Earthquake-Induced Liquefaction Features in the Coastal Setting of South Carolina and in the Fluvial Setting of the New Madrid Seismic Zone

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By S.F. Obermeier,[1] R.B. Jacobson,[1] J.P. Smoot,[1] R.E. Weems,[1] G.S. Gohn,[1] J.E. Monroe,[2] and D.S. Powars[1]

ABSTRACT

Many types of liquefaction-related features (sand blows, fissures, lateral spreads, dikes, and sills) have been induced by earthquakes in coastal South Carolina and in the New Madrid seismic zone in the Central United States. In addition, abundant features of unknown and nonseismic origin are present. Geologic criteria for interpreting an earthquake origin in these areas are illustrated in practical applications; these criteria can be used to determine the origin of liquefaction features in many other geographic and geologic settings.

In both coastal South Carolina and the New Madrid seismic zone, the earthquake-induced liquefaction features generally originated in clean sand deposits that contain no or few intercalated silt- or clay-rich strata. The local geologic setting is a major influence on both development and surface expression of sand blows. Major factors controlling sand-blow formation include the thickness and physical properties of the deposits above the source sands, and these relationships are illustrated by comparing sand blows found in coastal South Carolina (in marine deposits) with sand blows found in the New Madrid seismic zone (in fluvial deposits). In coastal South Carolina, the surface stratum is typically a thin (about 1 m) soil that is weakly cemented with humate, and the sand blows are expressed as craters surrounded by a thin sheet of sand; in the New Madrid seismic zone the surface stratum generally is a clay-rich deposit ranging in thickness from 2 to 10 m, in which case sand blows characteristically are expressed as sand mounded above the original ground surface.

Recognition of the various features described in this paper, and identification of the most probable origin for each, provides a set of important tools for understanding paleoseismicity in areas such as the Central and Eastern United States where faults are not exposed for study and strong seismic activity is infrequent.

INTRODUCTION

The near-surface and surface expressions of earthquake-induced liquefaction are highly dependent on the local geologic setting. This paper discusses the types of features, primarily sand blows, that formed on level or nearly level ground in two very different geologic settings, one in which the parent sediments have a marine or near-marine origin (coastal South Carolina) and one in which the parent sediments have a fluvial origin (New Madrid seismic zone). The diverse types of sand blows and the equally diverse controls on their formation became apparent during field studies recently conducted to locate regions subjected to strong earthquake shaking, as indicated by the presence of liquefaction-induced features. In coastal South Carolina, field studies focused on sand blows of very late Pleistocene to Holocene age, as well as features caused by the 1886 Charleston earthquake. In the New Madrid seismic zone, the search was largely restricted to sand blows caused by the 1811–12 earthquakes. In both areas, features of nonearthquake origin occur that might be confused with earthquake-induced features. Geologic criteria have been developed by which the earthquake-induced liquefaction features can usually be distinguished. ("Liquefaction" in this paper is defined as transformation of water-saturated sediments from the solid to liquid state as a consequence of increased pore pressure; this transformation indicates that grains are suspended by pore water. "Fluidization" is defined as the process of both suspension and transport of grains by pore water.)

The scope of the paper includes both intruded features (for example, dikes and sills) and vented features (for example, sand blows). The term "sand blows" is used to indicate features formed where earthquake shaking causes liquefaction at depth followed by the venting of the liquefied sand and water to the surface. Features described as sand boils in this paper form in the absence of earthquake shaking and involve transport of sediment to the surface by artesian flow (springs). This distinction in terms is made because the processes within the ground that lead to development of earthquake-induced features differ from processes not related to earthquakes. Sand blows are thought to often form in response to liquefaction that is in turn followed by segregation of sand and

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water at depth, upward development of a vent, and finally the violent discharge of soil and water that soon diminishes to an ebbing flow (Scott and Zuckerman, 1973). (A slightly different model for some sand deposits is proposed in this paper.) In contrast, sand boils form by the downward development of a vent that typically does not enlarge violently and that continues to flow as long as high artesian pressure remains, commonly for days or years. This difference in development processes between sand boils and sand blows can often be distinguished in the field by use of techniques discussed in this paper.

COASTAL SOUTH CAROLINA

The strongest historic earthquake in the Southeastern United States took place in 1886 near Charleston, S.C. Throughout much of the epicentral region, an area about 35 km wide and 50 km long, the Modified Mercalli intensity ranged from IX to X (Bollinger, 1977). The estimated body-wave magnitude (m_b) was between 6.6 and 7.1 (Nuttli, 1983a). The potential for a future earthquake having the strength of the 1886 earthquake is a major concern in engineering design in the Southeast. The concern is reinforced by a 300-year historical record of continuing weak seismic activity near Charleston. The source of the earthquakes in the Charleston area remains unknown, and seismotectonic hypotheses are widely disparate, despite many geologic, geophysical, and seismic studies during the past decade. No faults or fault systems have been identified that adequately explain the large 1886 Charleston earthquake or the other smaller, historic earthquakes that have occurred throughout much of South Carolina (Hays and Gori, 1983; Dewey, 1985; Science News, 1986). Because direct evidence of seismotectonic conditions is lacking and because the historic earthquake record is too limited to provide a dependable basis for estimating the frequency of moderate to strong earthquakes, we undertook a search for pre-1886 sand blows.

Results of the search are shown on figure 1. Figure 1 shows the approximate boundary of the 1886 Charleston earthquake meizoseismal zone; the sites conspicuous in 1886 for development of many sand blows, described as "craterlets" by Dutton (1889); and the sites of pre-1886 sand blows that we discovered. The unshaded portion of figure 1 encompasses the area that was searched for sand blows. Radiocarbon ages of sand-blow materials at site HW show that at least three pre-1886 Holocene earthquakes have produced sand blows in that area (Weems and others, 1986). It is not yet known how many earthquakes are represented by the sand blows at the widely scattered sites on figure 1, nor are the seismic source zones known for these prehistoric sand blows. What is known is that the prehistoric sand blows extend far beyond the limits of 1886 sand blows (see fig. 1) in

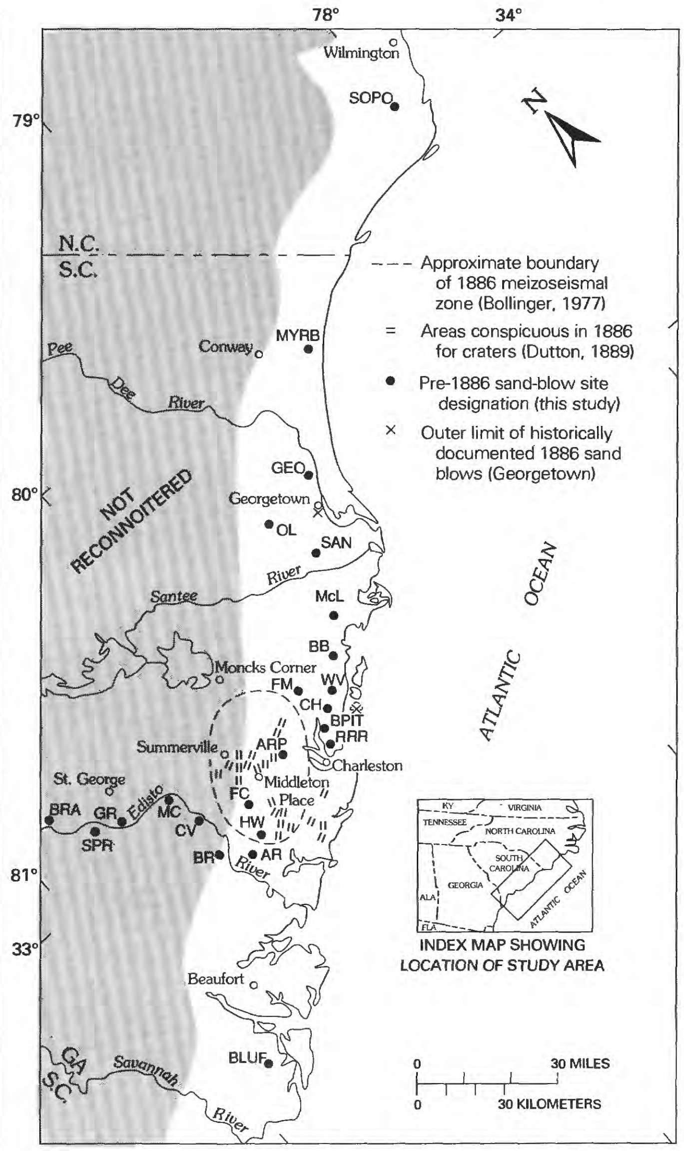

FIGURE 1.—1886 and older sand-blow sites. Unshaded onshore region is predominantly of marine sediments younger than about 240,000 years. Shading pattern denotes region of older marine sediments that was not reconnoitered. Younger fluvial sediments occur locally. All sand-blow sites in the region with no shading are in marine-related deposits, and all sites in region with shading are in fluvial deposits. Numerous sites discovered in and near the 1886 meizoseismal zone are not shown because of lack of space. Abbreviations adjacent to sand-blow sites are specific site designations.

sediments having approximately the same liquefaction susceptibility (Obermeier and others, 1986; unpublished data) and that the strongest Holocene shaking has probably been near Charleston (Obermeier and others, 1989).

All the pre-1886 sand blows that we found have no expression on the ground surface that is discernible by onsite examination or on aerial photographs. The sandblows are seen only where exposed in walls of excavations at least 1.5 m deep, typically in drainage ditches and borrow pits. At most sites shown on figure 1, at least three or four sand blows are exposed within a few hundred meters of one another. The following section focuses on the geologic setting in which these sand blows originated.

REGIONAL SETTING—GEOLOGY AND LIQUEFACTION SUSCEPTIBILITY

In South Carolina, the coastal region is known locally as the "low country" because it has low local relief (1–3 m) and low elevation (0–30 m) and because vast expanses of swamp and marshland are under water much of the year. Most of the Carolina low country is covered by a 5- to 10-m-thick blanket of unconsolidated Quaternary marine and fluvial deposits, which lies on semilithified Tertiary sediments (McCartan and others, 1984). The Quaternary sediments primarily occur as a series of six well-defined, temporally discrete, interglacial beaches and associated back barrier and shelf deposits that form belts subparallel to the present shoreline. The oldest beach deposits are farthest inland and are at the highest altitudes; younger beach deposits are progressively closer to the ocean and are at successively lower altitudes. Most beach deposits are 8 to 15 m thick.

Cutting across these marine and marginal-marine deposits at nearly right angles are four major rivers, the Savannah, Edisto, Santee, and Pee Dee (fig. 1). Bordering these rivers are fluvial terrace sediments of rather limited extent that consist almost exclusively of clean sand (that is, sand without clay, silt, or gravel).

Figure 1 shows the approximate areal extent of the marine-related deposits (beach, shelf, and open-sound back barrier) designated as Q1, Q2, and Q3 by McCartan and others (1984). The part of figure 1 containing units Q1, Q2, and Q3 is shown without shading. Q3 deposits are about 200,000 to 240,000 years old (Szabo, 1985) and are present as far as 20 to 40 km inland from the modern coast. The intervening Q2 deposits are about 80,000 to 130,000 years old (Szabo, 1985). Unit Q1 is closest to the ocean and is made of younger deposits. The search for sand blows was generally restricted to units Q2 and Q3. Older units have such a low susceptibility to liquefaction (due to effects of chemical weathering) that the likelihood of forming sand blows has been extremely low during the late Pleistocene and Holocene. Because unit Q1 generally has such a high ground-water table, the possibility of finding exposed sand blows was quite limited.

Formation of sand blows in any geologic setting depends primarily on the depth to the water table, the properties and thickness of materials in the depth range susceptible to liquefaction during shaking, and the thickness and characteristics of sediments above the zone that was liquefied during shaking. Specific relations between liquefaction susceptibility and subsequent formation of sand blows in the South Carolina low country are as follows:

(1) A water table very close to the ground surface usually greatly increases the susceptibility to liquefaction, even in comparison with a water table at depths of only 3 to 4 m. The modern water-table depth is generally 0 to 1.5 m throughout the low country and is typically a subdued mimic of the surface topography. The water table is deepest under hilltops and may come to the surface in swales. As surface elevations decrease toward the ocean, the water table is generally nearer the surface; within 15 km of the ocean, the water table is rarely deeper than 1.5 to 2 m.

(2) Clean sands are generally the only materials observed to have liquefied in the Charleston region. At one site (site SAN, fig. 1) the source sand bed contains as much as 3 to 5 percent clay; typically there is less than 1 percent silt and clay at all other sites. The liquefied sands are generally fine-grained, well-sorted (that is, uniformly graded) beach sands.

Principal properties of sand that control liquefaction susceptibility during shaking are degree of compaction or state of compactness (known as "relative density" by geotechnical engineers), sand-grain size and sorting, and cementation of the sand at grain-to-grain contacts.

The state of compactness is commonly a reflection of the energy (or mode) of deposition. To illustrate, beach deposits laid down in a high-energy environment such as a pounding surf zone are generally more dense than sand deposited in a quiet zone away from the influence of the surf. Higher density makes the surf-zone deposits more difficult to liquefy. Locally within the surf zone, though, there are many regions of low-energy deposition where the sand is not so compacted by wave pounding. In addition, for a given relative density, the fine-grained, well-sorted sands of ancient and modern beaches throughout the low country are much more susceptible to liquefaction than standard sands used for engineering analysis (Cullen, 1985), and this increased susceptibility due to grain-size effects and lower local compactness must account in part for the widespread liquefaction during the 1886 earthquake.

Cementation in near-surface subaerial environments is chiefly a reflection of the age of the deposit. Even slight cementation by agents such as silica, calcite, or clay dramatically reduces liquefaction susceptibility; thus liquefaction potential is related to age of deposits in many geologic settings (Youd and Perkins, 1978), and this relation is particularly apparent in the low country. Table 1 lists deposits in the low country shown in figure 1 in terms of depositional environment (type of deposit) and age and provides an estimate of their relative susceptibilities to liquefaction during strong seismic shaking.

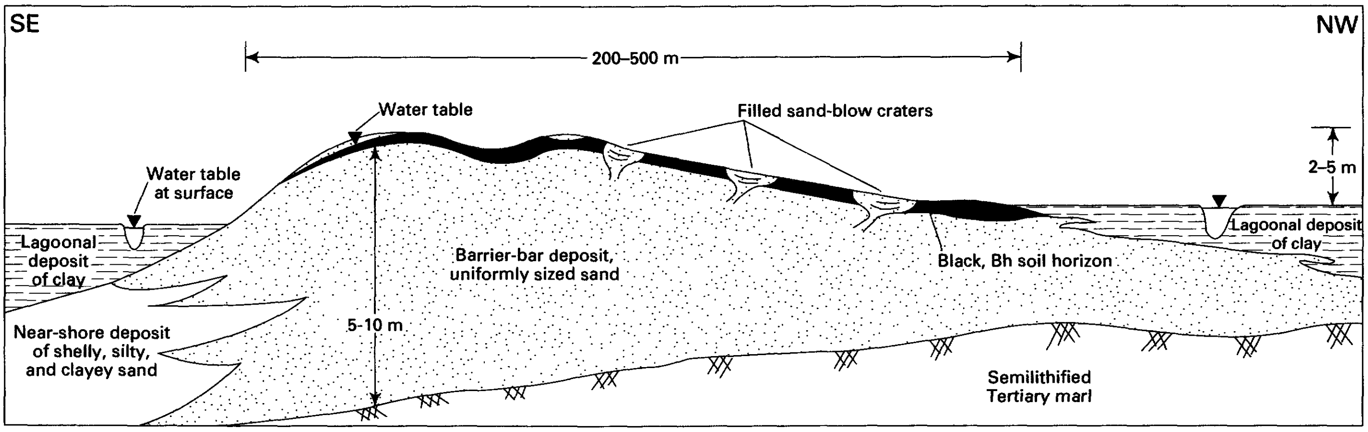

FIGURE 2.—Schematic cross section of representative barrier showing sediment types, ground-water table locations, filled sand-blow craters, and Bh (humate-rich) soil horizons. Modern shoreline is located southeastward. Lagoonal clay deposit at left is younger and lower in elevation than the barrier-bar deposit.

Features large enough to be interpreted as possibly having an earthquake origin in the low country were found only in sand deposits having a total thickness exceeding 2 to 3 m; within this 2- to 3-m-thick deposit, the thinnest individual source stratum was 0.3 m.

(3) The local geologic setting has a major role in the formation of earthquake-induced sand blows. The geologic setting most frequently associated with sand blows is the crest or flank of Pleistocene beach ridges, where a thin surficial cover of a clay-bearing sand or humate-rich sand overlies clean sand. According to first-hand observations of effects of the 1886 earthquake by Earl Sloan, "these craterlets are found in greatest abundance in belts parallel with (beach) ridges and along their anticlines" (Peters and Herrmann, 1986, p. 68). A schematic cross section through a typical low-country beach ridge, such as the ridges described by Sloan, is presented in figure 2. To a much lesser extent, sand blows have been found in back-barrier environments.

A thin clay-bearing stratum (or other stratum having very low permeability) above the liquefied zone is generally an important control on development of sand blows (Scott and Zuckerman, 1973; Obermeier, 1988; Ishihara, 1985). A veneer of nonliquefiable sediment, 1 to 2 m thick, aids greatly in the formation of and recognition of sand blows in the low country. On the other hand, a clay-bearing or other low-permeability stratum thicker than 4 to 5 m prevents sand-blow formation at the great majority of sites in the low country; a thickness greater than 2 to 3 m seriously impedes such formation.

CHARACTERISTICS OF AND CRITERIA FOR SAND-BLOW FORMATION

Two kinds of pre-1886 sand blows have been recognized: (1) filled sand-blow craters (craterlets) and associated sedimentary structures and (2) sand volcanoes that have vented to the surface, leaving relict sand mounds. The crater-type sand blows generally occur only where a surficial soil having less than several percent clay has formed on a parent material of clean sand. Where the surficial soil is richer in clay, vented-sand volcanoes are much more likely to form.

Exact locations were not known for 1886 craterlet sand blows when our study was initiated in 1983. Only one 1886 vented type of sand volcano had been discovered and examined (Cox, 1984). Thus it has been necessary to locate sedimentary features that display structures consistent with an earthquake origin and to develop criteria for interpreting whether or not these features have an earthquake origin.

TABLE 1.—*Estimated relative susceptibility of saturated cohesionless sands to liquefaction during strong seismic shaking*

[Source: Youd and Perkins (1978). Strong seismic shaking is determined by two parameters, peak horizontal acceleration and duration of largest acceleration. For a $m_b = 5$ earthquake, strong seismic shaking is defined as an acceleration of about 0.2 g for at least several seconds; for a stronger earthquake, the threshold acceleration is about 0.15 g for a duration of 10 seconds; for a much stronger and longer duration earthquake, the threshold acceleration can be less than 0.1 g (T.L. Youd, Brigham Young University, oral commun., 1985)]

Types of deposit	Liquefaction susceptibility for deposits of various ages: Holocene <500 yr	Holocene <500 yr	Pleistocene	Pre-Pleistocene
Dunes	High	Moderate	Low	Very low
Beach (low wave energy	High	Moderate	Low	Very low
Foreshore	High	Moderate	Low	Very low
Beach (high wave energy)	Moderate	Low	Very low	Very low
River channel	Very high	High	Low	Very low
Flood plain	High	Moderate	Low	Very low

Geologic criteria that we have developed for interpreting whether near-surface features are earthquake-induced sand blows generally have four elements:

1. The features have sedimentary characteristics that are consistent with an earthquake-induced liquefaction origin; that is, there is evidence of an upward-directed, strong hydraulic force that was suddenly applied and was of short duration.

2. Characteristics such as shape, width, and depth are consistent with historical observations of liquefaction during the 1886 earthquake.

3. The features are in ground-water settings where a suddenly applied, strong hydraulic force of short duration could not be reasonably expected except from earthquake-induced liquefaction. In particular, these settings are extremely unlikely sites for artesian springs.

4. Similar features occur at multiple locations, preferably within a few kilometers of one another, having similar geologic and ground-water settings. Where evidence of age is present, it should support the interpretation that the features formed in one or more discrete, short episodes that individually affected a large area and the episodes were separated by long time periods during which no such features formed.

As fewer of these criteria are satisfied, the confidence in an earthquake origin generally diminishes. Subsequent sections describe the application of the four criteria.

Whether earthquakes or other mechanisms induce small liquefaction-flowage features is often impossible to determine. Small synsedimentary liquefaction-flowage features such as dikes and sills as much as 0.25 to 0.5 m long and a few centimeters thick are not unusual in point-bar deposits and in beach surf zones. Very small features similar to sand blows are also common in beach surf zones. These small features result from a variety of forces, including dynamic wave loadings and static slumping. An earthquake origin was considered a possible mechanism in this study only if the flowage features were much larger than these small features and if, in addition, the flowage features cut the surface soil profile.

FILLED SAND-BLOW CRATERS

All filled sand-blow craters belong to a single morphologic group having many common features in the fill sediments, although systematically occurring variants also occur. The normal type of filled sand-blow crater is discussed first. In a later section there is discussion of two variants that may indicate association with earthquake-induced landslides.

Almost all pre-1886 sand-blow sites shown on figure 1 have sand blows whose original morphology and size are comparable to the 1886 craters described by Dutton (1889), except that the craters are now filled with sediment. Figure 3 shows craters that represent moderate- to large-sized craters produced by the 1886 earthquake. A crater is a hole at the ground surface that forms as liquefied sand vents to the surface. In the process, the forceful upward surging of sand and water also scours and enlarges a hole and deposits sediment beyond the crater. When flowage stops, a surficial sheet of ejected sand and soil surrounds the rim of the crater, and clasts of dark soil are commonly scattered along the base of the steep wall (fig. 3). Examination of photographs shows that, in 1886, the surficial sheets commonly appear to have had thicknesses near crater rims of about 15 to 20 cm and maximum diameters rarely exceeding about 3 to 4 m. The maximum reported thickness of vented sand was 1 m, and the maximum crater diameter was about 6 m (Dutton, 1889).

About 75 percent of the sites on figure 1 are on barrier and nearshore marine sands. The sands are almost exclusively well-sorted, fine- to medium-grained quartz, with less than 5 percent heavy minerals and mica (McCartan and others, 1984; Gelinas, 1986). Surface weathering has imposed a soil profile on these marine-related sands. The soils are classified as spodosols, alfisols, and ultisols, but most sand blows occur in poorly drained areas where humic spodosols (humods) have formed. Humods are characterized by a thin (<10- to 15-cm) surficial A horizon (organic matter and several percent sand) overlying a thin (<10- to 15-cm), very light gray E horizon; the E horizon overlies a thick (0.5- to 1.5-m) black Bh horizon (humate-enriched sand containing a few percent clay) and a variably thick (0.1- to 1-m), gray to light-yellow B–C horizon (transition zone between B and C horizons). The B–C grades down into C-horizon sands (parent material), which are very light gray in the upper 1 to 2 m and grade down to a greenish hue. Beneath the upper 0.5 to 1 m of the C horizon, the original bedding consists of thin, horizontal, black heavy-mineral laminae about 0.25 to 0.5 cm apart. The sand-blow features cut the solum and the C horizon. The source sand beds that liquefied during earthquake shaking generally occur within the depth range of 3 to 10 m.

Figure 4 is a vertical section of a filled sand-blow crater that is representative of the type observed at almost all of our study sites (fig. 1). The figure illustrates characteristics that we consider to be compatible with an earthquake-induced liquefaction origin. In figure 4, the soil horizon is cut by an irregular crater, which is filled with stratified to structureless (that is, nonstratified) and graded sediments. The fill materials are fine- to medium-grained sand and clasts from the Bh, B–C, and C soil horizons and sand from depths much below the exposed C horizon. Walls of the crater are generally

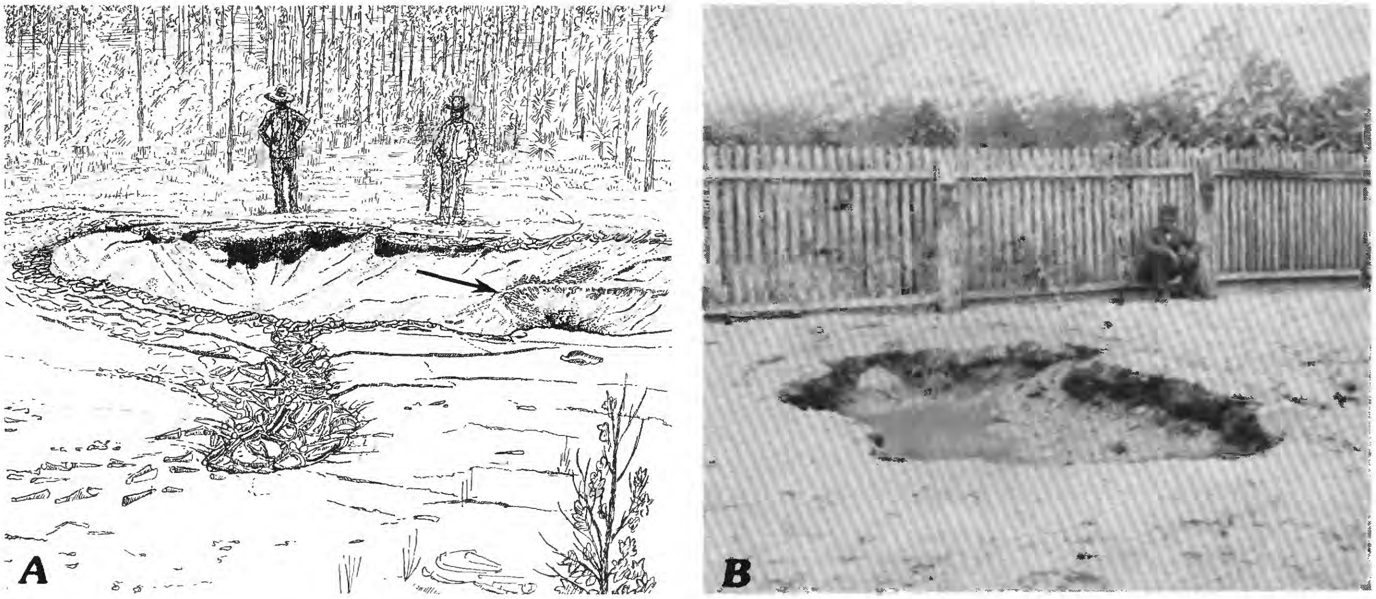

FIGURE 3.—Craters produced by the 1886 earthquake. *A*, Sketch from a photograph of an 1886 crater (sand blow at Ten Mile Hill, near the present Charleston airport). Note that the crater contains sand sloughing toward the lowest parts and that there is a constructional sand volcano located in the right part of the crater (at arrow). The crater is surrounded by a thin blanket of sand partly veneered with cracked mud. *B*, Photograph of typical crater produced by the 1886 earthquake. Note the thin blanket of ejected sand around the crater and sand and clasts of dark soil within the crater. (Photograph from the archives of the Charleston museum.)

smooth and sharp when viewed up close, especially in the lower part of the crater. Walls of some craters are very jagged, however.

The Bh horizon generally is much thinner over the central part of the filled crater than on the sides, and the Bh horizon of the laterally adjoining undisturbed soil is abruptly thicker. Clay content in the Bh horizon in the crater is much less than in the undisturbed soil. Figure 5 is a sand-silt-clay ternary diagram of particle sizes of pairs of samples taken at identical depths inside and outside the liquefaction craters. The lines connecting pairs of samples show that crater samples have consistently less clay and silt than the adjacent undisturbed soil. The Bh horizon on the filled crater typically is thicker, is more clay rich, and has better developed soil structure (that is, peds) with increasing age; craters older than about 5,000 to 10,000 years have Bh horizons that approach the thickness and development of those in laterally adjacent undisturbed soils.

A sequence of five layers, each having specific characteristics, occurs beneath the Bh horizon in the filled crater. The sedimentary characteristics of these layers are less distinct in older craters because of pedogenesis in the crater deposits. Layer 5 (fig. 4) is a structureless (that is, massive), gray, humate-stained sand, which overlies a thinly (2- to 3-mm) laminated sequence of alternating light- and dark-colored sands (layer 4). The lamina typically are discontinuous and irregular in thickness, as illustrated in figure 4. The dark color is generally imparted by humate staining; the dark lamina may contain appreciable amounts of silt and clay. The basal bed of this sequence (layer 4) is clay-rich in perhaps 10 percent of the filled craters and is rarely thicker than 1 cm. The basal bed sharply overlies a medium-gray structureless sand (layer 3). Layer 3 contains many small clasts (1- to 5-mm diameter) of Bh material, charcoal, and wood. This clast-rich layer grades down into a structureless sand zone (layer 2) containing many intermediate-sized clasts (5- to 20-mm diameter) of friable Bh material and occasional extremely friable clasts of light-colored sand. The clasts of Bh material and sand have, respectively, the same color, mineralogy, consistency (that is, resistance to being crushed between fingers), and friability as the adjoining sands of the B–C and C horizons. Layer 2 grades down into layer 1, containing densely packed intermediate-sized (1–5 cm) and large-sized (>5 cm) clasts of Bh material in a structureless sand matrix; the large clasts have diameters exceeding 25 cm in many filled craters. Many of the clasts in layer 1 have their long axes vertically oriented. At and below the thinly stratified sequence (layer 4), the sides of the bowl are sharply defined by a color boundary and by the presence of clasts within the bowl.

Beneath the bowl are vents containing structureless clean sand. Sides of the vents sharply cut bedding in the C horizon.

The matrix sand in layers 1 through 3 contains an extremely small percent of clay-sized material and clay

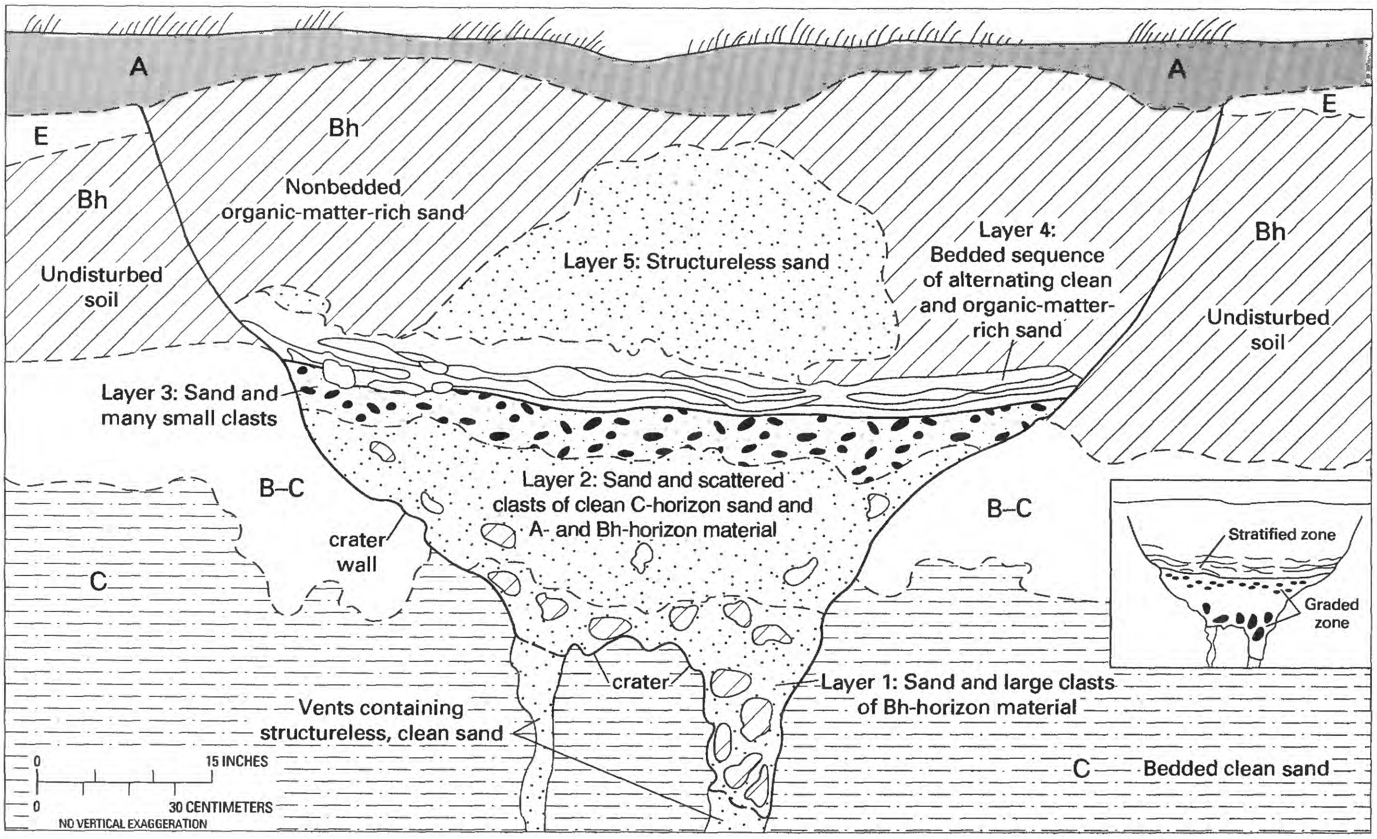

FIGURE 4.—Schematic cross section of normal type of filled sand-blow crater. Letters correspond to soil horizon designations. The filled crater in this figure much predates the 1886 earthquake, based on thickness of Bh horizon in the filled crater.

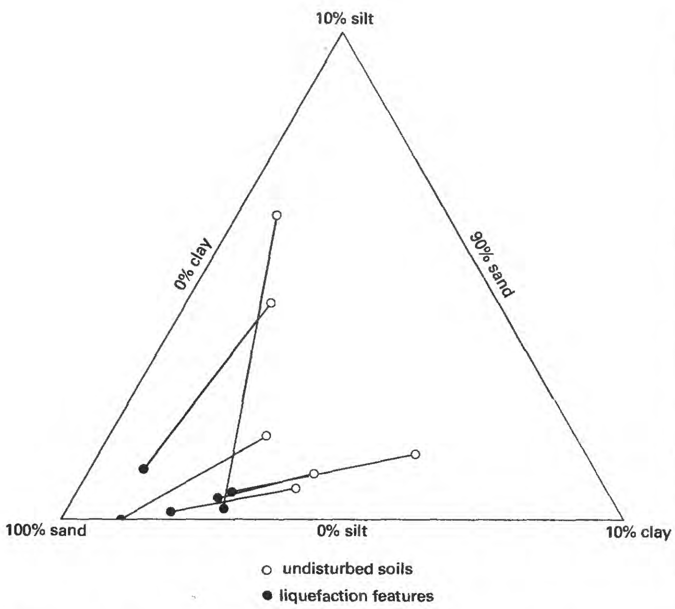

FIGURE 5.—Ternary diagram showing grain sizes of paired samples at identical depths inside and outside liquefaction craters.

minerals. Both the percent of clay minerals and percent of clay- and silt-sized material in layers 1 through 3 are much less than in the laterally adjoining undisturbed B–C and C horizons, particularly for craters formed less than a few thousand years ago (Gelinas, 1986). The maximum grain size of the matrix sand in layers 1 through 3 is greater than in the laterally adjoining and overlying undisturbed materials in some filled craters.

The following phases have been interpreted in the formation of the filled sand-blow craters: (1) after earthquake-induced liquefaction at depth, a large hole is excavated at the surface by the violent upward discharge of the liquefied mixture of sand and water; (2) a sand rim accumulates around the hole by continued expulsion of liquefied sand and water after the violent discharge; (3) sand, soil clasts, and water are churned in the lower part of the bowl, followed by settling of the larger clasts and formation of the graded-fill sequence of sediments; and (4) the crater is intermittently filled by adjacent surface materials to form the thin stratified-fill sequence, during the weeks to years after the eruption. In the craters predating the 1886 earthquake, the sand blanket ejected from the crater is indistinguishable in the field from the surface and near-surface (A, E, and Bh) soil horizons, because the blanket has been incorporated within these soil horizons.

This interpretation of origin of the filled craters is based in part on comparison with photographs and descriptions of 1886 craters. The filled craters are similar to the 1886 craters shown in figure 3 in that the dimensions (diameters) at the surface are about the same and the depths to the contact of the stratified fill to the graded fill (see fig. 4) are about the same. The lowermost stratified layers of filled craters (layer 4) are composed of humate-stained, slightly clayey or clean sand that generally contains a few clasts of Bh horizon material and, in a few filled craters, clasts of C horizon sand. (See fig. 4.) Presence of the intact very weak, friable clasts of C horizon sand suggests very rapid initial infilling of the crater, as illustrated in figure 3*B*. In the filled craters, clasts of C-horizon material first must have been ejected onto the adjacent ground surface or walls of the crater before falling back or washing back into the bottom of the crater and being deposited within the beds of layer 4. We know of no other reasonable explanation of the means by which these clasts could have been transported to the stratified zone.

The presence of friable, angular clasts of C- and Bh-horizon material in the graded-fill portion is consistent with a short-lived, churning type of upwelling from the vent. Water commonly flows for a day or so from earthquake-induced sand blows. The violent, boiling phase is much shorter in duration. Hence, the presence of friable clasts argues against a long-term artesian spring origin for these features; a spring-induced long-term churning phase, lasting days, that slowly diminishes in flow would abrade, round, and (or) disintegrate the clasts of Bh- and C-horizon material. Short-term, nonearthquake-related springs have been eliminated as a possible mechanism by which craters are formed along the crest of the beach ridges, because such springs cannot form in this topographic-geologic setting. (This point is discussed in detail later.)

Our interpretation of the origin of the craters is also supported by the presence of sand-filled tabular fractures, whose overall shape and dimensions strongly suggest that they are "incipient craters." These fractures are rather common at some places in the meizoseismal zone of the 1886 earthquake, where craters are plentiful. Figure 6 shows V- and U-shaped fractures (fig. 6*A* also shows a connecting vent) that are filled with sand we believe was transported upward from depth, on the basis of the freshness of minerals in the fractures. The fractures, which are tabular, generally widen with depth until they connect to a single, near-vertical large sand-filled fissure (that is, a vent). The sand-filled fractures probably represent the early phase of development of craters; for the features in figure 6, however, the upward forces were too weak to excavate the overlying material.

It is possible that liquefaction led to the production of craterlets because of a fortuitous combination of sediment properties in and above the zone that liquefied *during* earthquake shaking. The source beds that liquefied were exceptionally susceptible of liquefaction, in that generally they were very loose (engineering sense), fine-grained, uniformly sized, and free of clay (Dickenson and others, 1988); these properties would cause the source beds to liquefy abruptly and, once liquefied, to flow readily (Seed and others, 1983; Youd, 1973). We suspect that the liquefied sand strata suddenly applied a large point force to the overlying sediment (through a hole or weak zone such as that left by a decayed root), causing a V- or U-shaped crack to form, through which the liquefied sand violently vented because of its exceptional ability to flow. The V- and U-shaped cracks occurred because overlying sediment is humate cemented, has no pronounced planes of weakness, and is very brittle; the process is similar to formation conchoidal fracture in an isotropic, brittle medium, caused by the application of a point load.

In summary, the filled craters that we found have a morphology that is consistent with descriptions and photographs of craters caused by the 1886 earthquake. The general geologic setting (Pleistocene beach crests and flanks) and the locations of swarms of sand blows reported by geologists immediately after the 1886 earthquake (Peters and Herrmann, 1986) are very near or coincident with two sand-blow sites found during this study (HW and ARP on fig. 1). The morphology of the walls of the craters, the stratigraphy of the fill of the craters, and the evidence that they formed in relatively sudden, discrete episodes serve to demonstrate that the filled craters are the result of earthquake-induced liquefaction. Alternate origins, such as short- or long-term artesian springs, ocean wave pounding, thrown trees, or other mechanisms that have the potential to create similar features are discussed in detail in the section entitled "Other Possible Origins."

VENTED-SAND VOLCANOES

Figure 7 is a schematic cross section of a pre-1886 vented-sand volcano. Where the zone that liquefied at depth is overlain by a thick clay-rich stratum, a sand blow typically is expressed on the ground surface as a mound of sand. The thickest part of the mound ranges from a few centimeters to as much as 0.25 m. At a few

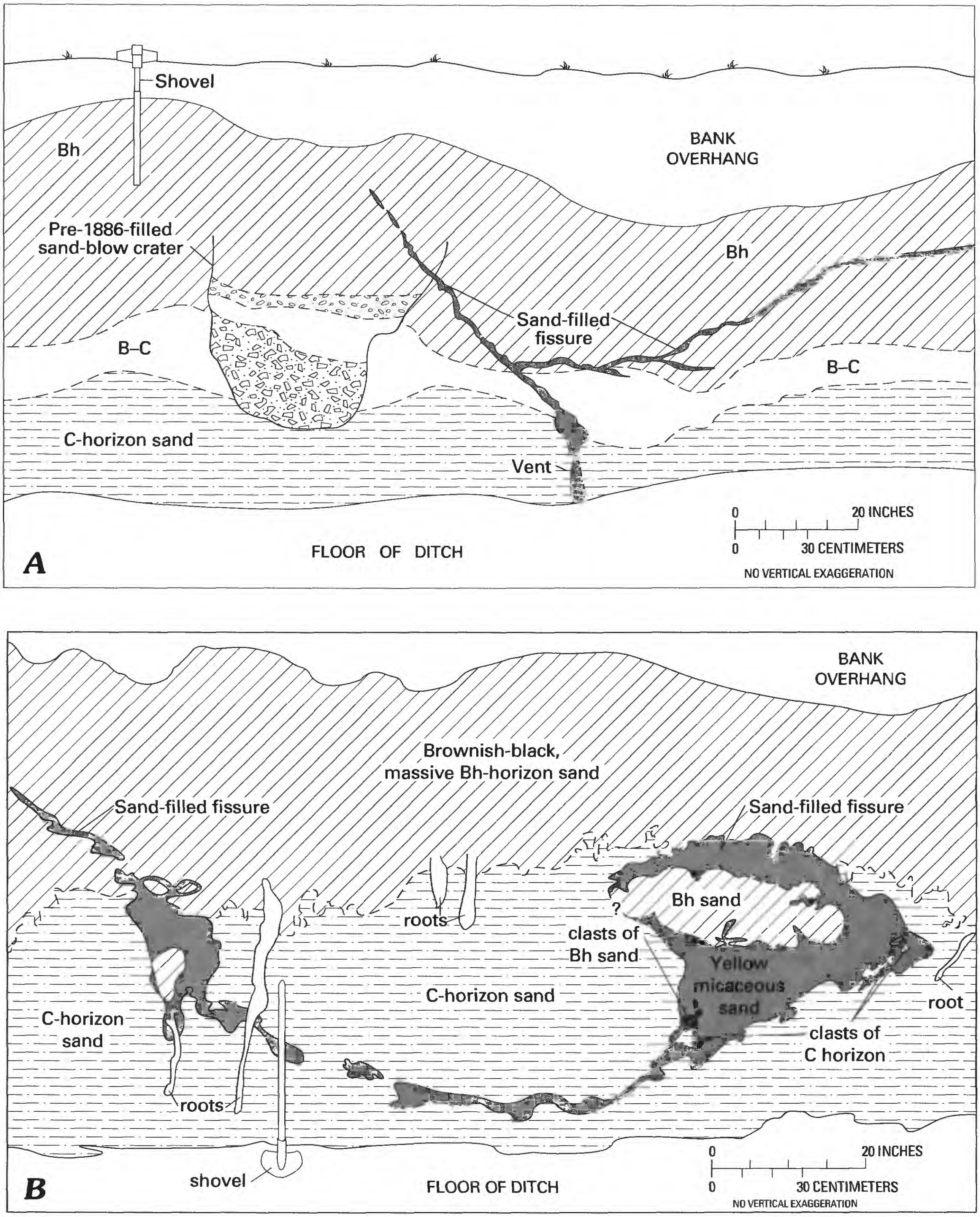

FIGURE 6.—Schematic cross sections showing sand-filled fissures interpreted as resulting from liquefaction and flowage during the 1886 earthquake. Light-yellow, intruded sand in fissures is determined to have been vented from a depth of 6 m, on the basis of grain-size and mineralogical data. *A*, V-shaped sand-filled fissure and vent. Fissure cuts soil horizon developed on pre-1886 filled sand-blow crater. *B*, U-shaped sand-filled fissure.

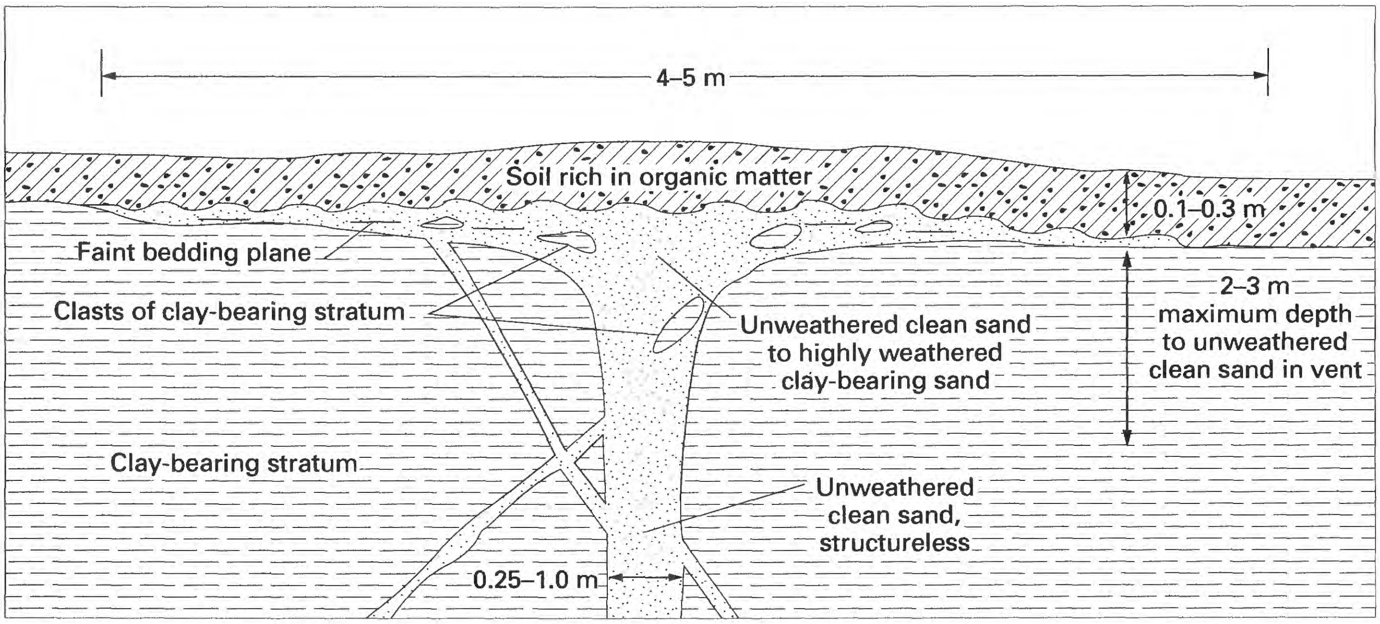

FIGURE 7.—Schematic vertical section of representative vented-sand volcano.

localities, mounds have discernible bedding. The mounds are generally thickest above their junctions with the widest steeply dipping sand-filled dikes (that is, feeder vents) that extend downward through the clay-bearing stratum. Often the sand in the mound and in the dikes contains clasts of the clay-bearing stratum that have been torn from the walls of the vent. The sand in the dikes is structureless and has no discernible bedding; the sand in the central parts of the dikes may be better sorted and coarser than that near the edges. Dike widths range from about 0.25 to 1 m.

An 1886 sand blow of this type was described by Cox (1984). Similar sand blows have been observed at many places around the world, including the New Madrid seismic zone (to be discussed later). Figure 8 shows a variant of this type of sand blow, in which the vented sand has coalesced into a continuous surface cover; some of the white surface sand is pedogenic E horizon, which is virtually indistinguishable from the older, vented sand. The volume of fluidized sand expelled to the surface has been so large at some places that the clay-bearing stratum has downdropped noticeably. At the site shown in figure 8, for example, comparison of soil structure and clay content of the undisturbed soil with the clay-bearing stratum above the hand shovel showed that the stratum above the hand shovel had been downdropped about 1 m.

Near-surface sediments in the depth range of 1.5 to 2 m are commonly so intensely weathered and discolored that textural analysis is required to determine a possible origin by venting. A useful (but not sufficient) test for venting is comparison of the coarsest sand fraction in the suspected vent with the grain size of sand in the laterally adjoining clay-bearing stratum or soil horizon. An earthquake origin for the sand is not considered likely unless the coarsest sand in the dike is significantly different in size from that in the crosscut clay-bearing stratum. Such a textural difference cannot be the result of soil-forming processes. In addition, there must be no possibility that the coarsest sand fraction has been introduced from above the vent. At three vented-sand volcano sites shown on figure 1 (sites CH, BR, and SAN) that have been interpreted as earthquake induced, other evidence of venting is also present. Examples include clay-bearing clasts in the sand mound combined with sand-filled sills and dikes that widen downward and, at depth, extend into slightly weathered and unweathered sediments. At some places, the sand-filled dikes and sills cut through ground that appears to have been shattered (that is, irregularly, intensively fractured and intruded), which is very suggestive of forceful intrusion.

Features similar to the vented-sand volcanoes, but for which springs or other nonearthquake sources cannot be easily eliminated as possible origins, are common in

FIGURE 8.—Vented-sand volcanoes that have coalesced to form a continuous sheet of sand on the ground surface. At least two episodes of venting separated by long periods of time are represented. Note irregular pattern of fracturing. (Shovel is 60 cm long.)

many lowland areas. These sites have not been included on figure 1.

FEATURES SHOWING EVIDENCE OF LATERAL SPREADS OR GROUND OSCILLATIONS

Lateral spreads[3] were commonplace features of the 1886 earthquake (Peters and Herrmann, 1986). Most of these spreads developed in fluvial terrace deposits, bordering streams in low areas. Obvious surface evidence of lateral spreads has not persisted to the present, and ditches and pits are so rare in the wet locales where lateral spreads formed that none has been found. However, along the flanks of some Pleistocene beaches, reverse shears are present that may indicate an incipient formation of lateral spreads or ground oscillations. Shear displacements commonly range from 1 to 4 cm. Slope of the ground surface is typically less than 1 percent for hundreds of meters downslope or upslope. These slopes are so gentle and the possibility of high artesian pressures is so remote that gravity-induced slumping is virtually impossible. Some of these reverse shears almost certainly have an origin in earthquake-induced liquefaction. At one site, for example, reverse shears dipping in opposite directions (toward one another) formed about 10 m apart in the stratum that liquefied during shaking, and sand blows traceable to this liquefied stratum formed between the shears; the most likely mechanism that could have formed the opposite facing shears was alternating directions of ground movement, caused by earthquake oscillations, in the liquefied stratum. (Such a mechanism is illustrated as fig. 2–11 in "Liquefaction of Soils during Earthquakes," published by the National Academy Press, Washington, D.C., in 1985.)

Reverse shears can occur as isolated features but are generally found in association with sand blows. Figure 9 shows the typical relationship for filled craters. A reverse shear is present on the downslope side of the crater, and the upthrown block cuts the C, B–C, and Bh horizons. The shear formed prior to venting of sand from the source stratum at depth, because the vent is not cut or distorted.

At some sites, the shears along crater edges could have formed only in response to earthquake-induced lateral spread movement because the shears are traceable into and along the bedding of the stratum that liquefied. Only rarely is an exposure deep enough to allow determination of whether the shear goes into a stratum that liquefied, and thus an earthquake origin cannot be confidently assigned to all sites. However, we are of the opinion that these reverse shears along crater edges suggest an earthquake origin, even where the

[3]A lateral spread is a landslide formed by a laterally moving slab, commonly of large areal extent; earthquake-induced lateral spreads commonly form where surface slopes range between 0.5 and 5 percent (Youd, 1978). The presence of lateral spreads generally indicates a large areal extent of liquefied material.

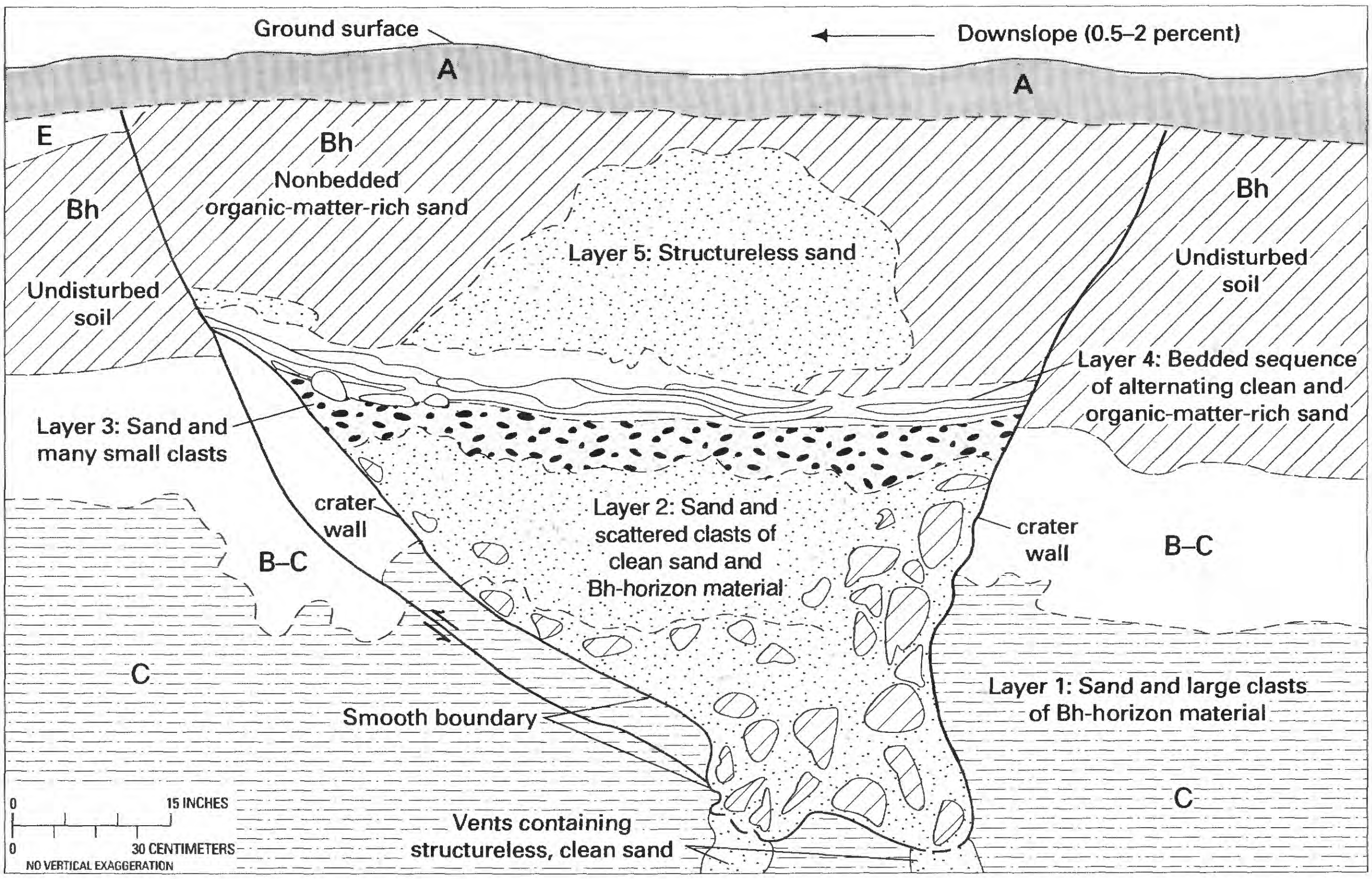

FIGURE 9.—Schematic cross section of filled sand-blow crater, illustrating aspects possibly associable with downslope movement. A reverse shear soil is along the lower left side of the filled crater; an underturned edge forms the lower right part of the filled crater beneath the B–C and C soil horizons; large clasts along the right side are much higher than in other soil parts of the filled crater.

shears have formed on gently sloping ground (less than 2 percent) as much as 5 m below the beach crest.

Another aspect of filled craters that possibly indicates an incipient lateral spread is also illustrated by the position of the clasts along the right side of the crater of figure 9. Large clasts occupy the region from the vent to the vicinity of the bedded sequence (layer 4). The lower part of this clast zone turns beneath the B–C and C horizons. These clast relations are found in perhaps 25 to 50 percent of craters having reverse shears. The consistent positional relationship of underturned edge with clasts along the crater edge suggests control by minor downslope movement, which took place while the sediments in the crater were still fluid enough to segregate slightly.

REGIONAL DISTRIBUTION OF SAND BLOWS

The regional distribution of sand blows is systematic and predictable in coastal South Carolina, in terms of size and number of sand blows. This distribution conforms with the pattern that would be anticipated from an earthquake origin.

Numerous worldwide observations show that the number, size, and type of earthquake-induced liquefaction features are a reflection of three principal factors. Of primary importance are (1) severity of shaking (controlled by earthquake accelerations and duration (Seed and Idriss, 1982)), (2) liquefaction susceptibility (controlled largely by state of compactness and cementation (Youd and Perkins (1978)), and (3) local geologic controls (affected primarily by depth to water table, thickness of impermeable sediments above liquefied stratum, deformation properties, permeability of sediments above liquefied stratum, and permeability of liquefied stratum (Ishihara, 1985; Obermeier, 1989; Obermeier and others, 1986)). The local geologic controls on the Pleistocene beaches in coastal South Carolina are constrained within a very narrow range because (1) bedrock accelerations should have been amplified about the same amount at many places by previous earthquakes, (2) the liquefaction susceptibility of the source sand beds has been about the same, and (3) local geologic controls on morphology and size of sand blows on beach ridges has been about the same, for a time period extending at least throughout the Holocene (Obermeier and others, 1989; also discussed later). Thus, in the vicinity of the epicentral region, the sand blows should generally be largest (have largest diameters) and be most abundant. This hypothesis is confirmed in the Charleston region by observations of effects of the 1886 earthquake (Peters and Herrmann, 1986). Sand blows produced by the 1886 earthquake were abundant within the 1886 meizoseismal zone, which was a region about 35 km wide and 50 km long. Beyond this zone, the sand blows were smaller and scattered. Only rare, small sand blows formed more than 10 km beyond the meizoseismal boundary. Because all the factors involving sand-blow production are about the same in many places throughout coastal South Carolina, this same relation of sand blows to epicenter location should hold true for pre-1886 Holocene earthquakes.

In particular, it has been found that for craters having one apparent age (based on soil profile) and whose diameters (measured near the ground surface) are less than about 1 m, there are at most two or three craters exposed in a nearby 1-km-long ditch cutting across beach ridge deposits. Wherever maximum diameters of craters having one apparent age are about 2 m, there are more (as many as 5 to 10) craters exposed in a nearby 1-km-long ditch. Wherever maximum diameters of craters having one apparent age are 3 m or larger, there are a greater number (as many as 20) exposed in a nearby 1-km-long ditch. This well-defined relation between number of sand blows and sand-blow diameters is consistent with an earthquake origin in our opinion.

We emphasize that a single, isolated feature that appears to be earthquake-induced would not be interpreted by us to be compelling evidence for prior earthquake shaking. Compelling evidence requires at least several features, scattered over a region of at least several square kilometers.

OTHER POSSIBLE ORIGINS

Origins other than earthquakes for filled sand-blow craters and vented-sand volcanoes that have been considered include compaction-induced dewatering and soft-sediment deformation; artesian springs; landslides; fillings in decayed stump and root holes; ground disruption by fallen, root-wadded trees; and liquefaction caused by storm-induced, ocean-wave pounding. Criteria for assessing each of these potential sources are discussed below.

Compaction-induced dewatering and deformation.— Syndepositional and postdepositional dewatering by compaction occurs in sediments during or shortly after their deposition. This dewatering generally takes place in response to rapid deposition of sediments (especially coarser, denser sediments) above very soft, clay-rich sediments, which causes buildup of pore-water pressure and gravitational instability (Dzulynski and Walton, 1965; Allen, 1984; Lowe, 1975, 1976); deposition of silts or fine sands rapid enough to cause buildup of pore-water pressure can also lead to instability (Dzulynski and Smith, 1963; Sanders, 1960). The kinds of sedimentary structures formed by these processes include load structures, dish structures, convolute lamination, sand dikes, and faults (Pettijohn and Potter, 1964). In our study, earthquake-interpreted features generally have soil horizons that are much thinner and less well developed (and thus younger) than soil horizons on laterally adjoining, undisturbed parent sediments; in the relatively few instances where both the parent sediments and crosscutting features have essentially the same degree of soil development, the earthquake-interpreted features contain clasts of Bh material at depths generally about 0.5 m below the laterally adjoining Bh material in the parent sediment. The apparent difference in age between the parent sediment and filled-crater or vented-sand volcano determined on the basis of soil development, combined with lack of reason to suspect sudden, nonseismic surface stressing or long-term pore-water pressure buildup, is sufficient reason to eliminate postdepositional dewatering as a source mechanism.

Artesian springs.—The regional and local topographic and ground-water setting of many sites rules out any significant likelihood for the occurrence of a sudden, strong increase in the hydraulic force of a spring. The beach crests are generally flat lying for many kilometers along their crests, and the crests are well above the lagoonal deposits. Where filled craters are found on ancient beaches, a short- or long-term spring origin is not believed possible if the following conditions are met: (1) the filled craters cut humate-rich sandy soils that are also cut with numerous, highly permeable sand-filled root holes and burrows that extend well into the C horizon, (2) the filled craters are much above the lagoonal deposits adjoining the beach (as illustrated in fig. 2), and (3) the filled craters are within 1 to 2 m of the beach crest. Short-term springs induced by a hurricane deluge (or any other mechanism) have not been observed where these criteria are met.

Short-term springs are suspected to be rather common in lowland areas of lagoonal deposits, however. Features interpreted as earthquake-induced sand volcanoes are restricted to sites where artesian springs are thought to have been unlikely, and in addition, there is other evidence for an earthquake mechanism, such as fractured ground (as illustrated in fig. 8). Typical sites are elevated locations near deeply entrenched rivers, where artesian pressures would have been relieved by lateral flow to the rivers rather than by vertical flow to higher elevations, which would have been required to form the volcanoes.

Landslides.—All sites on figure 1 are on level or nearly level ground, hundreds of meters from any steep slopes. Downslope movement on these nearly level sites could be initiated only by seismic shaking, considering relations between surface topography, ground-water setting, and strength of materials.

Fillings in root holes.—Holes caused by decayed stumps and roots, and later filled with clean sand, are common in the study area. Although the filled holes are generally circular in plan view, many display poorly defined layers of clay mineral segregation (due to weathering) and gradually taper downward; they do not have the well-defined laminar bedding of the filled craters and underlying graded zone of sediments. Typically, maximum depths of root penetration are about 2 to 3 m.

Ground disruption by trees.—Tree throw in the South Carolina low country is generally restricted to hardwood species having wide, shallow root systems. These trees may blow over in wind storms and rip up a wad of sediment caught within their root mats, thus creating pits and mounds. Taprooted pines and cypress trees very rarely tend to throw; instead, these species break off near the ground surface. Pits excavated by hardwood tree throw tend to be shallow (usually less than 50–100 cm deep); when filled, they do not contain sediment introduced from depth, and they almost always lack the orderly internal stratigraphy of liquefaction craters. At some few places, though, pits excavated by thrown trees can be distinguished from craters excavated by earthquake-induced liquefaction only by the presence of feeder vents or by the presence in the crater of sediment (generally sand or silt) that has been introduced from strata beneath the base of the crater. Verification that sediment has been introduced from depth may require mineralogical, weathering, and grain-size analysis (Gelinas, 1986).

Ocean-wave pounding.—The disruption of subaerially formed soils shows that liquefaction occurred long after deposition of the parent sediments. Storm-induced, ocean-wave poundings are not a credible mechanism of liquefaction at most of the widely scattered sites because the elevations are well above modern sea level (up to 15 m) and are too far inland. In addition, no sedimentary records indicative of inland surges of the ocean, such as soil horizons buried by storm-deposited sediments, have been found.

FEATURES OF WEATHERING ORIGIN

A wide variety of features produced by chemical weathering mimic earthquake-induced liquefaction features. Such weathering features include the white, E-horizon sand that commonly blankets the surface; pedogenic tongues; and BE or fragipan horizons. Distinguishing liquefaction features from weathering features is much more difficult where older liquefaction features have been extensively weathered.

A loose, clean, white sand blanket covers large parts of the South Carolina low country that is underlain by sandy sediments of barrier beach and nearshore marine origin. In undisturbed soil profiles, the white sand underlies a thin, 0- to 15-cm-thick, dark-gray A horizon. Although some of this white sand has been periodically remobilized by surficial processes, it is a pedogenic E horizon, formed by weathering and leaching of the underlying sandy sediments (Gamble, 1965). The pedogenic origin of the E horizon is demonstrable by its eluvial-illuvial relationship with the underlying B horizon. Clays and labile minerals have been removed from the E horizon, and weathering products have been deposited in the B horizon; however, particle-size distribution and resistant heavy-mineral percentages are nearly identical in both horizons. In addition, the boundary between the E and B horizons is commonly abrupt and irregular and is characterized by narrow, near-vertical tongues of white E horizon penetrating downward into the B horizon. Laterally continuous, interpenetrating boundaries of this type are more likely to be indicative of geochemical leaching rather than clastic deposition. At some sites resistant sedimentary features, such as thin pebble beds, pass through the horizon boundary, showing that it is not a sedimentary contact.

In summary, although a blanket of ejected white sand often exists at earthquake liquefaction sites, all white sand blankets are not necessarily formed by earthquake liquefaction. In younger liquefaction features, bedding within ejected sand blankets helps to demonstrate liquefaction origin if other fluvial and eolian origins can be rejected. With age, the usefulness of this criterion decreases as soil mixing by flora and fauna destroys bedding, making ejected sand blankets appear superficially like massive pedogenic E horizons.

The pedogenic boundary between E and B horizons can be gradational or quite sharp and can also be highly convoluted. Gamble (1965) describes boundaries between loose, white E-horizon sand that grades over 1 to 2 mm of depth to brown, clayey sand. The boundary also commonly has 2- to 3-cm-wide, 5- to 10-cm-long tongues of E horizon descending vertically into the underlying B horizon. Locally, we have observed tongues of E-horizon sand that extend more than a meter into thick, red to brown (7.5*YR* 5/6–5*YR* 5/8), clayey B horizon (fig. 10). Tongues of this size and shape can give the impression of fractured and brecciated ground and might be mistaken for liquefaction features unless examined carefully for sedimentary characteristics.

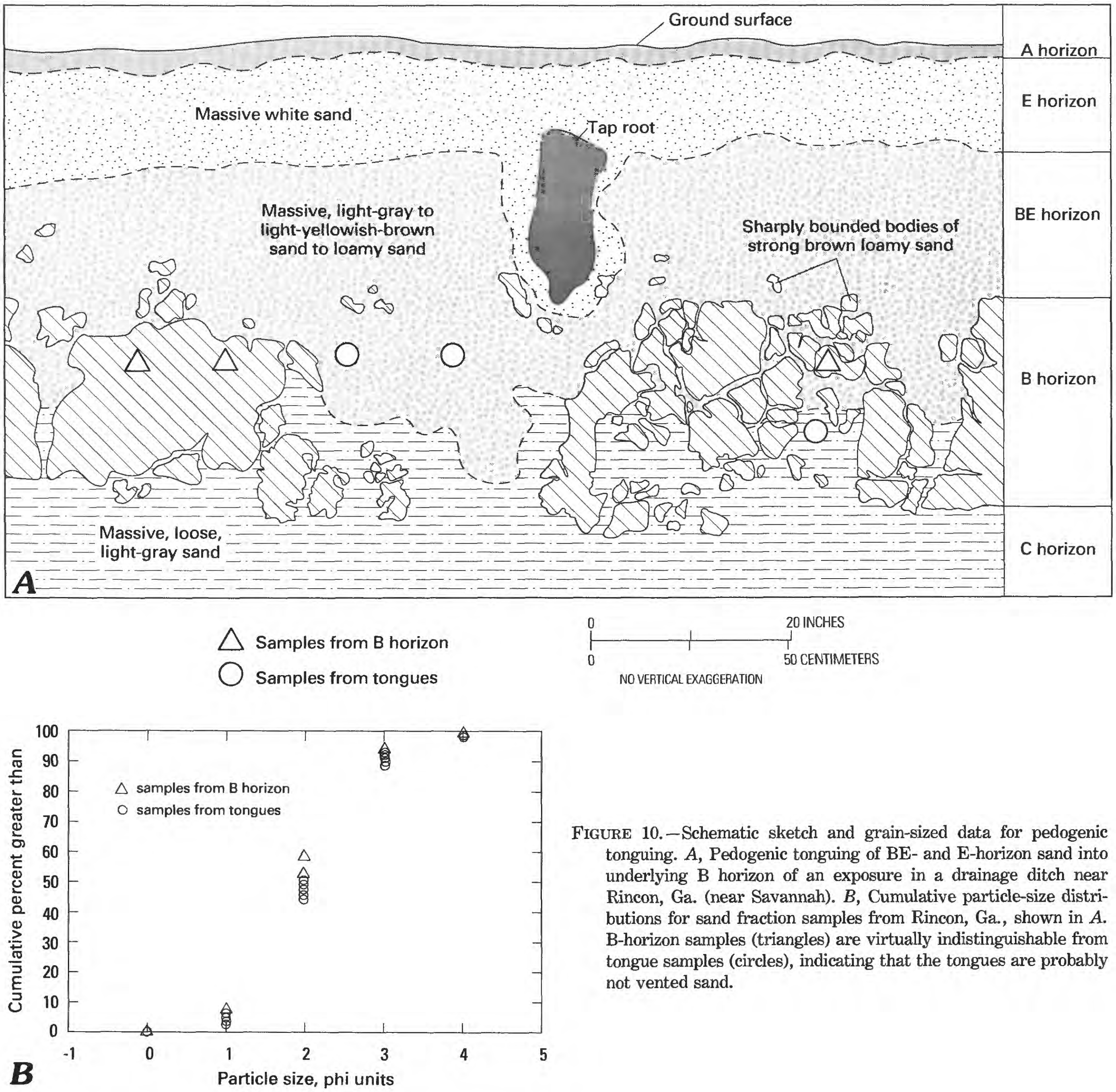

FIGURE 10.—Schematic sketch and grain-sized data for pedogenic tonguing. *A*, Pedogenic tonguing of BE- and E-horizon sand into underlying B horizon of an exposure in a drainage ditch near Rincon, Ga. (near Savannah). *B*, Cumulative particle-size distributions for sand fraction samples from Rincon, Ga., shown in *A*. B-horizon samples (triangles) are virtually indistinguishable from tongue samples (circles), indicating that the tongues are probably not vented sand.

The single field criterion that is most useful for distinguishing pedogenic tongues from liquefaction features is the downward closing of the narrow, nearly vertical sand bodies. Pedogenic tongues almost always close downward. Alternatively, if they connect at depth with a source bed, an earthquake-induced liquefaction origin is possible. Unfortunately, many exposures are too shallow to allow use of this criterion.

Pedogenic tongues can range in morphology from tubular (Gamble, 1965) to planar (defining B horizon polygons; Nettleton and others, 1968a). Hence, tongue morphology is tenuous evidence for pedogenic or earthquake origin.

An example of pedogenic tonguing is shown in figure 10*A*, which is a schematic drawing from an exposure in a drainage ditch near Rincon, Ga. The tongues extend

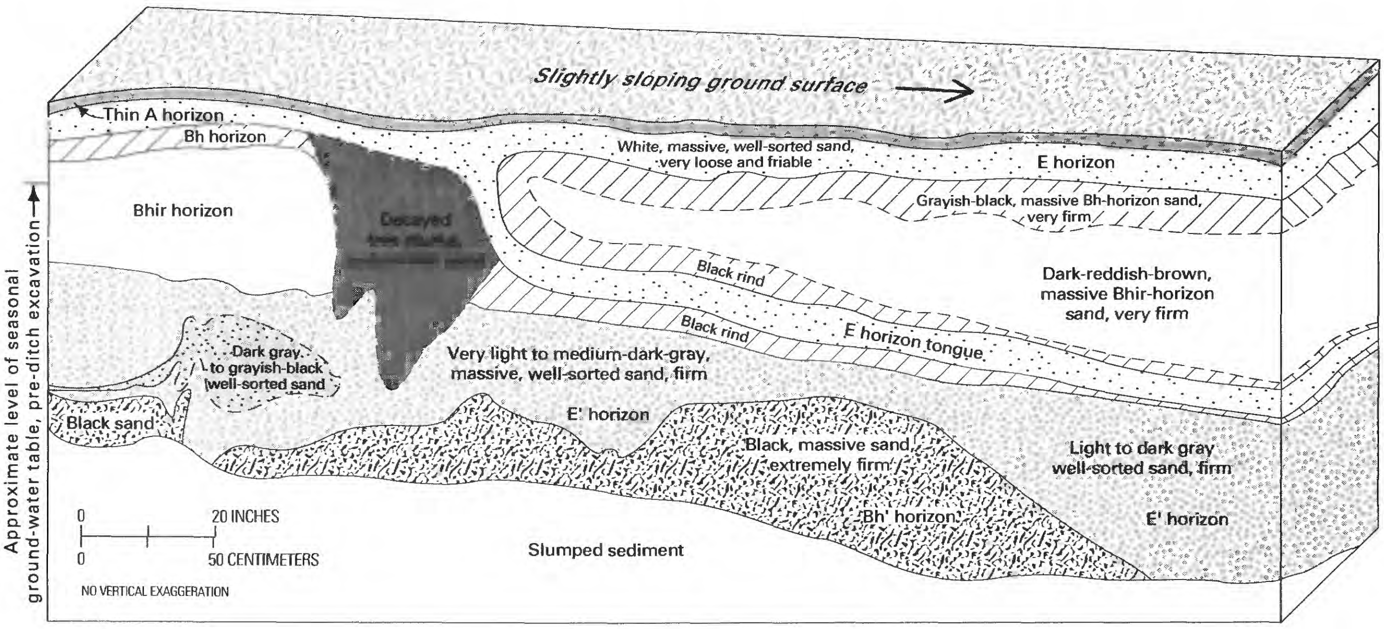

FIGURE 11.—Cross section through a white, pedogenic sand tongue from an exposure in a drainage ditch at site HW, near Hollywood, S.C. Prime designations on E and Bh horizons indicate the lower horizons of a bisequal weathering profile. Bhir is iron-enriched Bh material.

downward from a massive, pale-brown to gray E horizon and narrow with depth. Tongues may form preferentially along soil structure polygon boundaries (ped boundaries) or where tap roots have penetrated the B horizon. Once formed, tongues become the loci of concentrated weathering because the higher permeability of the leached E-horizon sand guides infiltrating solutions through the clayey B horizon. Increased weathering in the tongues may alter the particle-size distribution of the original sediment either by producing or destroying minerals in the finer particle-size fractions. Therefore, a comparison of the particle-size gradation exclusive of the clay and silt fractions (less than 63 micron) may show whether the tongue sand has been injected from a separate source. Sand size and cumulative particle-size distributions of six samples taken from areas marked in figure 10*A* are shown in figure 10*B*. The extreme similarity between the tongue and the nontongue portions strongly suggests that the tongues were weathered in place, although there is a remote possibility that injected sand could have the same sand-particle-size distribution. Alternatively, a liquefaction origin would be more strongly suggested if the tongue sand-particle-size distribution was distinctly different from that of the adjacent soil.

Other evidence for pedogenic versus liquefaction origin of tonguelike features is the mineralogy of the sand fraction. Because weathering is concentrated in the E-horizon tongues, the presence of abundant fresh, labile minerals, relative to adjacent soil mineralogy, would be good evidence for injection. Conversely, if the tongue and adjacent soil have similar mineralogies, or if the tongue has more highly weathered minerals than the adjacent soil, a pedogenic origin is more probable. By using mineralogic criteria, Gelinas (1986) assigned a liquefaction origin to sand vents at the McL, SAN, and CH sites (fig. 1) by demonstrating that suspected vent sands have greater abundances of easily weathered feldspars and hornblendes than the adjacent undisturbed soil.

Another type of pedogenic tonguing is sketched in figure 11 from an exposure in a drainage ditch near site HW (fig. 1), where the ditch crosses the crest of a relatively well drained beach ridge. At least 10 features of this size and orientation were exposed in a 60-m section of the ditch. In vertical exposures, 10- to 30-cm-thick tongues of white sand dip gently downslope. The tongues are surrounded by black to dark-orange, humate and iron-oxide-stained sand. Some of the near-horizontal parts of the tongues can be traced for distances up to 7 m. At the upslope end of each tongue, the feature abruptly turns up, breaks through the overlying Bh soil horizon, and flares toward the surface. At the surface, the tongue is continuous with white sand of the E horizon. The downslope end of the tongue is commonly terminated by black-and-orange-stained rind that is continuous with the Bh horizon; in other cases the tongue-rind contact becomes diffuse downslope.

The white sand tongue features were originally interpreted by us as intrusions of liquified sand (Gohn and others, 1984, p. 11); however, subsequent observations have led to the conclusion that the tongues are weathering features unrelated to liquefaction events. Essentially, they are extreme convolutions in the pedogenic E horizon caused by concentrated infiltration of soil solutions at places where taproots of trees have broken through brittle, relatively impermeable Bh horizons. This interpretation relies on several field observations:

- The tongues are continuous with white E-horizon sand overlying the Bh horizon. Pedogenic origin of the E horizon sand is established by satisfaction of the sedimentologic and mineralogic criteria discussed previously.
- The rind surrounding the white tongues is zoned into black Bh and dark-orange, iron-enriched Bh material (Bhir) stripes identical to the sequence found at the top of undisturbed B horizons. The vertical sequence of horizons in an undisturbed soil profile shows that dark organic compounds are deposited closer to the surface than iron oxyhydroxides after solutions infiltrate through the overlying E horizon. A similar zonation in the rind surrounding the tongues suggests that soil solutions are migrating laterally along the tongues and outward toward their margins.
- The sharp to diffuse contacts between the tongues and surrounding sands show no appreciable sand-grain-size variation across them. The only difference between the tongues and the rinds is that the tongues contain black and orange colloidal material that bridges and coats sand grains.
- No bedding is present in the tongues.
- The flared portion of each tongue (the portion closest to the surface) is characterized by evidence of a taproot, which is either a rotted root in-place or a narrow V-shaped zone of loose, organic-stained sand subjacent to the white sand tongue. E-horizon tongues were found in all stages of development, from incipient leaching and bleaching surrounding areas adjacent to relatively fresh taproots to wider and more deeply bleached zones that formed as the taproots became more decayed.
- The tongues all dip downhill in the direction of the low-gradient, shallow ground-water flow system. Mottles around rootlets in the C horizon are also elongated in downhill directions. Coincidence of the tongue orientation with ground-water flow direction suggests that after the infiltrating solutions break through the Bh horizon where taproots have decayed, the infiltrating solutions are entrained in the shallow ground-water flow system. Dissolved organics and iron oxyhydroxides in the ground water are then deposited downgradient as a rind adjacent to the E-horizon tongues.

Another category of pedogenic feature that might be confused with earthquake-induced liquefaction is the BE or BE′ horizon. (The prime designation is used to denote the lower sequence of eluvial and illuvial horizons in a bisequal profile, in the general order A, E, Bh, E′, BE′, B (t,x), C.) Bisequal profiles with BE′ horizons form at the transition between poorly drained and moderately well drained landscape positions (Nettleton and others, 1968a,b). BE or BE′ horizons are transitional between eluviated E horizons and illuvial B, Bt (argillic), or Bx (fragipan) horizons and are characterized by irregular to prismatic bodies of clay-rich sand surrounded by leached, clean sand. Studies of this type of weathering profile in the North Carolina coastal plain indicate that the clean sand forms from the progressive destruction of a clay-rich Bt horizon (Daniels and others, 1966; Nettleton and others, 1968a, b; Steele and others, 1969). In this process, a Bt horizon is leached at the depth of a fluctuating water table, which removes clay and labile minerals and leaves patches of white to gray quartz sand that often coalesce to define a polygonal pattern in plan view. This horizon, composed of patches of leached and unleached material, is termed an E′ or BE′ horizon, depending on the extent of leaching. With continued leaching of clay and labile minerals, the soil volume decreases, resulting in collapse and the formation of a dense Bx horizon.

In the ditch at the HW site, bisequal soils occur between red, oxidized spodosols on beach ridge crests and black, organic-rich spodosols in interridge depressions. For the example shown in figure 12, the BE′ horizon is a layer of pale-brown, clayey sand and white sand. The top of the layer is 1 to 2 m beneath the ground surface, and the layer is 0.3 to 0.5 m thick. Both above and below the layer is light-colored, well-sorted, fine- to medium-grained quartz sand. This sand cuts vertically across the clay layer at many places and creates irregular masses of the brown clayey sand. In three dimensions it can be seen that the sand forms irregular vertical walls. In vertical section, the sand walls are typically about 5 to 10 cm wide but can be as much as 30 cm wide. They are spaced at both irregular and at regular intervals. The contacts of the sand and the brown clayey sand are typically sharp with regard to both color and texture. The sand-grain sizes in both the leached sand and clayey sand masses are essentially the same.

Sharp contacts between the leached quartz sand areas and the pale-gray to pale-brown clayey material originally suggested that this feature was formed by some type of ground disruption, possibly liquefaction. However, the similarity with features attributed to pedogenesis elsewhere and the consistent occurrence in horizons underlying undisturbed pedogenic horizons led us to interpret these features as pedogenic.

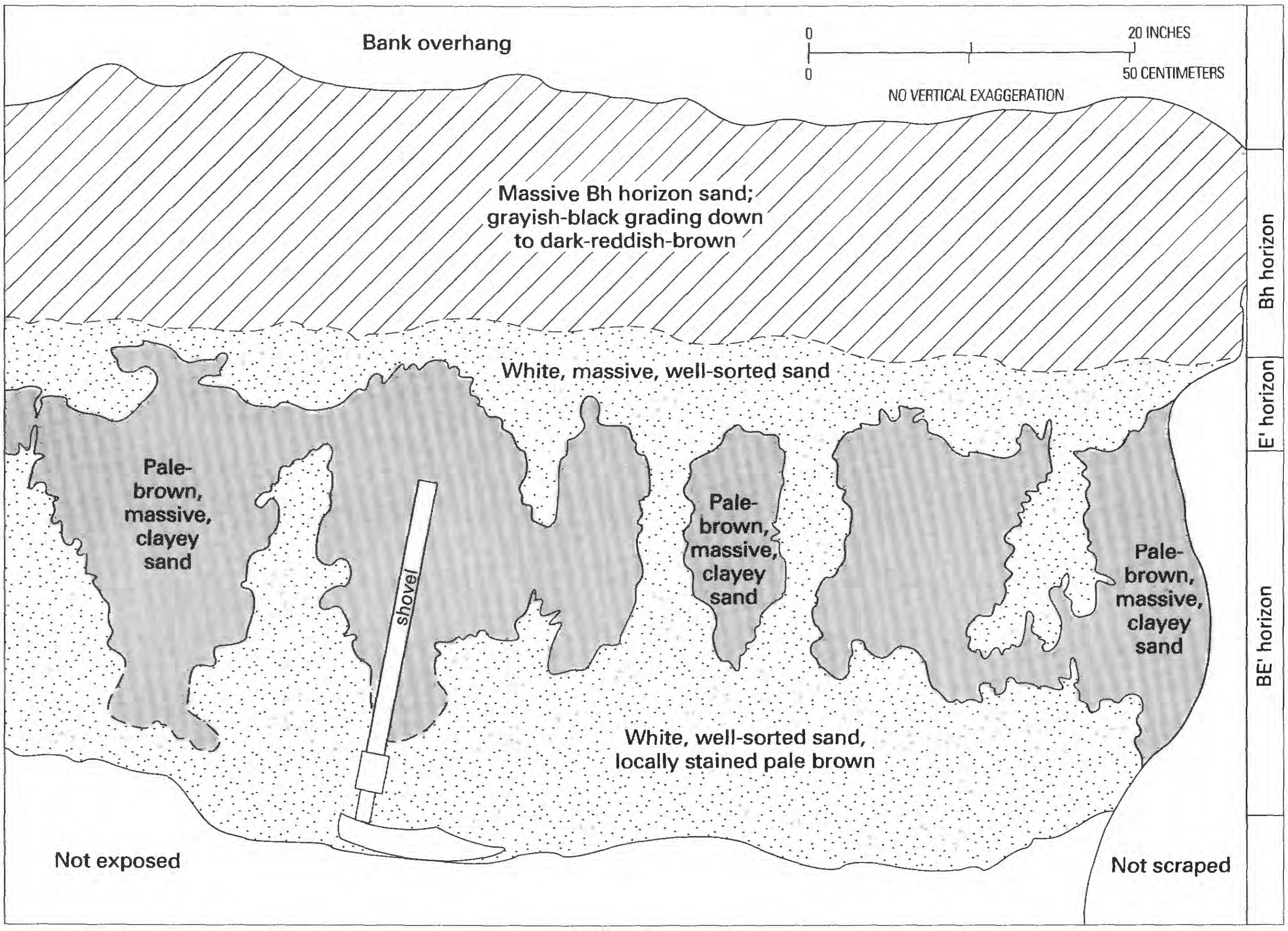

FIGURE 12.—Vertical section through BE′ horizon at site HW, near Hollywood, S.C.

OVERVIEW OF THE SOUTH CAROLINA LIQUEFACTION STUDY

Pre-1886 sand blows were first discovered in 1983 (Gohn and others, 1984; Obermeier and others, 1985) near the town of Hollywood (site HW, fig 1). There, the sand blows are exposed in an unusually deep ditch (up to 3 m) for a distance of 5 km. Over a hundred sand blows have been discovered in this ditch, and the depth of the ditch has permitted detailed study of entire features to depths that include source vents. For these reasons, site HW is generally used for the descriptions of filled craters in this paper. Since the discovery at Hollywood, features we interpret as sand blows have been found at sites throughout much of coastal South Carolina.

Features at all sites shown on figure 1 are interpreted to be of earthquake origin, although the confidence level differs for various sites. Sites where we have greatest confidence of an earthquake origin are those where the following features occur: (1) craters have formed on topographically high beach crests, (2) numerous craters are present near where craters were reported to have been abundant during the 1886 earthquake, (3) ground oscillation shears have formed in opposite directions, (4) lateral spreads that could not be gravity-induced have formed and have shears traceable into a liquefied stratum, or (5) shattered ground is cut by numerous sand-filled dikes in settings where high artesian pressures could not have been involved. Sites for which we have the highest confidence of an earthquake origin are BR, AR, HW, ARP, RRR, CH, FM, WV, SAN, OL, and SOPO.

All other sites on figure 1 are filled craters that are more than several meters below the beach crest, causing the confidence level to be lower. Some craters at site MYRB have reverse shears, however, an occurrence that makes an earthquake origin seem likely.

Elimination of all sites from figure 1 except those in which we have the highest confidence does not affect our interpretation of Holocene seismic activity (discussed subsequently). Our interpretation about earthquake ages and relative severity of shaking is based only on data from the sites for which we are most confident of an earthquake origin.

EARTHQUAKE AGES

Craters are typically the only features for which radiocarbon ages related to earthquake ages can be generated, because other liquefaction-related features are not found in association with preserved organic matter. Three methods have been used to bracket the times of crater formation (Weems and others, 1986): (1) radiocarbon ages of woody material (tree limbs or pine bark) that fell into the open crater soon after crater formation, (2) dating of roots sheared off at the edge of the crater (predating crater formation) and dating of roots that grew into the stratified fill portion of the crater (postdating crater formation), and (3) dating of clasts of Bh material that fell into the graded fill zone of the crater. The first method yields a highly accurate age for the time of earthquake occurrence, whereas the other two yield a broad range of possible ages. Sufficient data have been collected at site HW (near Charleston; fig. 1) to show that at least three pre-1886 earthquakes produced sand blows within the past 7,200 years. Radiocarbon dating of a clast of pine bark in a crater at site ARP (also near Charleston) independently verifies the middle of these three events. The only definitive statement about earthquake recurrence that can presently be made is that near Charleston there have been at least four sand-blow-producing (m_b probably >5.5, discussed later) earthquakes within the past 7,200 years (including the 1886 event). Accurate ages of crater formation have been obtained from some sites more than 100 km from Charleston. These dates differ from ages near Charleston, thereby suggesting that the craters far from Charleston originated from epicentral regions also far from Charleston. At many sites far from Charleston, there are at least two generations of craters that are long separated in time of formation.

SHAKING SEVERITY ESTIMATION

Insufficient radiocarbon ages have been collected from liquefaction features throughout the Carolina coastal region to define epicentral regions of separate earthquakes. Adequate data have been collected, however, to estimate the relative shaking severity throughout the coastal region during the Holocene. This estimate is provided by measurement of the number and size of craters at the sites shown on figure 1.

The methodology for estimating shaking intensity is based on the premise that the number and size of liquefaction features are greatest where earthquake shaking has been strongest for a fixed geologic setting and liquefaction susceptibility. The condition of a fixed geologic setting is met almost ideally on many of the Pleistocene beaches, as discussed in an earlier section.

The condition of a fixed liquefaction susceptibility is also almost certainly satisfied at the widely scattered sites on figure 1. Source sands typically are loose (based on limited Standard Penetration Test data and numerous observations of ease of augering) and have about the same thickness. Moreover, the thickness and properties of nonliquefiable sediments overlying the source stratum lie within a narrow range. It is also a certainty that recurrences of liquefaction do not greatly diminish the potential for formation of more large craters in loose sands. This statement is verified by the observation that at site HW there are many large craters that formed in each of at least three generations of Holocene earthquakes, with each generation widely spaced in time. Thus, at sites in beach deposits on figure 1, liquefaction susceptibility is generally high and has not been greatly reduced by previous occurrences of liquefaction.

The other major variable, depth to the water table, is about the same from site to site (very shallow depth) and probably has been shallow throughout the Holocene (Obermeier and others, 1987). Evidence for location of the water table throughout the Holocene is provided by location of the base of the Bh horizon. (See fig. 4 for location of this horizon at a typical filled crater.) The maximum depth of the seasonal water table during the Holocene is marked very nearly by the base of the Bh horizon. (The Bh is defined as the subsoil zone of accumulation of organic matter and is formed in these soils at the lower limit of vertical infiltration of water.) Throughout the coastal region, the base of the Bh (generally 0.6 to 1 m below land surface) is nearly coincident with the present-day water table. Radiocarbon ages from the basal Bh horizon are 5,000 to 10,000 years at site HW (Weems and others, 1986, p. 7). Because these ages are mean residence times of organic matter in a dynamic system characterized by continuing vertical infiltration of younger organic matter, some of the organic matter has been there even longer. Thus, it can be concluded that the water table has been very shallow throughout the Holocene over wide areas of the South Carolina Coastal Plain.

HOLOCENE EARTHQUAKE SHAKING

Both the abundance and diameters of pre-1886 Holocene craters are greatest within the 1886 meizoseismal zone for a given age of craters. On the basis of these criteria, we know that shaking has been much weaker north of the Santee River (Obermeier and others, 1989). Intermediate shaking has taken place between Charleston and the Santee River and also between Beaufort and the Savannah River.

Confidence in this interpretation is high for the area between Charleston and Wilmington, because of the hundreds of kilometers of ditches we searched. Our confidence is also high for the 1886 meizoseismal zone and for the area between Beaufort and the Savannah River. Our confidence is not nearly as high for the area between the Beaufort and the Edisto River nor for the area south and southeast of the 1886 meizoseismal zone; this lower confidence is caused by the limited number of ditches and pits available for inspection.

Whether or not the pre-1886 Holocene shaking in the 1886 meizoseismal zone is associable with earthquakes stronger than the 1886 event can be determined only by additional radiocarbon ages for craters at sites far beyond the 1886 meizoseismal zone.

Based on worldwide observations in the field, the minimum earthquake strength required for liquefaction-induced features is m_b equal to approximately 5; such features are rare for m_b less than 5.25 to 5.5 (Carter and Seed, 1988). It is likely that an earthquake slightly stronger than about 5.5 is adequate to produce numerous small liquefaction features in coastal South Carolina because of the exceptional liquefaction susceptibility of many marine sand deposits there (discussed previously and discussed in Dickenson and others, 1988).

Many of the pre-1886 craters near Charleston that we have observed probably were not produced by earthquakes as small as m_b 5.5, however. Many pre-1886 craters are large (diameters of as much as 3–4 m are not uncommon) in comparison to historical descriptions of 1886 craters (Dutton, 1889) and our observations of the sizes of 1886 craters. As a result, we suspect that some of the pre-1886 craters were formed by earthquake shaking that was stronger at those sites than in 1886. Direct comparison of 1886 and pre-1886 crater sizes cannot be used for estimating earthquake magnitudes, however, because of the lack of knowledge about distances between craters and their associated epicentral regions and lack of knowledge about changes in liquefaction susceptibility caused by a previous occurrence of liquefaction. About the only comment that can be made with much assurance is that the prehistoric craters were very likely the result of earthquakes stronger than m_b 5.5; some of the earthquakes probably were much stronger because of the widespread distribution of large craters.

NEW MADRID SEISMIC ZONE

The succession of great earthquakes collectively designated as the New Madrid earthquakes of 1811–12 caused severe and widespread ground failure throughout a large area near the Mississippi River in southeastern Missouri, northeastern Arkansas, western Kentucky, and western Tennessee. The earthquakes caused multitudes of fissures and sand blows (Fuller, 1912), set off numerous landslides (Fuller, 1912; Jibson, 1985), and caused localized doming and submergence of the ground surface (Russ, 1982). Earthquake effects were particularly severe in the St. Francis Basin (fig. 13). According to the earliest extensive documentary account (Fuller, 1912), three major shocks occurred: December 16, 1811; January 23, 1812; and February 7, 1812. The surface-wave magnitudes (M_s) of these earthquakes are estimated to have been between 8.3 and 8.8 (Nuttli, 1983a). Estimated Modified Mercalli (MM) intensities are XI to XII throughout large areas near the epicenters (Nuttli, 1973); these intensities are regional values based largely on historical accounts of liquefaction-induced ground failure. No faults associated with the 1811–12 earthquakes have been found that cut through alluvium, loess, or older strata to the surface. The epicenters are estimated to lie within the large area of sand blows near the Mississippi River, shown on figure 14. This interpretation of epicenters is based on the study by Fuller; a study of MM intensities by Nuttli (1973); recent geologic, geophysical and modern seismicity studies (McKeown and Pakiser, 1982; Crone and others, 1985); and engineering-geologic studies of factors controlling the distribution of sand blows (Obermeier, 1989). (The epicentral regions during the 1811–12 earthquake and the region of epicenters on figure 14 define the "New Madrid seismic zone.")

Figure 14 is a map by Obermeier (1988) that shows the percent of the ground surface covered by sand vented to the surface in the St. Francis Basin. The map is the latest of a series of maps (Fuller, 1912; Saucier, 1977; Heyl and McKeown, 1978) showing sand blows throughout the alluvial lowlands of the basin. Significant differences appear between each of the maps, particularly north of the town of New Madrid. These differences occur for a number of reasons, the most likely being that (1) some surface soils are so sandy that sand blows can be observed only on the older aerial photographs used by Obermeier, (2) the modern farming practice of land-leveling has destroyed many sand blows since the 1950's, (3) some nonearthquake features cannot be distinguished from sand blows except by field excavations, and (4) the map by Obermeier extends more to the limits of ground covered by vented sand, whereas the earlier maps emphasize areas of abundant sand blows.

Sand-blow deposits and other manifestations of liquefaction-induced flowage occurred far beyond the limits of sand-blow deposits shown on figure 14, but beyond these limits the deposits were generally

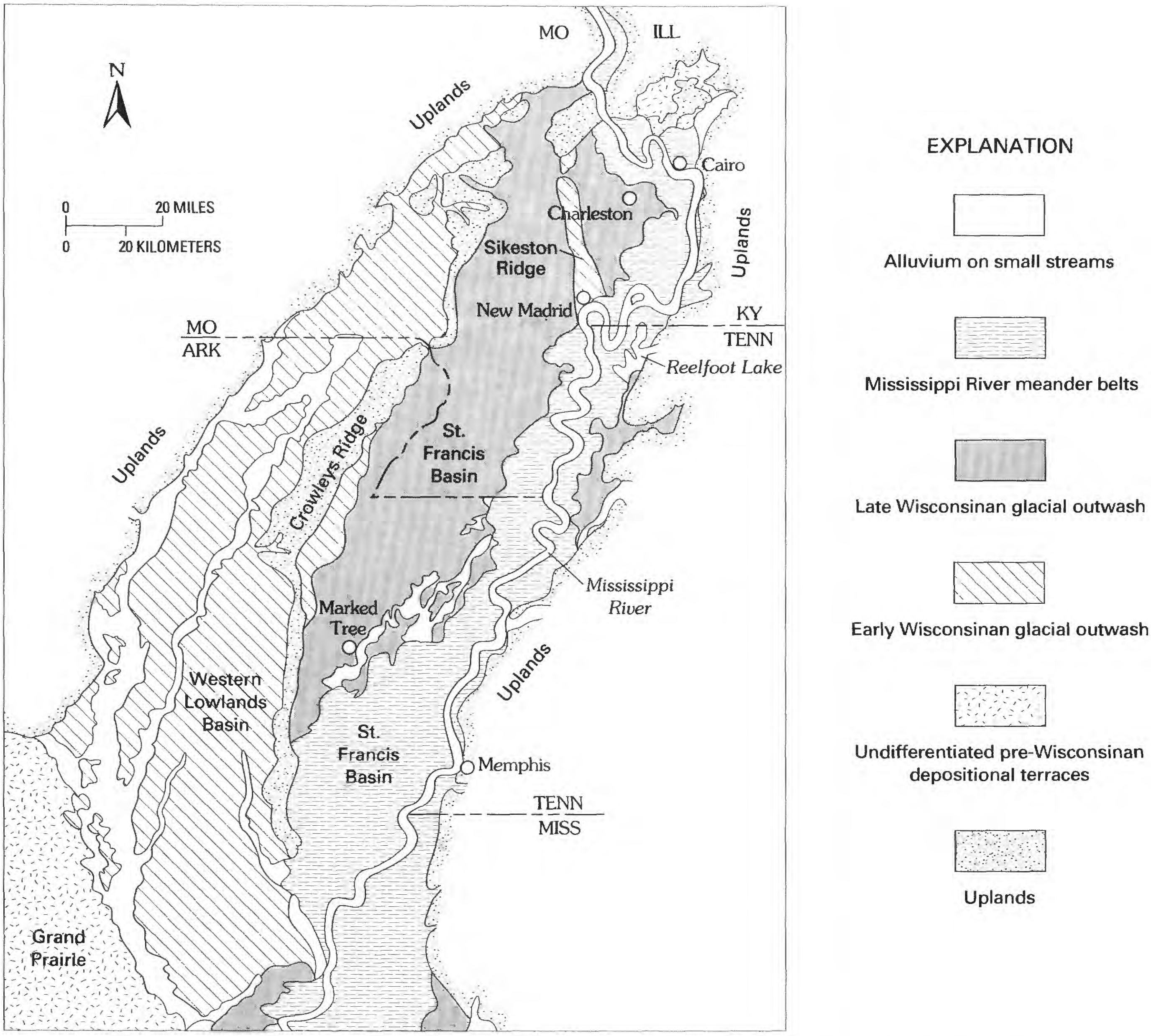

FIGURE 13.—Late Quaternary alluvial deposits of St. Francis and Western Lowlands Basins (from Saucier, 1974).

restricted to modern flood-plain sediments adjoining rivers. These very young (late Holocene) sediments generally have higher liquefaction susceptibility than the late Wisconsinan and early Holocene fluvial deposits, which underlie the area of sand-blow deposits shown on figure 14. Liquefaction features were especially commonplace in alluvium along some of the small streams of the Western Lowlands Basin (fig. 13), particularly those nearest Crowleys Ridge. Sand blows were reported (McDermott, 1949) in alluvial deposits as far north as Cahokia, Ill. (which is very near to St. Louis, Mo.), and as far northeast as the Wabash River valley (Street and Nuttli, 1984). Some possible sand blows were formed in some of the river valleys beyond the limits of sand blows on figure 14. For example, Dickey (1985) reports features on aerial photographs in the Western Lowlands that appear to be sand-blow deposits. At many places, though, sand-blow deposits on modern flood plains have been covered by a veneer of alluvium deposited since the 1811–12 earthquakes.

In the New Madrid area, visible sand-blow deposits are still plentiful on the ground surface, and the epicentral regions are relatively well established. These characteristics permit a more detailed evaluation of geologic

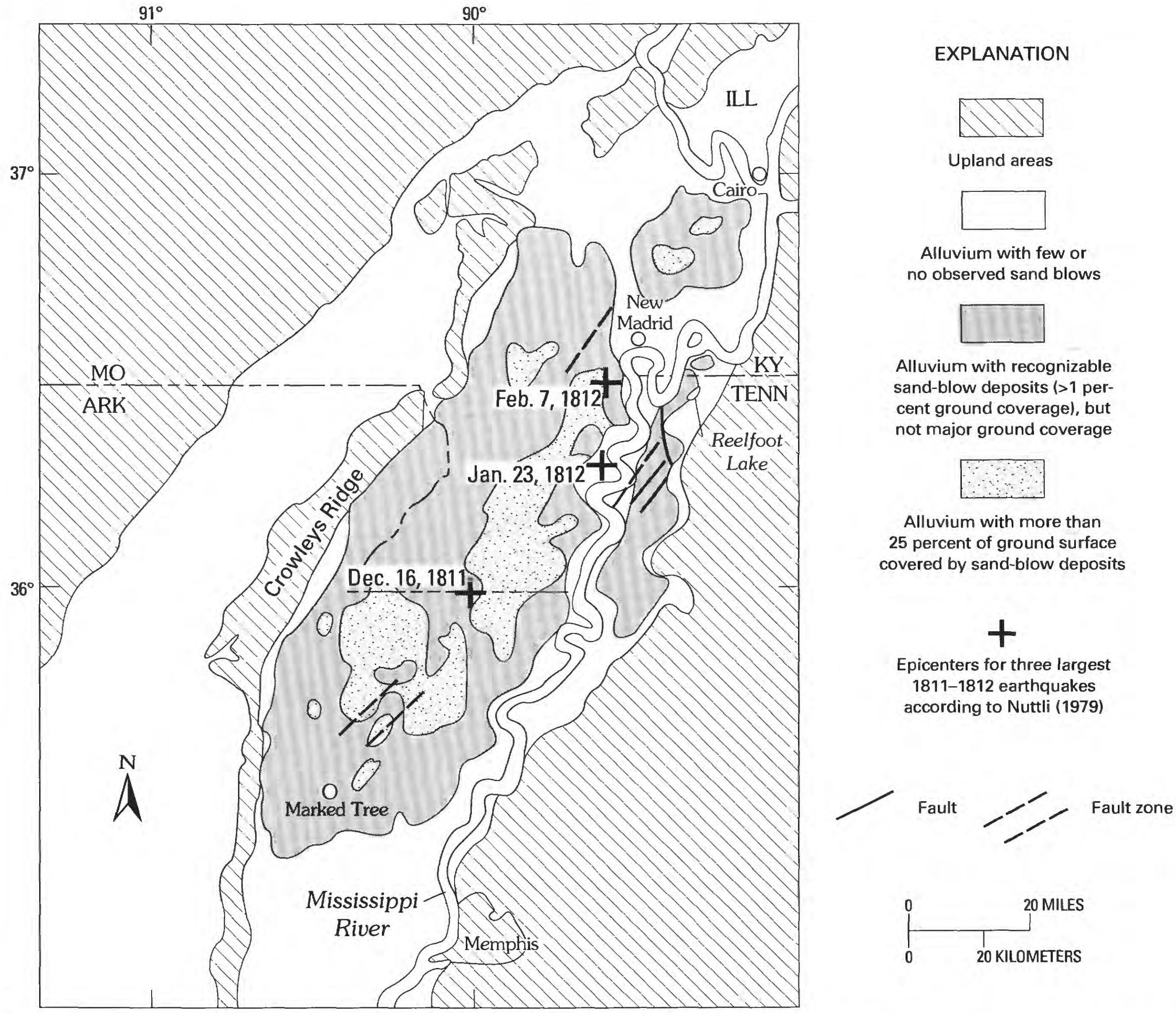

FIGURE 14.—Map showing area covered by vented sand (Obermeier, 1988), estimated epicenters of strongest 1811–12 earthquakes (from Nuttli, 1979), and faults and fault zones (from Hamilton and Zoback, 1982).

controls on the formation of liquefaction features than is possible in coastal South Carolina.

REGIONAL GEOLOGIC AND SEISMOTECTONIC SETTING

The St. Francis Basin is a topographic lowland containing 30 to 60 m of late Quaternary alluvium, which originated from deposition associated with the Mississippi and Ohio Rivers and their tributaries. The lowlands are typically very flat, having about 0.3 to 2 m of local relief. Numerous small scarps, 2 to 6 m high, result from depositional or erosional processes. The water table is very high and is within 1 to 2 m of the ground surface at many places. Elevations in the lowlands exceed 100 m near Cairo, Ill., and decrease southward to less than 60 m near Memphis, Tenn.

Widespread floods were common prior to the construction in this century of manmade levees along the Mississippi River. Standing water occupied the lower parts of the lowlands throughout much of the year before the levees were built and large drainage ditches were excavated. Until about 50 years ago, dense forests and cypress swamps covered most of the basin; today the forests have been replaced by large fields of soybeans, rice, and other row crops.

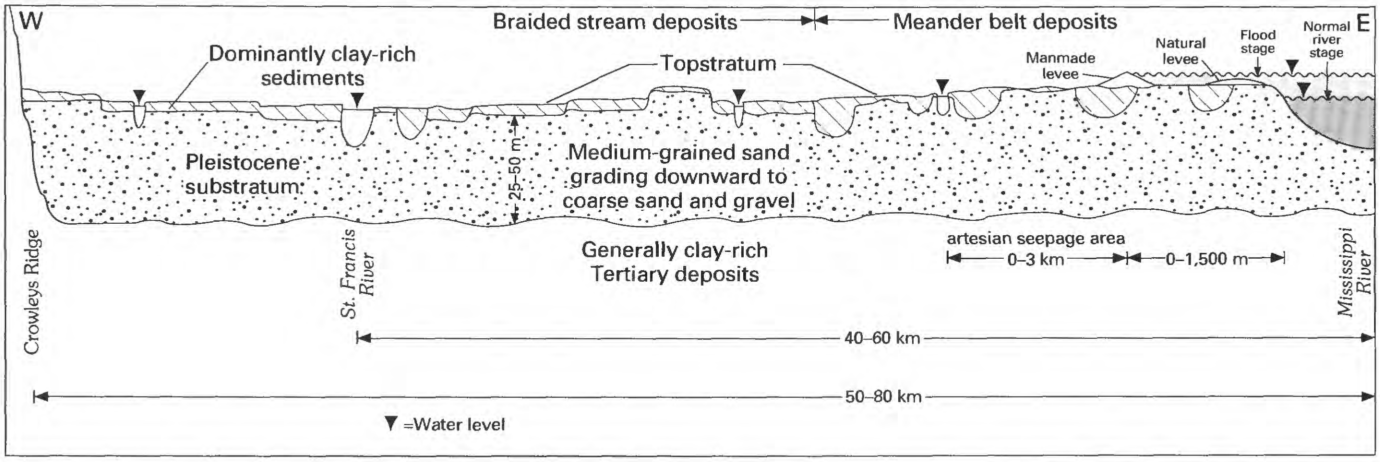

FIGURE 15.—Schematic east-west cross section showing geologic and ground-water setting of St. Francis Basin.

CHARACTERISTICS OF QUATERNARY ALLUVIUM

Cyclic Pleistocene glaciations directly and indirectly controlled the origin, character, and distribution of virtually all the Pleistocene deposits in the basin (Saucier, 1974). Although continental glaciers did not extend into the St. Francis Basin, they supplied large volumes of glacial meltwater and outwash to southward-flowing river systems. Up to 60 m of valley fill composed of very coarse grained, well-graded (engineering sense) glacial outwash (gravel and sand) were first deposited by aggrading braided streams. After maximum aggradation, numerous braided-stream terraces were deposited. The ancestral Mississippi River changed from a braided to a meandering regimen in early Holocene time, and since that time, slow aggradation has taken place in most parts of the valley. Most meander-belt deposits are medium-grained, well-graded sand; locally, silts and clays have been laid down.

Sikeston Ridge (fig. 13), the highest terrace of St. Francis Basin, is early Wisconsinan in age, while the other braided-stream terraces are late Wisconsinan. The earlier braided-stream deposits are topographically higher, have greater relief, and have sandier surfaces than younger deposits. The older braided-stream deposits occur in several terrace sublevels separated by 2 to 6 m. These deposits generally are capped with 3 to 7 m of silty or clayey sediments (the topstratum) containing some thin sand strata. The topstratum abruptly overlies the clean, outwash sands and gravels (the substratum) over large areas except at the highest sublevels and interfluves. The topstratum was deposited either by relatively slack streams as individual braided-stream levels were successively abandoned or by overbank deposition during widespread flooding of the Mississippi River and local streams that now occupy the topographic lows. Substratum sands and gravels generally are found in intercalated strata and lenses ranging from a few centimeters to a meter thick and tend to become coarser with depth.

The Mississippi River meander belts represent successive courses formed by lateral migration of the river. Most of the meander belt consists of point-bar "accretion" topography of parallel arcuate ridges and swales, abandoned channels in various stages of filling, and natural levees. Point-bar deposits are generally clean, well-graded sands[4] that are capped by the topstratum except locally at the highest elevations. Point-bar deposits of the Mississippi River typically have 2 to 3 m of local relief. Many abandoned channels are filled with 30 m or more of soft clays and silts and are swampy and densely forested.

Large backswamp areas are found along the margins of some Mississippi River meander belts (Saucier, 1974). Backswamp deposits were formed by seasonal floodwaters depositing silts and clays in low areas of the flood plains during the Holocene. The deposits average 12 m in thickness but are as much as 18 m thick.

The ground-water setting is generally simple, as shown in figure 15. Major rivers such as the Mississippi and St. Francis both supply and drain large volumes of water in the substratum. Locally, especially within 0.5 to 1 km adjacent to levees, artesian conditions occur many years during flooding. Beyond this area of flood-related springs near levees, the streams have cut through the topstratum, thus preventing artesian conditions. In addition, throughout the St. Francis Basin many topographically elevated areas make artesian conditions impossible. Present ground-water conditions are shown in a fairly detailed manner in a ground-water report by

[4]In this paper, the clean sands beneath the fine-grained cap are referred to as the "substratum," and the fine-grained cap as the "topstratum," irrespective of origin as meander-belt or braided-stream deposits.

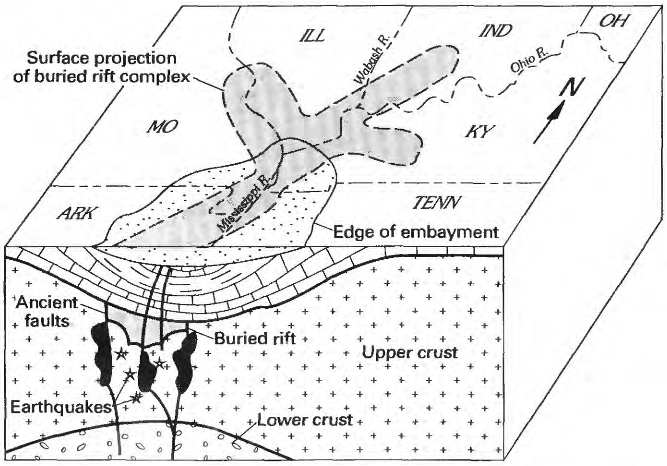

FIGURE 16.—Configuration of the buried New Madrid Rift Complex (from Graile and others, 1984). The structurally controlled rivers, Paleozoic rocks in the cratonic sedimentary basins, and the Mississippi Embayment, all associated with the buried rift complex, are also shown. Dark areas indicate intrusions near the edge of the buried rift.

Krinitzsky and Wire (1964); detailed geologic maps by Saucier (1964) and Smith and Saucier (1971) can be used to help determine present and Holocene ground-water conditions. No evidence indicates that ground-water conditions (especially the piezometric surface) were significantly different during the past few thousand years in the St. Francis Basin, except locally near manmade levees during times of flooding.

SEISMOTECTONIC SETTING

The St. Francis Basin lies in the northern Mississippi Embayment, which is a broad, southward-plunging syncline (fig. 16). The northern embayment is underlain by a late Precambrian intraplate rift that has been active periodically since its formation (Braile and others, 1984). Within this rift lies the New Madrid seismic zone. This zone is the most seismically active area in the United States east of the Rocky Mountains. Continuing fault movement, in response to regional compressive and perhaps thermal stresses, is the most likely cause of modern seismicity (Braile and others, 1984).

Since 1812, at least 20 earthquakes having estimated body-wave magnitudes between 3.8 and 6.2 have occurred in the region (Nuttli, 1982). Most estimates for the return period of great earthquakes are between 500 and 700 years (Hopper and others, 1983).

Figure 17 shows locations of instrumentally recorded earthquakes and principal structural elements in the New Madrid seismic zone. The trend of modern epicenters extending from Marked Tree, Ark., to Ridgley, Tenn., coincides at least approximately with the epicenters interpreted by Nuttli (1979) of the three principal shocks of the 1811–12 earthquakes. Subsurface faults are inferred by seismic-reflection data to lie beneath the most seismically active areas (O'Connell and others, 1982; Crone and others, 1985).

CHARACTERISTICS OF AND CRITERIA FOR EARTHQUAKE-INDUCED LIQUEFACTION FEATURES

Reports of level-ground and near-level ground failure features made shortly after the 1811–12 earthquakes noted great multitudes of vented-sand volcanoes, linear fissures up to 6 m deep and hundreds of meters long, craters many meters in diameter, and lateral spreads hundreds of meters long (for example, see Penick (1976)). Fuller (1912) described vicinities where these features were especially abundant. Individual vented-sand volcanoes and some long linear fissures through which sand vented are the only features that are still readily visible on the ground surface. Intruded dikes and sills are common in walls of deep (>3 to 4 m) drainage ditches in the clay-rich topstratum. We have not found other types of level-ground failure features such as deep craters or lateral spreads having downdropped blocks at the head such as those described by Fuller (1912, p. 48), primarily because we have not searched extensively in areas where these features would have formed (that is, locations in sandy topstratum for craters and locations near streams or scarps for lateral spreads). Limited observations in the New Madrid seismic zone by Wesnousky and others (1987) and by us of the characteristics of sand blows formed in a sandy topstratum indicate that open, deep craters are atypical forms. Instead, intrusions that extensively shattered the ground near the surface locally erupted to form shallow craters (up to 2 m deep).

The geologic criteria (previously discussed) for interpreting an earthquake origin with regard to specific features are identical in both the South Carolina and New Madrid earthquake areas.

VENTED-SAND VOLCANOES

The vented-sand volcanoes in the New Madrid area are similar to, though much larger than, those discovered in South Carolina. (See fig. 7.) Individual vented-sand volcanoes induced by the 1811–12 earthquakes are domelike accumulations of clean sand on the ground surface. Fuller (1912, p. 79) noted that "the normal blow is a patch of sand nearly circular in shape, from 8 to 15 feet across, and 3 to 6 inches high." Such small sand blows as Fuller described can rarely be found at present. The sand blows that are now obvious range from about 0.3 to 1.3 m in height at the center and thin to a feather-edge at a diameter of 20 to 60 m.

Most vented-sand volcanoes have a well-defined internal stratigraphy. Along the base of the vented-sand

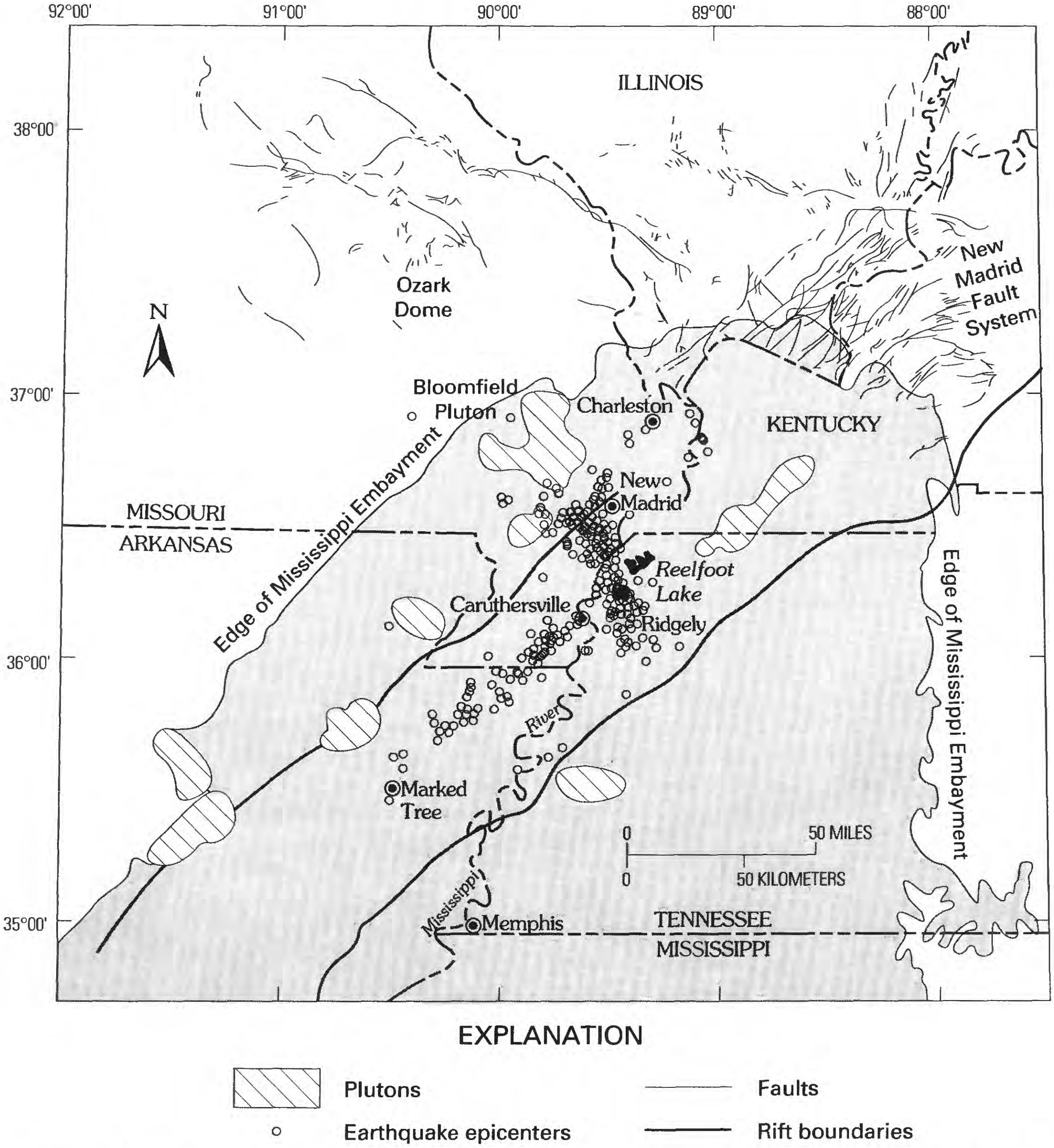

FIGURE 17.—Northern Mississippi Embayment (shaded area) showing earthquake epicenters, plutons, rift boundaries, and faults (from Zoback and others, 1980).

volcano deposits, where the vented sand spread over the ground surface, many irregular 1- to 3-cm-long clasts of topstratum clay (generally a blue-gray, highly plastic clay) are scattered throughout the sand. The clasts are largest and most plentiful closest to the vent, which is beneath the central part of the dome. The basal part of the sand-blow deposit contains numerous pieces (1–3 cm in diameter) of rounded charcoal fragments and other very low density materials vented to the surface from the substratum. This basal part contains and is overlain by a very clean, generally medium- to coarse-grained sand. This sand grades upward to a much finer sand, which is capped by a clayey, organic silt stratum, 0.5 to 4 cm thick (fig. 18); this cap may also contain multiple very thin (1-mm-thick) clay-rich layers (Saucier 1989). The organic matter in the silt is made up of small pieces of coal, lignite, and wood. The fining-upward sequence, from the basal clay clast-bearing sand to the organic silt stratum, represents the transition from the initially rapid, turbulent eruption to the final ebbing flow out of the vent.

FIGURE 18.—Stratigraphy of vented-sand volcano showing organic-rich silt between two fining-upward sequences of sand. Presence of two fining-upward sequences is inferred to represent two separate liquefaction events or reactivation during a single earthquake. Height of shovel is 40 cm.

Locally, near the vent, the beds may dip steeply toward the vent if flowage has been sufficiently slow. (Dipping beds in an excavation through vented-sand volcano deposits are shown in fig. 5 in an article by Newmann-Mahlkau (1976)). If a violent eruption has taken place, the lower and central portion of the vented-sand volcano may be sheared and disrupted. Figure 19 shows earthquake-induced "eruptive vent" in sand-blow deposits of the 1811–12 earthquakes.

A basal bed, several centimeters thick, of pea-sized clean gravel is the coarsest basal material that we have found in the sand blows. Within a sand matrix, isolated gravels up to 3 cm in diameter have been vented; these coarsest materials were vented only near the earthquake epicenters. Sand blows far from the epicenters, near the boundary of sand-blow occurrences (fig. 14), are rather small and are made up of finer grained material. Here the basal sands are often fine-grained and fine upward to a silty sand.

At least two fining-upward sequences characteristically occur in vented-sand volcanoes near epicentral regions; each sequence represents a separate occurrence of venting induced by the 1811–12 earthquake.[5] However, generally only one sequence occurs near the regional boundary of sand blows, which is shown in figure 13.

Not only are the largest sand blows near the epicentral regions, but in some places the vented sand coalesces into continuous sheets. The sheets are up to 2 m thick and at several places blanket tens of square kilometers (for example, just south of the Missouri-Arkansas boundary and near the December 16, 1811, epicenter shown on fig. 14).

Vented-sand-blow deposits are obvious on aerial photographs at many places. Figure 20 shows some examples. The light-colored spots show vented sand blows, and the light-colored linear features show fissures through which sand has vented. (Some of the linear features near streams probably mark edges of lateral spreads, but we have not confirmed this suspicion in our study, and therefore in this paper such features are simply called fissures). Figures 20*A* and 20*B* illustrate how sand blows typically occur at irregular, almost erratically spaced intervals, and the crazing pattern of vented sand suggests that the ground is fractured. Figure 20*C* illustrates that where less sand is vented to the surface, the venting is more erratic and is restricted to localized areas and that relatively small differences in site characteristics become important controls on vent locations. Comparison of figures 20*A* and 20*B* illustrates that sand-blow deposits may be more apparent in the older photographs than the more recent photographs.

Some surficial features unrelated to earthquakes have a similar appearance to earthquake-related features shown in figure 20*C*. An earthquake origin was attributed to questionable features only if the features had a concentration of irregular, large (2–3 cm) clay clasts at the base (implying a strong hydraulic force); had a fining-upward clean sand (implying diminishing flow within a short time period); were irregularly spaced (suggesting ground fracturing); and (4) were located where spring flow from uplands or from beneath levees could not have occurred. (Sand boils formed by flow beneath levees are discussed later.)

Only limited data have been collected to characterize vent shapes and sorting and flow structures in the vent filling. The vents are frequently steeply to vertically oriented and are fissure shaped. Vents generally appear to be filled with a structureless mixture of silt, sand, and clay clasts ranging from sand sized to having a length of up to 20 cm. The clasts have been transported up the vent and are derived from sidewalls and beds at depth.

FISSURES

Earthquake-induced fissures presently appear on the ground surface as lines of continuously or discontinuously vented sand. The fissures are common on level ground far from any (erosional) scarps but are especially commonplace on those parts of terraces immediately above scarps and on river banks. Vented sand locally forms small ridges, but the ridges are generally not as high as nearby vented-sand-volcano deposits and sometimes are

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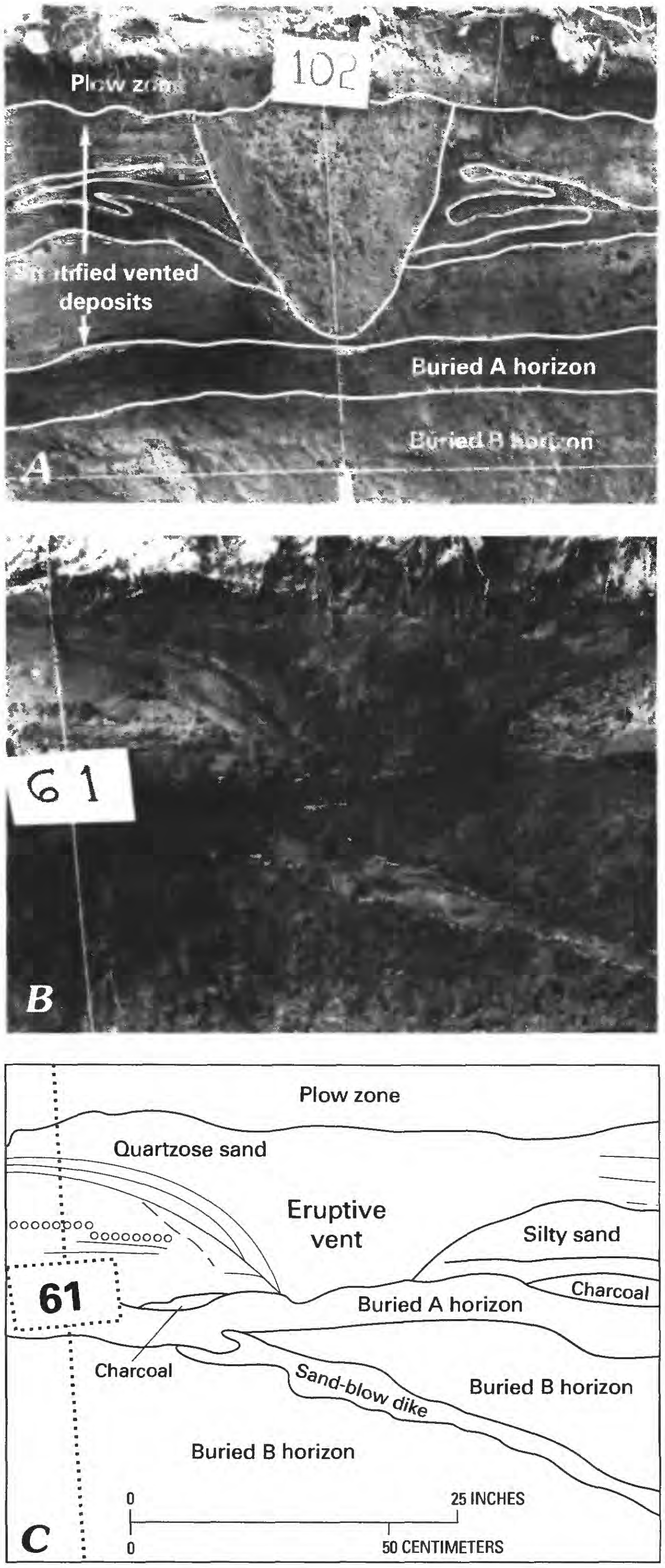
[5]Interpretation of 1811–12 earthquakes is based on the absence of a soil profile in any of the fining-upward sequences.

barely discernible. Widths of fissure openings tend to range between 0.3 m to a feather edge, most being small; the wider openings almost certainly indicate association with lateral spreads.

Fissure patterns can be highly sensitive to the local geologic setting. Fissures have a strong tendency to follow the crests of point-bar deposits; on braided-stream deposits having essentially a uniformly thick topstratum, the fissures are commonly oriented randomly (Obermeier, 1989). Topstratum thickness appears to be an important control on development of fissures; more fissures develop at locations having a thin topstratum.

Long irregular fissures through which sand has vented, such as those shown in figures 20*A* and 20*B*, have not been observed to have originated by springs in the alluvial lowlands of the St. Francis Basin. Slumps and lateral spreads along rivers (generally caused by rapid hydraulic drawdown after floods) can generate fissures, but we have not observed sand flowing to the surface through the fissures except for small amounts associated with water oozing through the slide mass. In addition, aseismic lateral spreads and slumps that we have observed do not extend back from banks along large rivers more than 50 m. Thus we believe that fissures associated with vented sand that is located more than 50 m from present or former banks of rivers can generally be attributed to seismically induced slope instability. Confidence in this interpretation increases with increasing width of the sand-filled fissure and with increasing distance from the banks. In order to confidently attribute an earthquake mechanism to a specific vented-sand fissure, however, the location of the river at the time of fissure formation must be established, an undertaking that may be quite difficult.

Fissures formed on terraces far from streams can be attributed to earthquakes more confidently. In interpreting an origin for any feature, it is still necessary to assess ground-water conditions at the site and ascertain that the fissures could not have originated by some process related to syndepositional deformations.



◀ FIGURE 19.—"Eruptive vent" that cut stratified vented deposits of a vented-sand volcano (from Haller and Crone, 1986). *A*, Number 102 is above a severely disrupted zone. Distances between vertical edges of photograph is approximately 1 m. *B*, Number 61 is at contact between vented sediments and the original ground surface. Distance between vertical edges of photograph is approximately 1 m. *C*, Line drawing of "eruptive vent" in *B*. Heavy lines are stratigraphic contacts; fine lines show laminations within the "eruptive vent," dashed where discontinuous. Open circles are aligned clay fragments.

INTRUDED FEATURES

Intruded features are sand-filled cracks that do not reach the surface, and therefore the features are visible

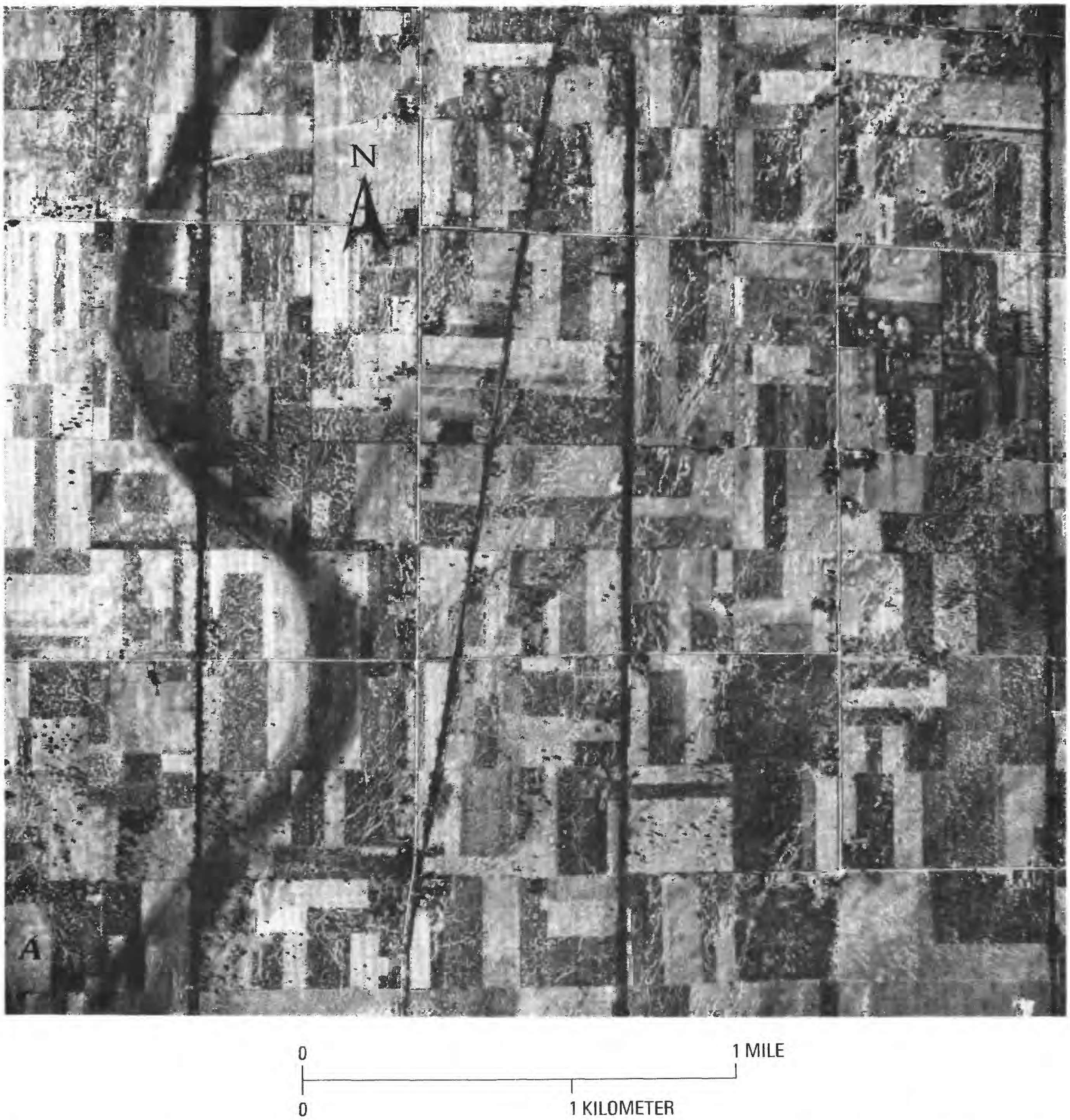


FIGURE 20.—Aerial photographs showing vented sand, interpreted as the product of liquefaction and flowage during the 1811–12 earthquakes. Note pattern of crazing in figures 20*A* and *B*, which is especially indicative of ground breakage caused by earthquake-induced liquefaction. *A*, Taken in 1941, showing more than 25 percent of ground surface covered by vented sand (small white spots and thin linear zones) near Marked Tree, Ark. *B*, Taken in 1959, showing more than 25 percent of ground surface covered by vented sand (small white spots and thin linear zones) near Portageville, Mo. *C*, Taken in 1940, showing localized venting of sand near Reelfoot Lake, Tenn. Regions with vented sand are outlined.

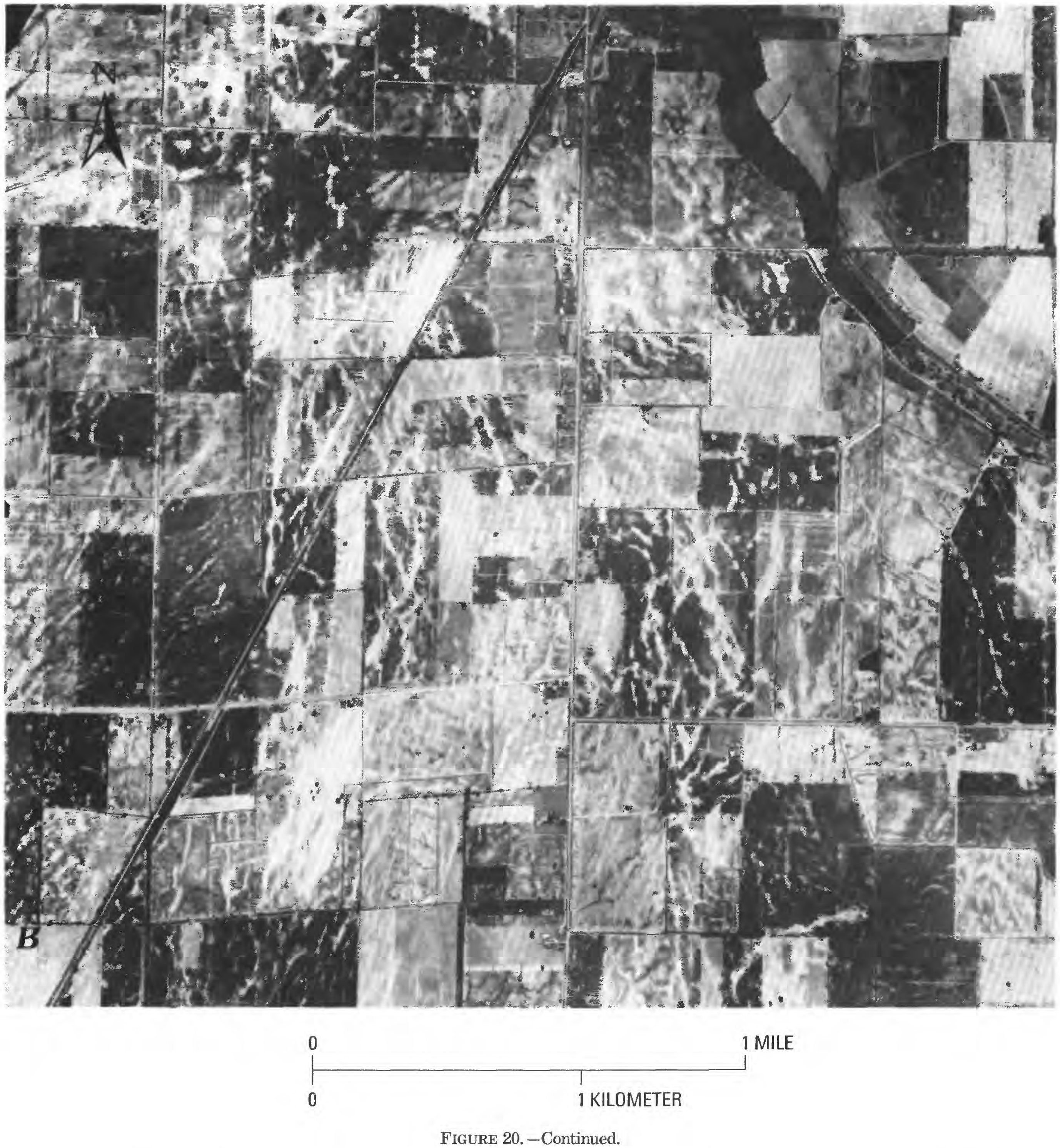


FIGURE 20.—Continued.

only in exposures cut into the subsurface. A simple, very common type of intrusion having an earthquake origin is a near-vertical, wedge-shaped, sand-filled dike. At many places in the region of vented sand, the clay-rich topstratum is cut by vertical dikes that are spaced tens to hundreds of meters apart. The dikes extend upward from the substratum. Dike widths are several centimeters near the substratum; widths commonly narrow upward to an apex near the ground surface. The sand in the dike is generally structureless and can contain clasts of topstratum carried upward; the larger clasts are often oriented vertically.

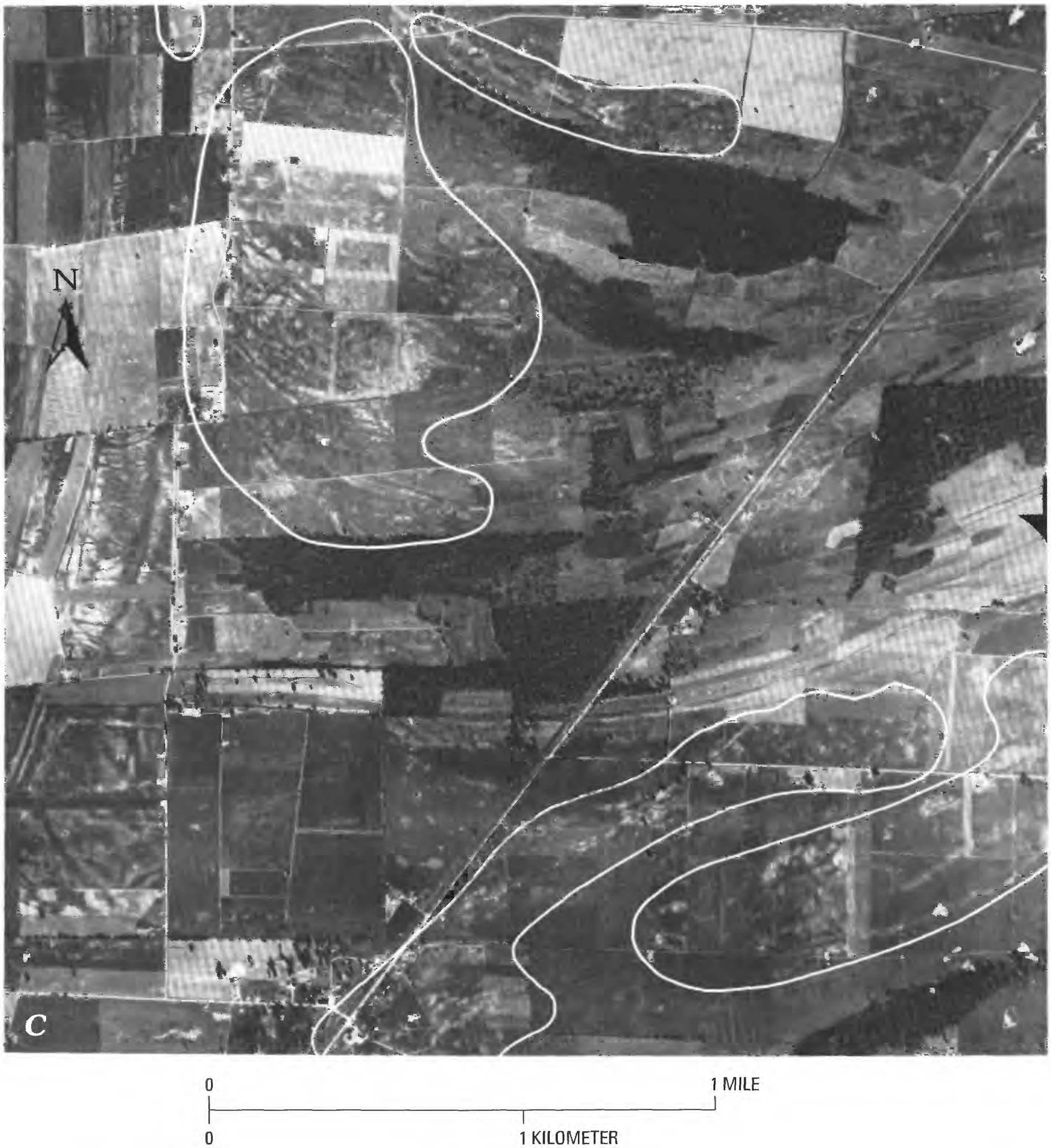


FIGURE 20.—Continued.

Combinations of dikes and sills are also common in the topstratum. The sills generally follow bedding planes or other horizontal planes of weakness in the topstratum. Sills tended to form in thin beds of silt or sand beneath only slightly deformed clay beds, but locally thin clay beds can be strongly warped. Figure 21*A* shows a typical relationship of near-vertical dikes connected to a laterally extensive sill. (The sill extends beyond the photograph and is at least 25 m long.) The internal layering of this sill is also typical of many of those we have studied. Individual laminae are composed of small pieces of charcoal or of much finer grained sand and silt. (See figs. 21*B* and 21*C*.) A single laminae of silt or very fine sand can have a length as long as 10 cm. Lamina made up of pieces of charcoal can be much longer, and lengths of 300 cm are not unusual. In some places, the sills are structureless, having no traces of bedding, and contain many clay clasts in a sand matrix; in other places, sills have graded bedding with clay clasts concentrated along the base. The absence of bedding or the presence of graded bedding seems quite understandable; however, the presence of planar strata of fine sand and silt within a much coarser sand cannot be explained by us with any degree of confidence. (Possibly the planar stratification develops by repeated intrusions of fluidized sand into an open crack.) Most certainly, however, sills having stratified bedding are common earthquake-related features that resulted from intrusion of liquefied substratum sands. (Sills that have stratified and graded bedding originating from earthquake-induced liquefaction of much deeper source beds have also been observed in the South Carolina study at sites HW and BLUF on fig. 1.)

The clay clasts in the sills are generally angular, suggesting a brittle or shattering mode of breakage of the clay stratum from which the clasts were derived. This brittle mode of breakage also suggests that when the stratum was "shattered" the clay stratum was dewatered significantly from the initial depositional condition. Such shattering indicates forceful intrusion and thus is not equivalent to the process causing formation of syndepositional soft-sediment deformation features (discussed later). We have observed that the shattered clay strata can have a consistency that is so soft that a person can force his thumb a few inches into a stratum.

Sills in the topstratum at depths of about a meter or less below the ground surface may be very wavy in vertical section and can thin and thicken dramatically within a meter, producing bulges at the surface. Sills can be as much as 0.5 m thick near the surface. Such large thicknesses seem less common at greater depths.

Earthquake-induced sand dikes that cut the upper substratum sands and the lower topstratum are shown in figure 22*A*, which is a schematic depiction of a vertical ditch bank exposed about 15 km northwest of Marked Tree. The exposure also illustrates many of the other common forms of intrusions in the topstratum. In the

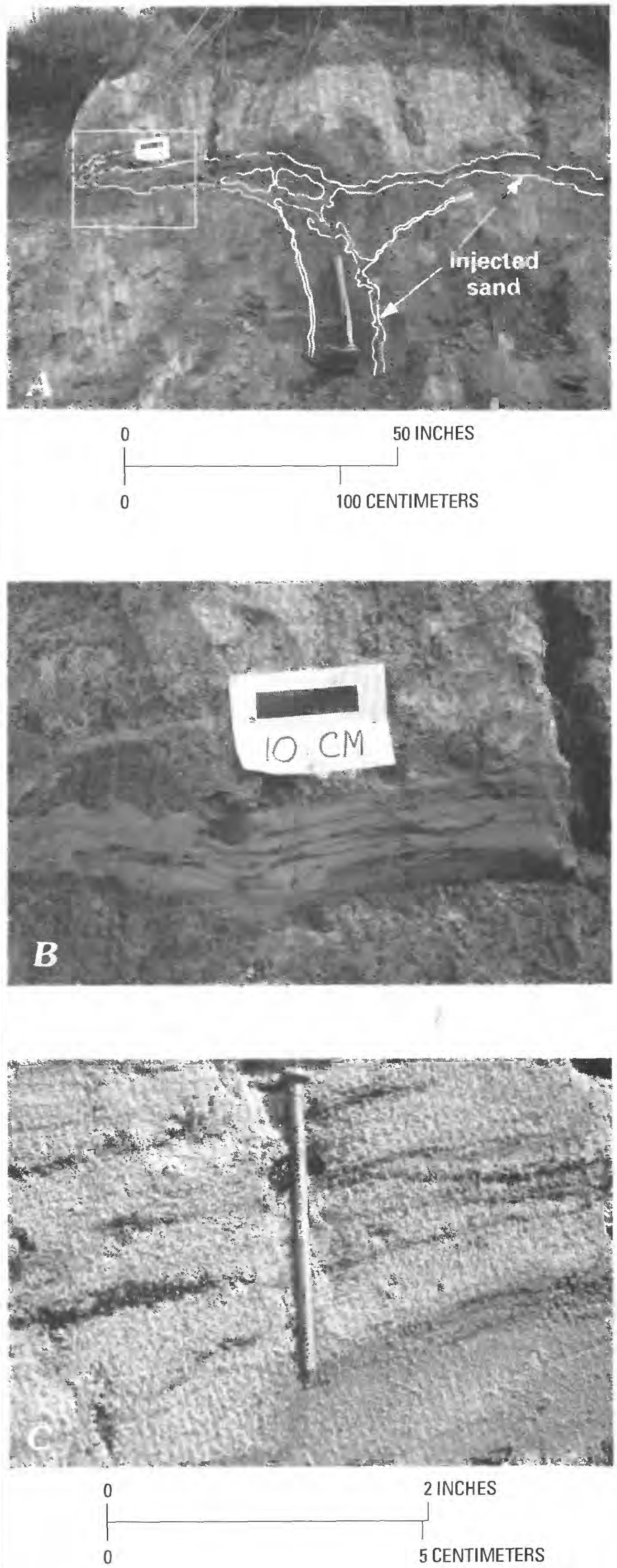


FIGURE 21.— Sand dikes and sills, interpreted as having originated ▶ by liquefaction and flowage during the 1811–12 earthquakes. *A*, Overview and line drawing of typical dikes and sills. Rectangle shows area of figure 21*B*. *B*, View showing detailed layering in sill. *C*, Very close view showing detailed layering in sill.

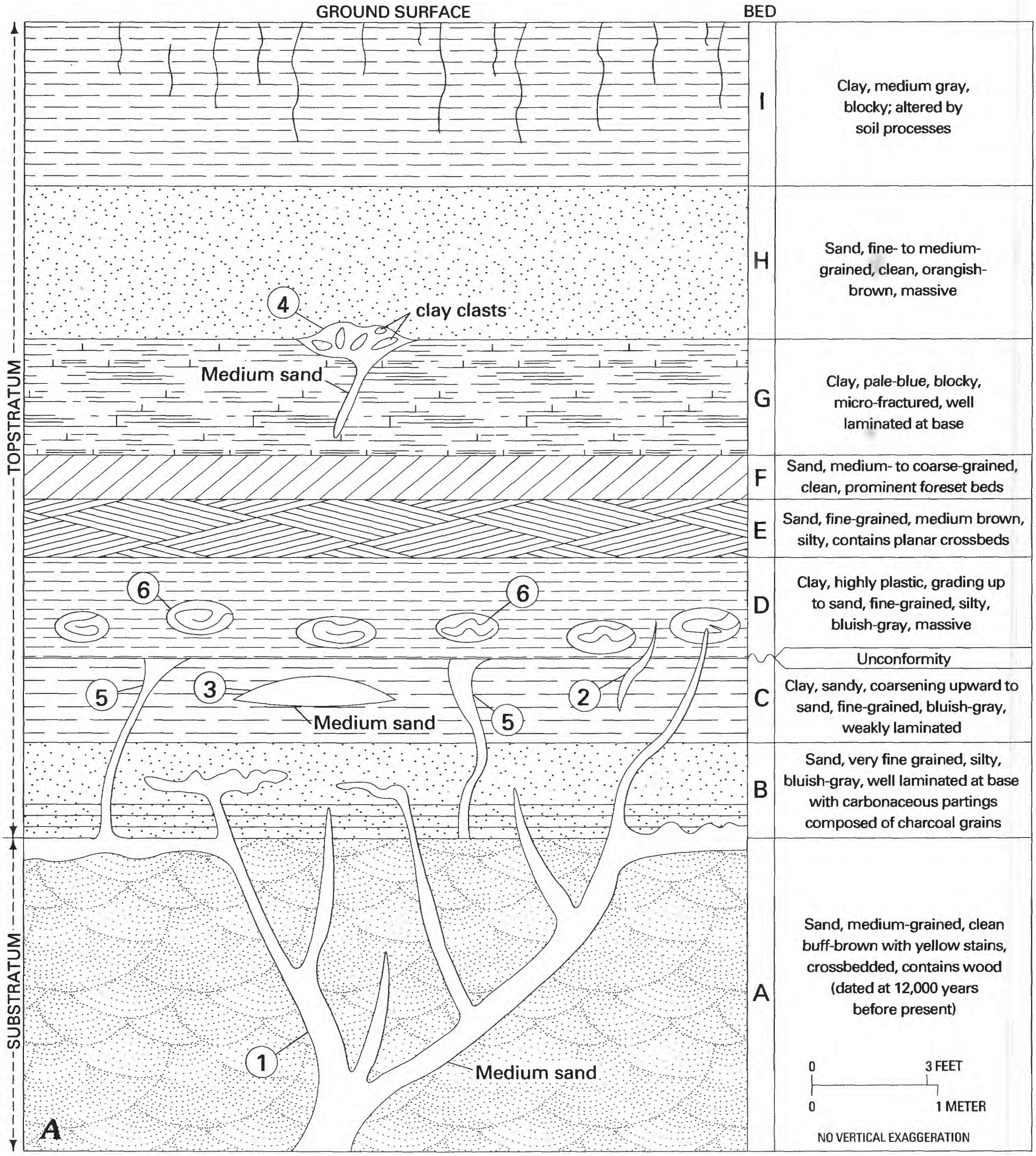


FIGURE 22.—Section showing Holocene sediments (topstratum) and underlying Wisconsinan braided-stream sands (substratum) in ditch about 15 km northwest of Marked Tree, Ark. Earthquake-induced intrusions cut section at many places. *A*, Schematic diagram of stratigraphic relationships and earthquake-induced liquefaction features (numbered 1 through 5). Feature 1—Dike of medium sand that cuts substratum and topstratum. Features 2 and 3—Intruded dikes and sills of massive, clean, medium sand. Feature 4 —Dike of medium sand and large clasts from bed G. Feature 5—Dikes of medium sand, truncated by unconformity. Feature 6—Pseudonodules collapsed into bed D.

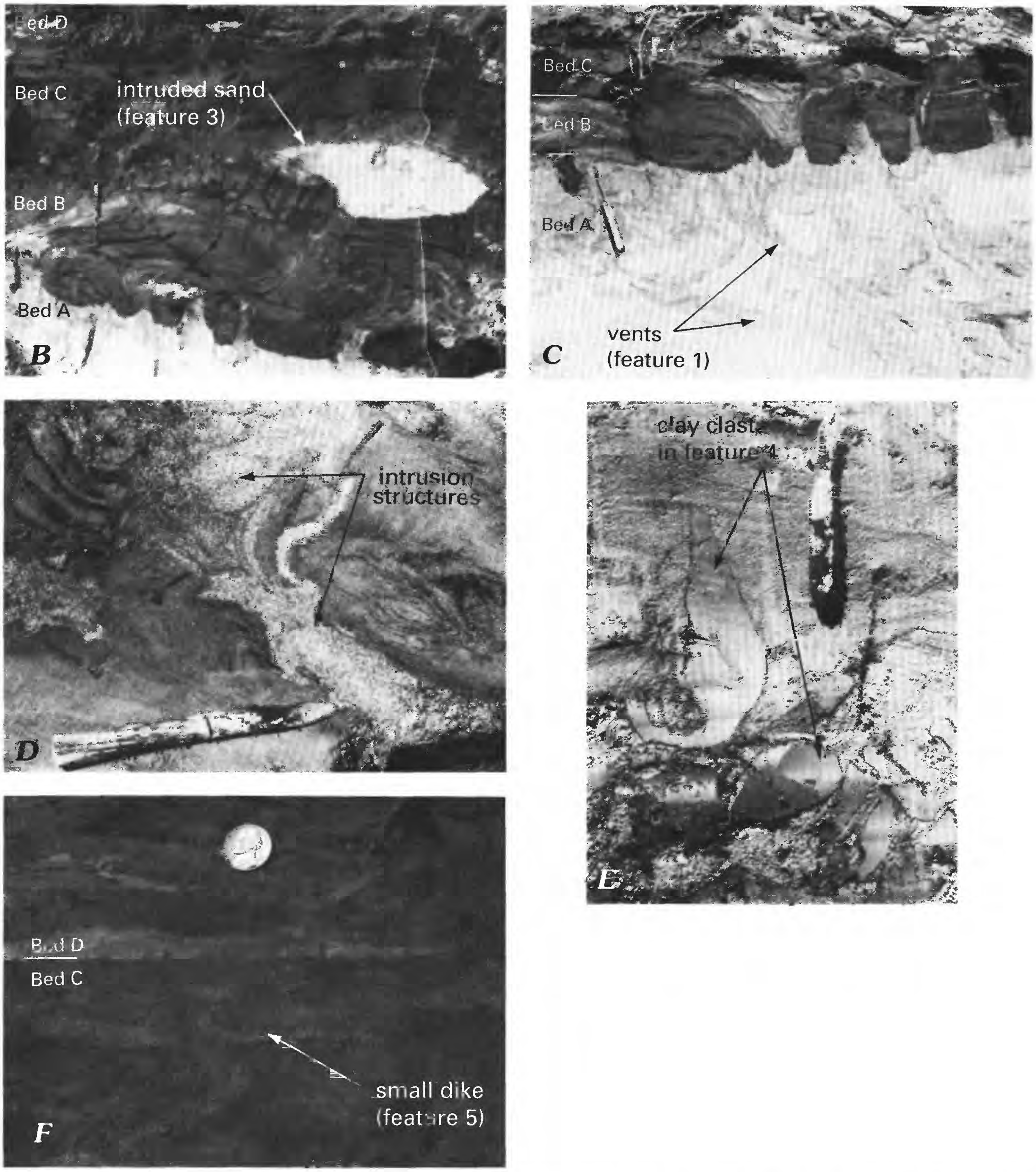


FIGURE 22.—Continued. *B*, Photo graph of beds A–D, showing sand intrusion (feature 3). Knife is 12 cm long. *C*, Photograph of vents (feature 1) cutting substratum sand (bed A) and bed B. Knife is 12 cm long. *D*, Photograph of plan view of bed B, showing intrusion structures caused by feature 1; the plan view is in the area of the knife oriented vertically in figure 22*B*. Knife is 12 cm long. *E*, Photograph of part of feature 4, showing clay clasts in sand matrix intruded into bed H. Knife is 12 cm long. *F*, Photograph of dike of unknown origin (feature 5), cutting through bed C and truncated by bed D. Coin diameter is about 1 cm.

upper part of the substratum (bed A, fig. 22*A*), small dikes branch out from a large dike (feature 1), cut through the basal topstratum bed (bed B) at horizontal intervals of 0.5 to 1 m, and extend upward about 0.5 to 1 m. However, a few dikes and sills have intruded at much higher levels (features 2, 3, and 4). At places nearby (not shown) dikes extend to the surface, and sand has been vented to produce sand blows.

The dikes and sills shown on figure 22*A* as features 1 through 4 contain clean, structureless, medium-grained sand. The edges of the intrusions are very sharp in clay-rich beds (such as beds C, D, and G). Edges are generally less distinct in beds of clean permeable sand. Dikes commonly terminate in a permeable sand bed (bed H); near the terminus there are often numerous clasts of clay-rich material from the subjacent bed.

Features labeled 1 through 4 in figure 22 are probably earthquake-induced because (1) they are widely distributed (common for tens of kilometers), (2) they contain evidence of intrusion by large volumes of water-saturated sediment (dikes and sills are commonly up to 15 cm wide), (3) there is evidence of forceful intrusion (clean, medium-sized sand containing large clay clasts), and (4) they are found in sites where artesian conditions are unlikely. Because of the lack of weathering of sand in the intrusions near the ground surface, we believe that features 1 through 4 probably formed during the 1811–12 earthquake.

The dike shown as feature 5 has an uncertain origin. Three small dikes that are truncated at the contact of beds C and D were exposed in a 25-m section exposed along the ditch but were not found in other nearby exposures of beds C and D nor in the lowermost portion of the topstratum at exposures further away (where beds C and D cannot be traced). The dikes contain a large amount of silty fine sand and cannot be traced far down into substratum sands. From this information, we interpret that the dikes quite possibly originated due to springs formed near the base of a streambank or, less likely, as slump-related features.

Feature 6 (pseudonodules) is discussed in the next section of the paper.

FEATURES OF UNKNOWN OR NONEARTHQUAKE ORIGIN

The principal features in this category include sand boils, mima mounds, load structures, sand dunes, “deer licks,” and sand ridges on point-bar deposits. Some of these features can be distinguished from sand blows on aerial photographs. Others may remain uncertain in their origin even after field studies.

SAND BOILS

During floods that occur every few years, sand boils originate due to artesian flowage beneath levees. Sand boils are common in lowlands near levees along the large rivers, particularly the Mississippi River (Kolb, 1976), and are generally restricted to the areas within 0.5 to 1 km of the levees. (See fig. 15.)

The sedimentary structures and morphologies of sand boils are very similar to and may not be distinguishable from those of (earthquake-induced) sand blows. For example, both processes vent sand to the surface from source beds at depth. The stratigraphic position of the bed that provided the sand can be used to indicate origin at some places. In a sand boil, the sand on the surface commonly can be traced to a sand bed that immediately underlies the cohesive (and thus relatively impermeable) topstratum. In contrast, the source bed for a sand blow may be much deeper. By itself, this criterion is of limited value and generally is best used in a context that includes tests to determine the earthquake-induced liquefaction susceptibility of the various beds (discussed later).

The shape of the vent in plan view may help distinguish between sand boils and sand blows. Both sand boils and sand blows can have circular vents; however, many sand blows have linear vents that follow large fissures. Where large sand-filled fissures widen downward and extend for tens to hundreds of meters laterally, they are probably vents associable with earthquake-induced liquefaction (providing the possibility of nonseismic slumping can be eliminated as the source of the fissures).

Sand boils have vents whose diameters range from 3 cm to greater than 1 m. Numerous small sand-filled fissures or tubes can be connected to the main vent of a sand boil; clay clasts up to 7 cm in diameter, derived from sidewalls, have been observed within the vent. The upper parts of the vents (as deep as 1 m or more below the original ground surface) of sand boils may contain stratified deposits of sand layered with other materials that fell into an open hole. A nonearthquake origin can be proven at some places because of the age of the materials in the vents. For example, the presence of agricultural produce such as rice or soybeans far down in the vent virtually eliminates an association with historic earthquake origin in the New Madrid region, especially if the vent is within 1 km of a levee. (Soybeans and rice have been cultivated only within the past 50 years or so, much postdating the last liquefaction-producing earthquake, the 1895 Charleston, Mo., earthquake.)

MIMA MOUNDS

Mima mounds (sometimes called prairie mounds) are domes of sand-rich material, usually less than 30 m in diameter and 1 m high, that were not formed by venting;

in exceptional cases the domes are as much as 1.7 m high but have diameters of only 10 to 12 m. Mima mounds are present in alluvial lowlands (especially in the northern parts of the St. Francis and Western Lowlands Basins, north of the town of New Madrid) and are also in upland areas west of the Western Lowlands (fig. 13). The origins of mima mounds are not understood. In many upland areas, the mima mounds are formed on nonliquefiable deposits and therefore are of nonearthquake origin. Mima mounds on alluvial lowlands can be identified as not resulting from earthquakes if excavation shows an absence of vents to connect the mounds to source beds below (Fuller, 1912, p. 80).

On aerial photographs, mima mounds can often be distinguished from sand blows by criteria illustrated in figure 23, which is an aerial photograph of mima mounds taken in the northern part of the St. Francis Basin. These criteria are regularity of spacing, alignment, and size. In contrast, sand blows are typically irregularly spaced, have widely differing sizes (heights and diameters) at locations nearby, and are not aligned along such precisely defined curves. Comparison of sand blows formed on point-bar deposits (fig. 20*C*) with figure 23 illustrates the contrast.

Locally it can be difficult to distinguish between sand blows and mima mounds on aerial photographs, and excavation is required to examine for feeder vents or other evidence of origin. Mima mounds in the vicinity of the St. Francis Basin characteristically have well-developed soil horizons that obviously predate the 1811–12 earthquakes and in addition are slightly cemented, making them difficult to auger using hand tools.

LOAD STRUCTURES

Many Holocene fluvial deposits in the Central United States have abundant syndepositional, soft-sediment deformation features known as "load-flow" or simply "load" structures. Load structures are bulbous downward intrusions of sandy or silty material into underlying weaker, finer grained muddy sediment. Two common types of load structures occur in the topstratum of the St. Francis Basin—pseudonodules and load-casted ripples. Pseudonodules occur when overlying sandy or silty sediments become detached and sink to become isolated subspheroidal bodies in the underlying clayey bed. Pseudonodules are usually found laterally adjacent to other undetached load structures (Allen, 1984, p. 359–360). Pseudonodules were experimentally produced in the laboratory by Kuenen (1958) by hammering a container in which sand overlies water-saturated, very soft clay. (See fig. 24.)

Sims (1975) correlated subspheroidal pseudonodules found in modern lake sediments, similar to those produced by Kuenen (and some possibly morphologically the same as those produced by Kuenen), with known earthquake events. Sims argued that structures such as those produced by Kuenen could be interpreted as earthquake-induced if (1) they occur in a seismically active region, (2) they are restricted to specific stratigraphic horizons, (3) they are correlative over large areas within a sedimentary basin, and (4) there is no detectable influence of slope movement or failure. However, pseudonodules similar to Kuenen's can form without shaking by the gravity-induced instability of denser sediment overlying less dense sediment (Allen, 1984, p. 363). The degree of soft-sediment deformation is controlled by the difference in densities between the two adjacent layers and the strength of the underlying layer. Sand deposited rapidly over water-saturated, muddy or extremely soft clay is ideal for deformation.

In the case of load-casted ripples, sandy or silty intrusions form because of the unequal loading of migrating ripples of sand on a clayey mud substratum. Load-casted ripples show progressively deformed radial internal lamination caused by the rotation of the ripple cross-laminations as the ripples sink (Dzulynski and Walton, 1965, p. 146–149). The asymmetry reflects current direction, as illustrated in figure 25. The development of load-casted ripples requires local deformation synchronous with deposition of overlying sand, and therefore load-casted ripples cannot likely be related to earthquake events.

The distinction between possible earthquake-induced load structures and those produced by rapid sedimentation is illustrated in a ditch exposure near Marked Tree, Ark. The section in plate 1*A* is composed of bluish clays interbedded with layers of sand, apparently related to intermittent crevasse delta sand deposition in a water-filled swale as illustrated in plate 1*B*. Small, deltalike sand bodies formed convex-upward sand lenses containing climbing ripple cross-laminations. The lenses are coarser upward, fine toward the edges, and are laterally adjacent to layers of pseudonodules (25–50 cm long and 5–25 cm wide) made up of similar grain sizes. The base of each sand lens has load structures, including load-casted ripples (pl. 1*B*). The load structures make up a greater percentage of each sand lens as it thins toward the edge (see north-south view of pl. 1*B*); downstream, each sand lens apparently grades into the pseudonodules. (See east-west view of pl. 1*A*.) The cross-lamination in the sand lenses has a gradational contact to the deformed ripple cross-laminae in the underlying load structures, whereas ripple cross-laminae above and laterally adjacent to the load structures are undeformed. (See north-south view of pl. 1*B*.) The pseudonodules are internally ripple cross-laminated, and some have load-casted ripples at the base. Isolated pseudonodules with the

FIGURE 23.—Aerial photograph of mima mounds in northern part of St. Francis Basin. Mima mounds often appear as trails of light dots and are commonly formed on slightly elevated ridges of point-bar deposits.

internal fabric of load-casted ripples have an asymmetry reflecting a current direction about the same as that for the sand lenses. The clay surrounding the pseudonodules has streaks of highly contorted silt, indicating flowage

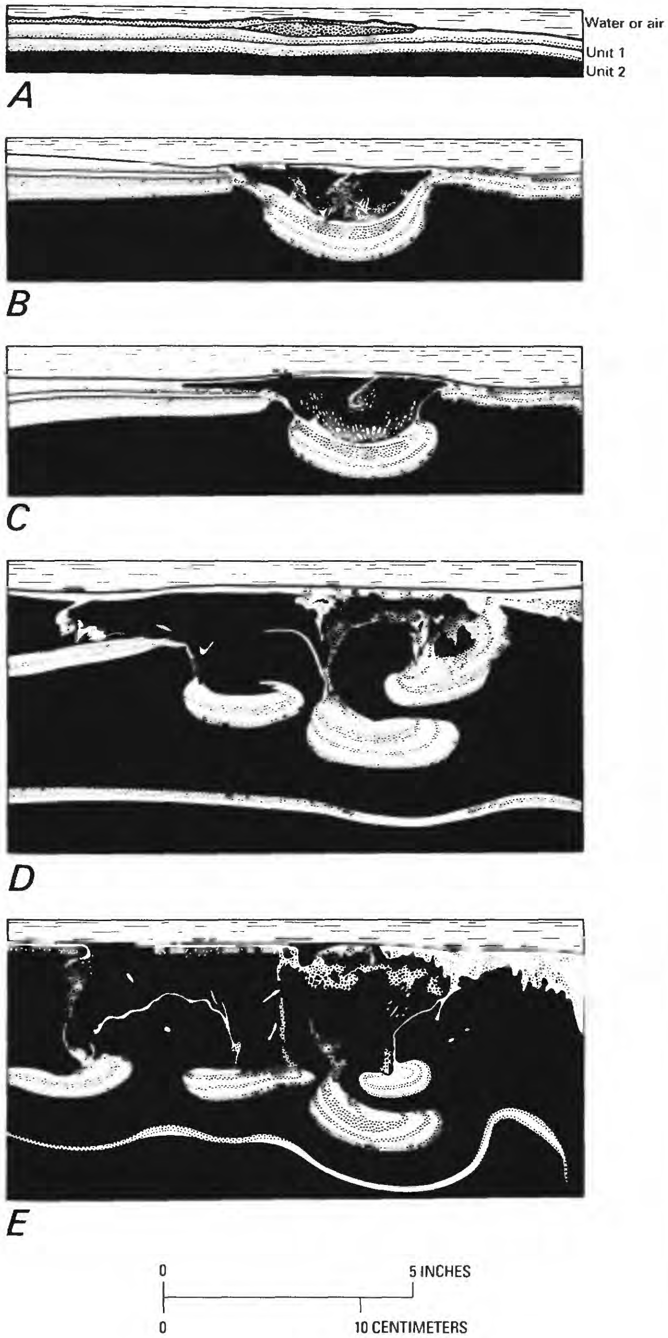


FIGURE 24.—Pseudonodules formed by shaking, from Keunen (1958). A layer of sand (unit 1) overlies very soft clay (unit 2) in *A*. With shaking, pseudonodules developed (*B* and *C*) and became completely enclosed in clay (*D* and *E*). Note the destruction of layering from *C* to *D*. This deformation is due to sinking of the entire sand layer into the very soft clay and also to localized sinking caused by uneven loading.

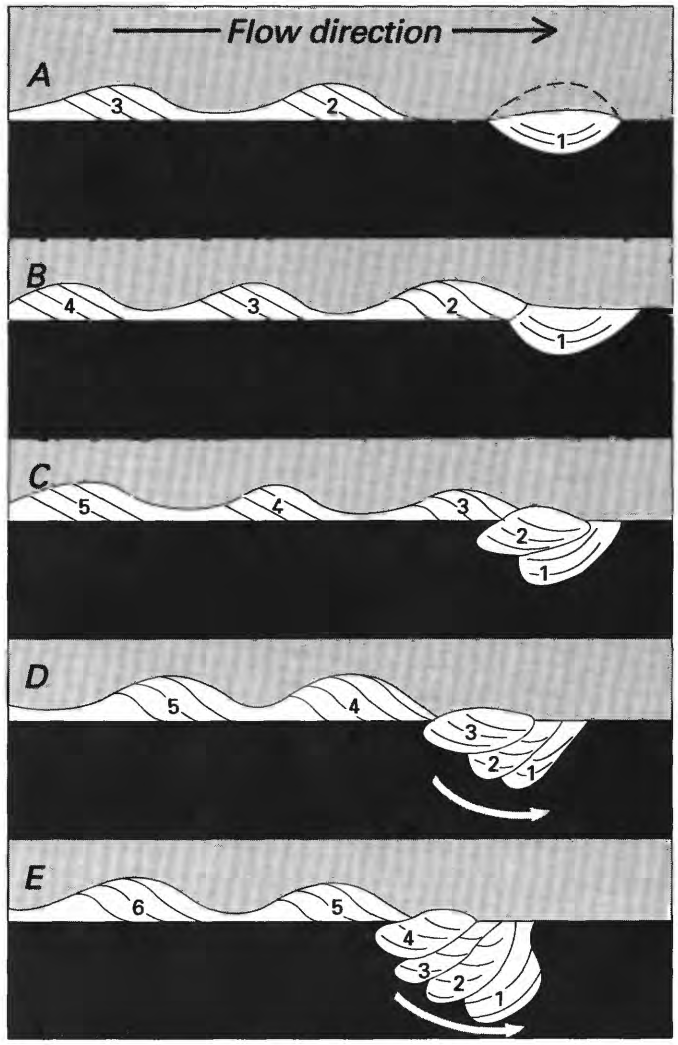


FIGURE 25.—Development of load-casted ripples, caused by ripple crests sinking into soft mud. Note the progressive tilt of the internal cross-lamination, which causes the downflow portion to be more steeply inclined than the upflow portion (modified from Dzulynski and Kotlarczyk, 1962). Load-casted ripples can also have an opposite sense of rotation (upflow more steeply inclined than downflow), as evidenced in field observations in the St. Francis Basin.

around the pseudonodules. The lower portion of the sequence of sediments in the trench is brecciated and faulted; the breccia, including pseudonodules, is surrounded by a coarse-grained sand that has been injected from beneath. These larger scale deformation features are truncated by the uppermost sand lens (sand unit X), which is also laterally equivalent to pseudonodules.

The load structures in plate 1*A* are interpreted as synsedimentary in origin and *not* as earthquake-induced on the basis of the following criteria: (1) The pseudonodules are gradational into the sand lenses, including load-casted ripples, with the same sense of flow direction, (2) there are several layers of pseudonodules laterally equivalent to largely undeformed sand lenses, and (3) the recurring conditions of sedimentation (rapid progradational deposition of sand over soft clay) were conducive to this type of soft-sediment deformation. Although load structures such as these may be formed by earthquakes, their repeated occurrence in this sequence (as also shown by Allen, 1984, figs. 9, 10; or in Coleman and Prior, 1980, figs. 15, 21, 22, and 32) and their apparent localized occurrence argue for a depositional origin. (The injection features, brecciation of clay, and faulting in this exposure that crosscut the syndepositional features are probably of earthquake origin.)

Pseudonodules formed by earthquake shaking would be similar in shape to those formed by depositional processes, because the mechanism of sinking is similar. The environmental conditions conducive to pseudonodule development are also ideal for the formation of convoluted bedding, fluidization dikes, and slumping. Any depositional environment in which coarse-grained sediment is introduced rapidly in otherwise quiet, standing-water conditions, such as fluvial overbanks, lacustrine or marine deltas, and submarine fans, may include these features as part of its regular sedimentary record. We suggest that the difference between syndepositional deformation and earthquake-induced deformation would be in the relationship of the pseudonodules to the surrounding sediments. Intense earthquake shaking might cause collapse of an entire sand layer into clay, rather than only the thin edge, and a small portion of the base of a sand lens; in addition, the structures should be widespread in deposits having the same age. A widespread distribution of load structures formed at the same time is difficult to prove in most fluvial depositional environments.

OTHER FEATURES

Field examination is often required to distinguish between sand blows and small aeolian sand dunes, particularly on high terraces having very sandy surface soils. Another feature requiring field examination is a "deer lick," which is a very light-colored area, up to 15 m in diameter, caused by a high sodium concentration in poorly drained silty soil on level ground (Brown and others, 1973). They are relatively common in western Tennessee.

Aerial photographs often show arcuate ridges of sand on modern and Holocene point-bar deposits, as illustrated in figure 20*C*. These sand ridges form where the fine-grained topstratum is especially thin, and thus more quickly eroded, or was never present over the point-bar deposits. Field excavations are often required to determine if the sands have been vented versus exposed by erosion.

LOCAL GEOLOGIC CONTROLS ON PRODUCTION OF VENTED SAND

Some of the geologic controls on production of vented sand during the 1811–12 earthquakes were first noted by Saucier (1977) and were later described in much more detail by Obermeier (1988, 1989). Principal controls are topstratum thickness and lithology, substratum sand grain size (and, indirectly, permeability), susceptibility to liquefaction of source sand (as related to age of the sediment and measured by the Standard Penetration Test (SPT)), and earthquake accelerations. The influence of these parameters was determined by hundreds of test borings, which were used to formulate a general notion of the geologic setting where sand was vented and to back-calculate 1811–12 earthquake accelerations. Along the border where sand blows developed (fig. 14), the Simplified Procedure of Seed and Idriss (1982) was used to estimate peak earthquake accelerations. Accelerations along the border probably did not exceed 0.25 g (Obermeier, 1988, 1989).

Substratum sands in the St. Francis Basin are typically at least moderately dense (engineering sense). N_1 values (SPT blow counts adjusted to an overburden stress of 1 ton/ft^2) in the depth range most likely to liquefy (5–15 m) are rarely less than 10, and median values are generally about 25.

Other basic data that should be kept in mind during this discussion of geologic controls for liquefaction features of the 1811–12 earthquakes are that the substratum sands generally have a narrow range of thickness (between 25 to 40 m); the epicenters of the three major shocks in 1811–12 were almost certainly somewhere within the areas where large volumes of sand were vented; and the M_s values of the three major shocks were probably about 8.5 to 8.7.

TOPSTRATUM THICKNESS

In general, where nonliquefiable topstratum thickness exceeded a critical threshold, there was increasing resistance to development of sand blows and venting of large volumes of sand (Obermeier, 1989). A topstratum thicker than 10 m presented a major barrier to more than scattered venting, even in close proximity to the epicenter. No sand blows formed where the topstratum was thicker than about 15 m. Within 20 to 25 km of the epicentral regions, a 5- to 6-m-thick topstratum was insufficient to prevent severe, extensive venting. At more distant localities, near the border of sand blows, a 3- to 4-m thickness of topstratum noticeably reduced sand-blow development. Where a thick topstratum restricted venting to the surface, intrusions within the lower part of the topstratum were still plentiful.

TOPSTRATUM LITHOLOGY

Even though the great majority of our field observations were at sites where the topstratum was dominantly clay, we observed enough areas where the surficial deposits were sand rich to make some general observations about the formation of craters and vented-sand volcanoes. The largest craters occurred where the topstratum was most sand rich at the ground surface. The stratigraphy in the craters is generally not well defined in comparison to craters in South Carolina. The largest clay clasts are at the base, and clast size decreases upward in a structureless sand matrix. The transition from graded zone to the overlying laminated bedding is difficult to find in some craters; the laminated bedding is generally defined only by thin, wispy layers of slightly different sizes of sand and silt. There is no evidence of organic material accumulating in an open hole, as in South Carolina.

We did not observe V- or U-shaped fractures in regions of craters, as in South Carolina. Partly for that reason, and partly because of the difficulty of making moderately dense, well-graded sand flow extensively after being liquefied, we suspect that the mechanism for formation of craters in the New Madrid region differs from the mechanism in South Carolina, where the eruptive craterlets were abundant. In the New Madrid region, the craters may have formed primarily because of abrasion and widening of the sidewalls as fluidized sand and water vented for a prolonged time. Similarly, large craters did not tend to form extensively in the clay-rich topstratum in the New Madrid region because of the difficulty of abrading the sidewalls.

SUBSTRATUM GRAIN SIZE

Medium-grained and coarse-grained, clean, well-graded sands greatly predominate in the substratum, in the depth ranges that were liquefied in 1811–12. Other factors being equal, the most extensive development of sand blows corresponded to areas underlain by medium-grained sand. Much smaller and more scattered sand blows developed over coarse-grained substratum sands than over medium-grained sands. Coarse-grained substratum sands have much higher permeabilities than the medium-grained sands; possibly the higher permeability permitted pore pressures of the liquefied sands to dissipate rapidly enough to curtail sand-blow development.

RESISTANCE OF SOURCE SAND TO LIQUEFACTION

Only small regional variations in the relative density of the late Quaternary substratum sands are found in the St. Francis Basin; these sands are typically at least moderately dense and thus moderately difficult to liquefy. However, historical accounts (Street and Nuttli, 1984) and data by Fuller (1912) show that 1811–12 sand blows were common on modern flood plains far from the St. Francis Basin. The modern flood plains are underlain by younger and generally looser and more liquefiable sand beds than those in the St. Francis Basin (Obermeier and Wingard, 1985), thus explaining the development of sand blows much further from the epicenters.

Liquefaction features in the New Madrid seismic zone have not been observed to be induced by numerous historic earthquakes having m_b of 5.3 to 5.5, or by a single earthquake having m_b of 6.0; in 1895, many sand blows (generally small) were produced by an m_b 6.2 earthquake whose epicenter was near Charleston, Mo. (Obermeier, 1988). These sand blows developed in alluvium that is slightly less susceptible to liquefaction than that at most other places in the New Madrid seismic zone, excluding the modern (<500 yr) point-bar and sand-bar deposits of the rivers in the region. Thus a reasonable earthquake threshold is approximately m_b 6.0 for small liquefaction features in braided stream and meander belt deposits, excluding very young alluvial deposits.

OVERVIEW OF STUDIES

Geologic criteria have been developed and geologic controls have been evaluated for earthquake-induced, level-ground liquefaction features in alluvium of coastal South Carolina and the New Madrid seismic zone. Many different types of earthquake features occur in these areas, in addition to features of unknown or nonseismic origin that might be interpreted as having an earthquake-induced origin.

GEOLOGIC CRITERIA

Assigning an earthquake origin to possible sand blows generally requires that four criteria be satisfied:

1. The features must have sedimentary characteristics that are consistent with earthquake-induced liquefaction origin; that is, there is evidence of an upward-directed, strong hydraulic force that was suddenly applied and was of short duration.
2. The features must have sedimentary characteristics that are consistent with historically documented observations of earthquake-induced liquefaction processes.
3. The features must occur in ground-water settings where suddenly applied, strong hydraulic forces of short duration could not be reasonably expected except from earthquake-induced liquefaction. In particular, such settings are extremely unlikely sites for artesian springs.
4. Similar features must occur at multiple locations, preferably at least within a few kilometers of one another, having similar geologic and ground-water settings. Where evidence of age is present, it should support the interpretation that the features formed in one or more discrete, short episodes that individually affected a

large area and that the episodes were separated by long time periods during which no such features formed.

As fewer of these elements are satisfied, confidence in interpretation of an earthquake origin generally diminishes. Considerable reliance has been placed on the second element for interpreting an earthquake origin in coastal South Carolina and the New Madrid seismic zone. If, however, all the other criteria are unequivocally satisfied outside these geographic areas, an earthquake origin can still be ascertained in areas that have not had historic earthquakes.

Criteria based on engineering analysis may be useful for diagnosis of the origin of features. To illustrate, if the source sand bed can be identified at a site of possible liquefaction, and the sand bed proves to be nonliquefiable on the basis of an engineering analysis, then an earthquake-induced origin can be eliminated. As another example, if mineralogical analysis associates vented sediment with a deep stratum of loose sand rather than with a shallower sand bed of much higher compactness, and an engineering analysis of nonearthquake-related groundwater flow forces shows that the shallower, more dense sand should have been transported up in preference to the deeper, loose sand, then an earthquake origin may be ascertained.

TYPES OF EARTHQUAKE FEATURES

In both coastal South Carolina and the New Madrid seismic zone, earthquake-induced liquefaction features originated in sand deposits that are relatively thick (3–50 m) and contain no or few intercalated silt- or clay-rich strata. At the ground surface there is a cap that is much less permeable than either the subjacent or deeper source sand beds. Properties of the cap have a major effect on the surface expression of the sand blows. In coastal South Carolina, where the cap is generally a 1-m-thick soil that is weakly cemented with humate, the sand blows are expressed as craters surrounded by thin sand sheets; in the New Madrid seismic zone, the cap is generally a clay-rich deposit about 2 to 10 m thick, and sand blows are expressed as sand volcanoes.

Both vented and intruded features can be related to earthquake-induced liquefaction. Features characterized by sand vented to the surface include more or less circular, individual sand blows and also long, irregular sand-filled fissures, hundreds of meters long. Where a clay-rich cap exceeds a critical thickness, sand is not vented to the surface; instead, earthquake-induced liquefaction is represented only by sand intrusions. Intruded dikes usually have a massive internal structure, whereas sills can contain both graded and laminar bedding in coastal South Carolina and the New Madrid seismic zone.

Lateral spreads produced by the 1811–12 earthquakes were common. (See Fuller, 1912, p. 48, for a schematic drawing.) We made no effort to locate lateral spreads in the New Madrid seismic zone, but evidence of lateral spreads is described in coastal South Carolina.

GEOLOGIC CONTROLS

The thickness of clay-bearing or impermeable surficial deposits can be an important control on development of surface venting. In the St. Francis Basin, our data show relations between the distance from the epicenter, thickness of topstratum, and development of surface venting produced by the 1811–12 New Madrid earthquakes; near the epicenters, a topstratum thickness more than about 12 m prevented venting, whereas near the farthest limits of sand-blow development, a topstratum more than 4 to 5 m thick greatly restricted development. In coastal South Carolina, a cap exceeding 3 to 5 m thick generally prevented development of sand blows for both 1886 and pre-1886 earthquakes. These differences in critical thicknesses between coastal South Carolina and the New Madrid seismic zone most likely reflect the influence of factors such as thickness of source sand bed, susceptibility of source sand bed to liquefaction, and earthquake magnitude. Similar relations between critical thickness of surficial cap and development of surface manifestations of liquefaction have been noted by Ishihara (1985).

The rate of expulsion of water to the surface while the sand blow was forming possibly controls whether sand blows formed as craters or as sand mounds, but this speculation has not been tested. Certainly, in coastal South Carolina, the sands are generally much looser than sands in the New Madrid seismic zone, and this looser state may have helped create a very thick, water-rich zone during shaking, which was probably the primary source zone during initial, very rapid explusion. Sands in the New Madrid seismic zone characteristically are much more permeable than the marine sands in coastal South Carolina, yet craters are not common features in the New Madrid seismic zone. Thus, the craters cannot be necessarily associated with higher permeability alone of source sands.

Vertical cracks formed by desiccation or other processes may predispose clay-rich sediments to form vented-sand volcanoes rather than craters. Vertical cracks are common in both coastal South Carolina and the New Madrid seismic zone.

Source beds for sand blows in the New Madrid seismic zone are predominantly medium- and coarse-grained sands; in the 1811–12 earthquakes, the coarse-grained and thus highly permeable source sands produced fewer and smaller sand blows than medium- to fine-grained source sands. Insufficient data are available to form conclusions about influence of grain size on sand-blow size in coastal South Carolina, because almost all sands are fine grained.

Geologic age is an important control on liquefaction susceptibility in both the New Madrid seismic zone and coastal South Carolina. Youngest sediments are generally most susceptible to liquefaction, for a given mode of deposition.

SUGGESTIONS FOR FUTURE RESEARCH

In recent years it has become increasingly clear that detection of liquefaction features is valuable as a tool for understanding earthquake activity. Liquefaction features can be used to identify strong earthquake shaking and to date the strong shaking events. We have described our interpretations of earthquake- and nonearthquake-related features from field studies in the New Madrid seismic zone and in coastal South Carolina. Still, more work is needed to better identify potential earthquake hazards in these geographic areas.

In the study of New Madrid 1811–12 liquefaction features, much was learned that is relevant to any search for paleoseismic liquefaction in moderately thick fluvial sediments of the Central United States. Aerial photographs are useful in a search for features formed in 1811–12, and earthquake deposits predating 1811–12 laid down on terraces having clay-rich surface soils may also be visible on such photographs. However, deposits laid down on terraces having sandy surficial soils have probably been so intensely reworked by wind that the deposits would be difficult, if not impossible, to identify. Aerial photographs should range from the earliest possible date (generally the late 1930's and early 1940's) to about the mid-1950's. The earlier photographs may show many features destroyed by modern farming, whereas more modern photographs generally have better overall clarity.

Field checking is generally required to determine the origin of features that on aerial photographs are suspected to have an earthquake origin. In particular, in point-bar deposits, field checking is often required to verify the origin of sand that appears to have been vented along slightly elevated, arcuate ridges. On terraces having sandy surface soils, sand blows are often indistinguishable from sand dunes, and distinctive signs of previous earthquakes may be restricted to long fissures that formed near and parallel to scarps along terrace sublevels.

Searches in drainage ditches and sand pits are probably the only means for identifying liquefaction features that much predate the 1811–12 earthquakes. The best areas to search can be predicted by evaluating the influence of geologic controls such as sediment age, topstratum thickness, grain size of source beds, and proximity to suspected epicenters.

Any search for paleoseismic features in coastal South Carolina and in the Coastal Plain of the Eastern United States should also make extensive use of exposures in ditches; aerial photographs, for example, are useless for locating 1886 and pre-1886 sand blows in South Carolina. Determination of an earthquake or nonearthquake origin for disrupted ground features in the Coastal Plain generally requires a team of people having many skills. In addition to having an understanding of sedimentation processes, soil science, and engineering mechanics, there may also be a need for geochemistry studies because weathering in these humid environments produces bizarre features not previously described in the literature. We have described some weathered and disrupted ground features that could cause confusion, but our catalog is far from complete.

Research is also needed to provide criteria for distinguishing between features having an earthquake-induced liquefaction origin and features having a short-term spring origin. Locations where springs can be easily eliminated as a cause, such as along the crests of beach ridges of coastal South Carolina, are relatively sparse in the Eastern United States. Fluvial terraces in lowlands contain the only potentially liquefiable deposits in most places of interest to paleoseismicity. Possible indicators of origin include the ground failure mode; the nature of the filling in vents, dikes, and sills; geologic characteristics of the source beds; and engineering characteristics of the source beds. The relevance of each indicator is discussed below.

Ground failure mode.—Ground failure mode includes features such as lateral spreads, single long fractures, and shattered ground at the ground surface. All of these features generally indicate an earthquake origin. Earthquake-induced liquefaction does not always produce these types of features, however, and in many cases they are not easy to recognize in ditch exposures even where they are known to have been relatively common, such as the meizoseismal region of the 1886 Charleston earthquake. Use of techniques that can rapidly develop a three-dimensional view, such as ground-penetrating radar, may prove to be of great value.

Nature of filling.—Vents associated with an earthquake-induced liquefaction origin generally appear to be filled with a structureless mixture of sand and silt grains and clasts derived from sidewalls and beds at depth. This structureless mixture apparently represents transport through the vent as a slurry. In some places, particularly near the top of the vent, the finer grained material has been winnowed out, thus indicating transport on a grain-by-grain basis. For either the slurry or the grain-by-grain mode of transport, it should be possible to place limits on the hydraulic forces that caused the transport and, by means of a flow-potential diagram, calculate whether nonearthquake flow forces could possibly have been the driving mechanism.

Numerous observations in the New Madrid seismic zone show a wide variation in the nature of sill fillings caused by earthquake-induced liquefaction. The fillings range from layered bedding to graded bedding to a structureless mixture of sand, silt, and clay clasts; for paleoseismicity interpretations, we suspect that the structureless mixture is associated only with the very forceful injection caused by earthquake-induced liquefaction, whereas layered and graded bedding may be associated with either a spring or earthquake-induced liquefaction origin.

In the New Madrid seismic zone, some of the larger sills within 2 m of the ground surface have domed the overlying beds as much as 0.3 m vertically over a horizontal distance of 5 m; such large doming may be unique to earthquake-induced liquefaction. The nature of the fillings in dikes may also be related to an earthquake-induced liquefaction or spring origin. Relevance of the fillings in vents, dikes, and sills can be verified only by research in the field.

Geologic characteristics of source beds.—Intuition suggests that earthquake-induced liquefaction should be much more effective than springs as a means to destroy original bedding over a large area. This suggestion has not been examined, however, and may be difficult to evaluate in practice. Earthquake-induced liquefaction would also probably produce many more widespread sills and dikes than springs.

Engineering characteristics of source beds.—The source beds containing the material vented by springs should be located selectively with respect to the ground-water setting and flow forces. For springs, the source beds must be connected to the source of flowage in such a manner that flowage goes toward the vent, and the flowage forces must be large enough to carry material from the source bed. This source bed would be, in general, the uppermost stratum of wide lateral extent with respect to hydraulic connection and would also be the uppermost fine-grained, noncohesive sand stratum. The source bed with regard to earthquake-induced liquefaction commonly lies much below (several meters) the uppermost sands of wide lateral hydraulic connection, according to limited observations in deep ditches in the New Madrid seismic zone. Thus, permeability relations may be quite useful for determining the causative mechanism responsible for venting materials.

Standard Penetration Test data that show relative ability to liquefy during earthquake shaking should also prove useful for determining origin at some places. Geologic investigation may also be required to determine origin in areas where Standard Penetration Test data indicate possible source beds that are thin (<1 m) and lie immediately beneath an impermeable cap, because (injected) sills beneath impermeable caps have been observed by us to approach a meter in thickness.

RELEVANCE OF LIQUEFACTION FEATURES

Recognition of the various features described in this paper, and identification of the most probable origin for each, provides a set of important tools for understanding the paleoseismicity in areas where faults are not obvious at the surface and where historic seismic activity is infrequent. Even where faults are available for study and offsets can be documented, there may be doubt that the offsets were associated with earthquakes; the presence of liquefaction-induced features is a means of verifying strong shaking.

The criteria we describe in this paper can be applied to many worldwide geologic settings. We caution, however, that verification of paleoseismic events becomes generally more difficult with increasing age of the event. The distinction between earthquake and nonearthquake liquefaction features is often impossible without knowledge of ground-water conditions. Application of the criteria to pre-Quaternary deposits may be extremely difficult, if not impossible, at many places.

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