meters) |
|---|---|---|
| Kiowa Formation: | | |
| Covered. | | |
| 10. Shale, light olive-gray to dusky-yellow. Highly weathered. Clay fraction composed largely of smectite and illite, but containing abundant kaolinite. Plastic; thin-laminated; scattered concretions of impure siderite less than 0.2 ft thick strung out along bedding ... Exposed, | 2.3 | 0.70 |
| 9. Shale, very light gray; abundant reddish-brown to dark yellowish-orange stain; weathers light brownish gray. Clay fraction composed mainly of illite and "degraded" illite, but contains abundant kaolinite. Plastic; thin-laminated; weathers to puffy slope littered with abundant gypsum needles. Contains abundant radial aggregates of gyp- | | |

sum as much as 0.2 ft in diameter; shaly lamination contorted about gypsum aggregates; aggregates associated with abundant "limonite" stain. Grades into next below ..... 1.8 0.55

8. Shale, light-gray to yellowish-gray; weathers very light gray. Composed largely of illite and smectite, but contains abundant kaolinite. Plastic; thin-laminated; sparse silty laminae; weathers to puffy slope. Pyritic nodules associated with abundant "limonite" stain and partly altered to fine-grained soft aggregates of gypsum. Grades into next below ..... 4.7 1.44

7. Shale, medium dark-gray; weathers medium light gray. Dominantly kaolinitic, but contains appreciable illite, smectite, and a vermiculite component. Plastic; laminated; blocky fracture; silty in basal 0.2 ft. Grades into next below ..... 1.0 0.30

Total thickness upper part Kiowa Formation measured ..... 9.8 2.99

Kiowa Formation, Longford Member:

6. Siltstone, pinkish-gray to dark pinkish-gray; weathers grayish orange. Thin-laminated to thin-bedded; hard; forms resistant ledge. Bedding surfaces show abundant symmetric and asymmetric transverse ripple marks with wavelengths from 0.1 to 0.3 ft and striking mainly E-W; micro-cross-stratification; sparse interference and linguoid current ripple marks; ripple marks mainly in upper 1 ft. Sparse nearly vertical burrows as much as 2 cm in diameter and 14 cm long. Micaceous; sparse black opaque grains; abundant iron-oxide cement in top 0.5 ft. Grades unevenly into next below; thickness measured ..... 4.7 1.43

5. Siltstone and mudstone, light brownish-gray grading down to medium-gray; weathers dark pinkish gray to yellowish orange. Clay fraction composed largely of kaolinite, but contains abundant illite and smectite. Thin-bedded grading down to thin-laminated. Abundant "limonite" stain; sparse carbon as flecks and fragments ..... 5.2 1.59

4. Claystone, medium dark-gray to dark-gray with brown overtones grading down to medium-gray to medium light-gray containing patches of dark-gray. Composed largely of kaolinite, but contains abundant smectitic and chloritic to vermiculitic mixed-layer clay. Plastic, nonfissile; weathers to puffy medium-gray slope. Upper foot very nearly lignitic, largely kaolinitic. Sparse pyrite or marcasite nodules; abundant hematite stain near top. Grades into next below ..... 9.4 2.86

3. Claystone and mudstone, light-gray to medium light-gray and brownish-gray; dark reddish-brown to dusky-red mottles; mottling most prominent near base. Composed largely of smectite and chlorite or vermiculite and interstratified smectite-chlorite mixed-layer clay, but contains appreciable kaolinite; basal parts composed almost exclusively of mixed-layer clay containing kaolinite and smectite. Plastic; no obvious lamination; weathers to puffy light-gray slope with reddish-brown stain. Near base locally contains contorted pods and fragments reworked from variegated claystone at top of Permian. Base Longford Member, Kiowa Formation ..... 5.4 1.65

Total thickness Longford Member, Kiowa Formation, measured ..... 24.7 7.53

Total thickness Kiowa Formation measured ..... 34.5 10.52

Unconformity.

Wellington Formation:

2. Claystone, variegated, mainly grayish-red but streaked, stained, and mottled moderate yellowish-brown, greenish-yellow, dusky-red, white, grayish-purple, and pale yellowish-gray. Top 0.2 ft (6 cm) generally stained dusky red by iron oxide. Composed almost completely of kaolinite; contains white nodules composed of kaolinite and halloysite. Weathers to puffy reddish-purple slope with abundant white blotches; silty near base; waxy conchoidal fractures in upper parts where nonsilty. Paleosol at top of Permian. Grades irregularly into next below. Thickness variable but approximates ..... 3.0 0.91

1. Mudstone, greenish-gray to grayish-yellow; abundant dark yellowish-orange to light olive-brown stain. Composed largely of illite and smectite, but contains appreciable chlorite and kaolinite. Silty near base where laminated and intercalated with limestone next below; blocky fracture. Thickness variable but approximates ..... 1.5 0.46

Carlton (?) Limestone Member, Wellington Formation. Platy, dolomitic limestone. Not measured.

Total thickness Wellington Formation measured ..... 4.5 1.37

# Index

Plate 1.—Reconnaissance geologic map showing distribution of Longford Member, Kiowa Formation, pinchout of Kiowa Formation, and locations of type and reference sections for Longford rocks in part of north-central Kansas.

Plate 2.—Graphic measured sections of Longford Member, Kiowa Formation, including representative X-ray diffraction traces of clay-mineral assemblages (air-dried smear mounts of fraction finer than 2 $\mu$).

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