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Paleoliquefaction studies in continental settings

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ABSTRACT

Research of the past 15 years has reported the manifestations of seismically induced liquefaction that occur in the sedimentary conditions commonly found in continental settings. And, criteria have been developed and published that can demonstrate a seismic origin for features of liquefaction origin. We present guidelines for conducting a paleoliquefaction search by means of geologic and geotechnical parameters. We also address the interpretation of results of a paleoliquefaction study in terms of locating the source region of a paleo-earthquake and back-calculating its strength of shaking and magnitude. Our critique of the geotechnical methods for these back-calculations points out uncertainties in the techniques that are most commonly used.

The guidelines that we present for a paleoliquefaction search are in terms of both geologic and geotechnical parameters because it is the combination that is critical. Neither suffices alone, and the relations between the geologic setting and geotechnical properties must be appreciated in order to understand why seismically induced liquefaction features are to be found in some locales and not in others.

INTRODUCTION

The study of paleoliquefaction features for seismic analysis is a new technique, developed over the past 15 years. The systematic search for paleoliquefaction features throughout large geographic areas is being used to interpret the paleoseismic record through much of the Holocene into latest Pleistocene time. Searches have been conducted chiefly in the southeastern, central, and northwestern United States, through different settings and seismotectonic conditions. Despite extensive studies on paleoliquefaction, their scope is not widely appreciated even by paleoseismic researchers. It is, thus, not well known that geologic field studies can yield clues about the severity of earthquake shaking and, in many settings, the probable location of the tectonic source zone. Realistic estimates can be made in many settings, even though some of the procedures are very recent and not fully developed and applicable in all situations. The techniques used in the continental United States to verify a seismic origin for suspected features are well developed (Obermeier, 1996). In contrast, some uncertainty is usually inherent

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in using a single procedure to back-calculate the strengths of shaking and magnitude (Olson et al., 2001).

This chapter critiques issues concerning field searches and the interpretation and back-analysis of strength of shaking and earthquake magnitude by using geologic and geotechnical-seismological procedures. The chapter is restricted to features developed on ground that was less than a few degrees in slope when an earthquake struck, and excludes slumps and flow failures induced by liquefaction, which may occur on steeper slopes. We focus on the following aspects of level-ground liquefaction: (1) mechanisms that form seismic liquefaction in the field; (2) field settings where liquefaction should be present if strong seismic shaking had occurred; (3) settings where the absence of paleoliquefation features indicates an absence of strong paleoseismic shaking; (4) how liquefaction features should be used to interpret the tectonic source region of a paleo-earthquake; and (5) how effects of liquefaction can be used to interpret the strength of prehistoric paleomagnitude and shaking. We include material relevant to geologists, seismologists, and engineers.

The summary here of the process of liquefaction is restricted to features that characteristically form in a clay- or silt-rich layer (i.e., host) lying above a liquefied sand-rich deposit. The discussion is based on the premise that collecting adequate data for the analyses described herein requires a search over an area at least tens of kilometers in radius. Exposures must be examined in scattered places often by searching banks of ditches or streams.

The paleoliquefaction record extends through much of Holocene time into the latest Pleistocene in many locales. A typical setting for these ages is in a river valley on the modern flood plain or a terrace a few meters higher, where the depth to the water table is less than several meters and where there are thick, sandy deposits. The liquefaction-induced features most commonly found here are steeply dipping, tabular, clastic dikes that cut a fine-grained host. Intrusions more horizontally inclined, such as sills, also abound in places.

Deformation of soft sediments that involves mud and freshly deposited cohesionless sediment (Allen, 1982; Obermeier, 1996) are not included in this discussion. Furthermore, not only is a seismic origin difficult to verify for plastically deformed soft sediments, but they may form without seismic shaking (e.g., Sims, 1975). Conversely, the origin of clastic dikes can usually be determined easily. Seismically induced dikes and sills typically involve a significantly elevated pore-water pressure. Their formation requires significant strength or duration of strong shaking. The minimum earthquake magnitude to form liquefaction features in most settings is moment magnitude M_w ~5.5 (Ambraseys, 1988), which is about the same as the threshold for damage to human-made structures. The minimum value of peak accelerations to form liquefaction features decreases with increasing magnitude; reported values are as low as 0.025 g for M_w 8.25 and 0.12 g for M_w 5.5 (Carter and Seed, 1988). It is commonly accepted that the vibration frequencies of interest are less than ~10 Hz, because higher frequencies do not induce shear strains large enough to break down the grain-to-grain contacts in granular sediments.

Herein we generally cite articles that contain expanded discussions and comprehensive references. The following section concerning the process of liquefaction and its manifestations is largely from Obermeier (1998a, 1998b); see also articles by Seed (1979), Ishihara (1985), Castro (1987, 1995) and Dobry (1989). Some recent critiques of geotechnical and seismological techniques are by Trifunac (1995, 1999), Pond (1996), Obermeier (1998a), and Olson et al. (2001). The Obermeier (1998b) paper includes numerous photographs showing features, with and without a fine-grained cap, in various types of field settings.

The purpose of this chapter is to serve as a state-of-the-art discussion of the seminal concepts being used for paleoliquefaction studies, as well as to discuss the application of those concepts and to properly credit, by references, from whence the concepts came. This chapter is not intended to serve as a listing of the numerous ensuing paleoliquefaction studies that use those concepts.

THE PROCESS OF LIQUEFACTION AND ITS MANIFESTATIONS

Following Seed and Idriss (1971) and Youd (1973), we define liquefaction as the transformation of a saturated granular material from a solid to a liquefied state due to increased pore-water pressure. Liquefaction is caused by the application of shear stresses and accumulation of shear strain, resulting in a breakdown of the soil skeleton and buildup of interstitial pore-water pressure. The process is typical of cohesionless, or nearly so, sediments, and most readily for fine- to coarse-grained sands, especially where uniform in size. The liquefied mixture of sand and water reacts as a viscous liquid with a greatly reduced shear strength. Liquefaction can be induced by seismic shaking, by nonseismic vibration, and by wave-induced shear stresses. In some loose sediments located on slopes, liquefaction can be triggered by static forces; static triggering mechanisms include an increased shear stress caused by toe erosion or an increased seepage force due to a changing water table.

The shear stresses that induce seismic liquefaction are primarily due to cyclic shear waves propagating upward from bedrock and through the soil column, although waves traveling along the ground surface can be important locally. Sediment on level ground undergoes loading, the shear stresses typically being somewhat random but nonetheless cyclic. Loosely packed, cohesionless sediments tend to become more compact when sheared. When subjected to earthquake shaking, the pore water does not have time to escape from the soil voids and allow the sediment to compact as the grain-to-grain skeleton is collapsing. Complete level-ground liquefaction occurs when the pore-water pressure increases to carry the static confining (overburden) pressure—i.e., the grain-to-grain stress equals zero, which permits large strains, flow of water, and suspension of sediment. Partial liquefaction occurs when the increase in pore-water pressure is not enough to fully carry the static overburden pressure. While, by definition, this is not true level-ground liquefaction, ground failure can occur under a condition of partial liquefaction.

A large increase in pore-water pressure commonly occurs during the transition to the liquefied state. The pore-water pressure carries the weight of overlying sediment and water. In many field situations, the pore-water pressure can increase several-fold within a few seconds, thereby hydrofracturing a fine-grained cap lying above the liquefied zone.

Subsequent densification occurs throughout the column of liquefied sediment during dissipation of excess pore-water pressure, and large quantities of water can be expelled. The water flows upward, carrying along sediment. This process is referred to as "fluidization" by some geologists (e.g., Lowe, 1975). The flowing water causes sediment to be carried or dragged along by other grains. The process of fluidization transports the sediment that fills clastic dikes and sills observed in paleoliquefaction studies.

Liquefaction can result from only a few cycles from many cycles of shaking. For a very loose packing of sediment grains, the breakdown of grain structure can be abrupt and liquefaction is virtually simultaneous with the onset of shaking (National Research Council, 1985, Fig. 2.26). Such loose packing is relatively common in delta and eolian dune deposits as well as in very young (less than 500 yr) river channel deposits (Youd and Perkins, 1978). However, some very young fluvial sands have such a dense initial packing that any pore-water pressure increase during seismic shaking is insignificant, and liquefaction does not occur (Seed et al., 1983). For older Holocene-age river deposits, the buildup of pore-water pressure generally tends to be more gradual and requires more cycles of shearing than for younger deposits. Deposits of Pleistocene age are often very resistant to liquefaction owing to effects of aging and weathering (Youd and Perkins, 1978), but deposits hundreds of thousands of years old and still highly susceptible to liquefaction have been encountered (e.g., Obermeier et al., 1993; Martin and Clough, 1994). A broad range of susceptibilities to liquefaction is commonly encountered in a local field setting.

Liquefaction Susceptibility

Liquefaction susceptibility refers to a sediment property and takes into account the depth to the water table and other factors (e.g., static stress conditions, density aging) that affect the ability of a deposit to liquefy. The relative state of packing of sand deposits (called the "relative density" by geotechnical engineers) is a principal determinant of liquefaction susceptibility in most Holocene deposits; the relative density is, in turn, related to standard penetration test (SPT) blow counts measured in situ (Table 1). Relative density is by definition a measure of how densely the sand grains are packed in comparison to the laboratory determined loosest and densest reference states (Terzaghi and Peck, 1967). Correlations of relative density with SPT blow counts are listed in Table 1. For practical purposes, sediments having blow counts in excess of 30 will not liquefy, even if other factors in the field are very favorable. Loose and very loose sands are generally highly susceptible to liquefaction. The moderately compact sands are generally moderately susceptible.

Liquefaction susceptibility is nearly always measured using an in situ test because of the extreme difficulty and expense of collecting samples that are sufficiently undisturbed for laboratory testing. In recent years, there has been a tendency to use the cone penetration test (CPT) to measure liquefaction susceptibility in situ (e.g., Stark and Olson 1995; Robertson and Wride, 1997). The CPT permits more detailed measurements of sediment properties and stratigraphy than does the SPT, and thereby is likely to provide a more accurate evaluation of in-situ liquefaction susceptability. Cone penetration testing is also relatively inexpensive compared to standard penetration testing. Because of the larger database where SPT results are available, however, particularly in the central and eastern United States, the following discussion focuses mainly on the SPT.

SPT data commonly provide a reasonable measure of relative density in Holocene clean sands (sediment that is composed almost entirely of the sand-sized fraction). Exceptions occur where the sediment has been cemented with chemical precipitates, where the fines content (silt, and clay fraction less than 0.075 mm) is greater than about 15%–20%, and where the mean grain size (50% of the material by weight) is greater than ~2 mm (Seed et al., 1983). Sites where stress conditions in the sediment are unusually high in the horizontal plane (as by prior glacial loading) can also cause misleading values of relative density from SPT readings (Terzaghi et al., 1996).

SPT data, in the absence of chemical precipitates, also provide a measure of the effects of aging and weathering on liquefaction susceptability. Aging and weathering effects can originate from both mechanical and minor chemical sources (Schmertmann, 1987, 1991; Mesri et al., 1990). In the short term (hundreds to a few thousand years), mechanical effects caused by adjustment of grains are likely to dominate aging (Olson et al., 2001). Fortunately, the total effect of chemical and mechanical aging is relatively minor from a practical viewpoint in some and perhaps many field settings. For example, in glaciofluvial deposits that abound throughout the central United States, the maximum change in SPT blow count resulting from aging is probably on the order of 3 or 4, on the basis of the difference between the loosest sediments of modern ages (with blow counts near 0) and the loosest deposits of early Holocene ages (Pond, 1996). This change in blow count almost certainly decreases

TABLE 1. RELATIVE DENSITY OF SAND AS RELATED TO STANDARD PENETRATION TEST

No. of blows	Relative density or compactness
0–4	Very loose
4–10	Loose
10–30	Medium or moderate
30–50	Dense
>50	Very dense

Note: Data from Terzaghi and Peck, 1967.

substantially with increasing initial relative density because of the diminished opportunity for mechanical adjustment of grains. The possible influence of aging and weathering should be evaluated on a case-by-case basis, depending on the geologic setting, in view of the uncertainty of factors that determine the effects of aging (Olson et al., 2001).

Field conditions favorable for formation of liquefaction features

Dikes and sills cutting a fine-grained host are generally readily visible in vertical section. Dikes and sills form most readily where a thick, sand-rich deposit is capped by a low-permeability deposit and the water table is very near the ground surface. Grain sizes that are generally the most prone to liquefy and fluidize, and to form dikes and sills, range from silty sand to gravelly sand. A thickness of 1 m of liquefied sediment generally suffices to form recognizable clastic dikes, although a much smaller thickness can be adequate, depending on factors such as the severity of liquefaction, the local field setting, and the mechanism that forms the dikes.

For the normal range of Holocene sediments in a river valley, clastic dikes cut readily through a cap about 1 m thick, and can cut through a much greater thickness where shaking has been severe (Ishihara, 1985; Obermeier, 1989). The strength of the cap generally has a minor influence on the development of dikes, and dikes have been observed to have formed in hard, massive silt and clay-rich glacial tills (Obermeier, 1996). Caps having large tensile strengths, however, such as fibrous mats, can greatly inhibit dike formation.

Liquefaction typically occurs at shallow depth, less than a few tens of meters (e.g., Seed, 1979). Paleoliquefaction searches by Obermeier in diverse field settings throughout the United States, where the water table was probably between 1 and 5 m deep at the time of the earthquake, show that dikes are found most often in fine-grained caps that are 1 to 5 m thick. Caps thicker than 10 m rarely host dikes, including small dikes along the base of the cap. Most likely, dikes in thicker caps are scarce because greater depths require exceptionally severe shaking for liquefaction. Liquefaction beneath thicker caps is unlikely because the increasing overburden pressure can increase the shear resistance of a cohesionless sediment beyond the shear stress induced by seismic shaking.

Liquefaction is most pronounced where the water table lies within a few meters or less of the surface. A change in water table depth of 10 m can change the ability of a deposit to liquefy from high to nil (e.g., National Research Council, 1985, Table 4-1).

Ground-failure mechanisms

Dikes in a fine-grained cap are induced chiefly by three ground-failure mechanisms: hydraulic fracturing, lateral spreading, and surface oscillations. All produce tabular dikes in plan view.

Hydraulic fracturing in response to seismic liquefaction was first deduced by Obermeier (1994), following discussion of the process in the failure of earth dams by Lo and Kaniaru (1990). Hydraulic fracturing begins at the base of a fine-grained cap sitting on liquefied sediment. Fracturing of the cap typically occurs in response to the high pore-water pressure entering naturally occurring flaws along the base of the cap, such as small root holes and other openings. The pressure causes vertical, tensile fractures that are tabular and are filled with a fluidized mixture of sand and water driven by the hydraulic gradient. Similarly, vertical tabular defects in the cap that formed by weathering can be opened by the high pore-water pressure, leading to the formation of tabular dikes either parallel to one another or irregular, having a nearly haphazard pattern in both plan and vertical views. Dikes from hydraulic fracturing are typically quite narrow, ranging from a few millimeters to less than 10 cm wide.

Lateral spreads reflect translational movement downslope or toward a stream bank. The movement occurs where there is only minor resistance to lateral translation of the fine-grained cap sitting on liquefied sediment. Dikes originating from lateral spreading, and especially the wider dikes, are the result of fluidized sand and water flowing into breaks through the cap that have been opened by shaking and/or downslope gravity (Bartlett and Youd, 1992). Dikes can be as much as 0.5 to 0.7 m in width even where shaking has been only moderately strong (about 1/4 g). Widths of as much as a few to several meters are not unusual. Lateral spreads are typically defined as occurring on slopes of 3° or less (Youd and Garris, 1995). On steeper slopes, liquefied deposits can flow tens of meters to 1 km (Tinsley et al., 1985).

Surface oscillations can cause tabular clastic dikes to originate in response to the fine-grained cap being strongly shaken back and forth above liquefied sediment. We use the term "surface oscillation" as a generic description of an end effect rather than a driving mechanism. This definition is in the sense commonly used by geotechnical engineers to describe liquefaction-related ground failure that requires, in plan view, large back-and-forth straining of the cap; high accelerations may or may not be involved. Indeed, during strong bedrock shaking at sites of liquefaction, the accelerations in the cap can be deamplified to a lower level even as straining of the cap is greatly augmented and breaks apart the cap (e.g., see analysis in Pease and O'Rourke, 1995). Surface oscillations can originate from either body (S) waves (Pease and O'Rourke, 1995) or surface (Rayleigh or Love) waves (Youd, 1984). The back-and-forth straining in the cap can be in the form of either axial or shear strains (Pease and O'Rourke, 1995, Fig. 2-1). Rayleigh waves are likely involved at sites of severe axial straining, and either S or Love waves are likely involved at sites of severe shear strain. The effects of Rayleigh waves are probably best manifested by dikes that tend strongly to parallel one another with a spacing that can range from tens to hundreds of meters apart; these effects generally are most severe in the *meizoseismal zone* (see Appendix) but can also extend far beyond (T.L. Youd, 1998, personal commun.). Surface oscillations from what are likely to have been Rayleigh waves are often seen by observers as traveling ground waves. Dike widths from surface oscillations may be as much as 15 cm (T.D. O'Rourke, 1998, personal commun.). Sites of severe shear straining may be indicated by lateral offsets along dikes and along fractures at the ground surface.

Factors controlling the ground failure. Different levels of shaking are required to form dikes visible at the ground surface for each of the mechanisms of lateral spreading, surface oscillation, and hydraulic fracturing. For cohesionless deposits that are very loose to moderate in relative density (Table 1) lateral spreads typically occur farthest from the meizoseismal region (if a stream bank is nearby at the time of the earthquake). The factors that determine the most distant occurrence of lateral spreading have not been verified, but such spreading could result when a stream bank offers little or no resistance to lateral movement during shaking, or it could be because the youngest deposits typically border a stream where the water table is shallow. Dikes from surface oscillations (Youd, 1984) can develop considerably beyond the meizoseismal zone, especially where conditions are favorable for developing surface oscillations from surface waves (e.g., broad valleys, alluvium at least tens of meters in thickness, and flat-lying bedrock). Such dikes are likely developed from S waves far beyond the meizoseismal, even at sites of marginal liquefaction and relatively low accelerations, because of the tendency for surface oscillations to develop for S-wave vibrations with longer periods (Pease and O'Rourke, 1995). In most field situations of moderate liquefaction susceptibility, hydraulic fracturing seems to cause only small, scattered dikes to form beyond the meizoseismal zone, even for earthquakes in excess of M_w ~7.

The formation of dikes from lateral spreading predominates near the stream bank, and effects from hydraulic fracturing predominate with increasing distance from the bank (Obermeier, 1996). Farther from the meizoseismal zone, the influence of hydraulic fracturing is often minor.

The thickness of cap penetrated by hydraulic fracturing depends primarily on the thickness of sand that liquefies, apparently because the liquefied thickness controls the magnitude of pore-water pressure increase as well as the volume of water expelled; the thickness of the penetrated cap is also strongly dependent on the severity of ground shaking (Ishihara, 1985). Youd and Garris (1995) also found that dikes caused by hydraulic fracturing commonly are much lower in height than those formed by lateral spreading or surface oscillations. They estimate the maximum thickness of cap that has been observed to be ruptured by hydraulic fracturing is about 9 m.

In contrast, the maximum cap thickness that can be ruptured by lateral spreading is commonly much greater than that ruptured by hydraulic fracturing (see Youd and Garris, 1995, Fig. 3). The maximum reported is ~16 m (T.L. Youd, 1997, personal commun.). In many settings, the maximum thickness is controlled by the maximum depth of liquefaction, because of the low tensile strength of the cap in relation to the stresses imposed on it by gravity and seismic shaking. Caps of Holocene and late Pleistocene ages composed of silt and clay sediments typically have very low tensile strengths and thus are easily pulled apart by lateral spreading.

The formation of lateral spreads is not nearly as dependent on the thickness of the liquefied zone as is the formation of hydraulic fracturing. Lateral spreads have not been observed to form on liquefied sand strata only a few centimeters in thickness (J.R. Keaton, 1993, personal commun.). Lateral continuity of the liquefied bed is especially important for lateral spreading, particularly on such a thin stratum.

The thickness of cap ruptured by surface oscillations commonly is greater than that ruptured by hydraulic fracturing (Youd and Garris, 1995), and the effects of oscillations tend to extend much farther from the meizoseismal zone. In general, though, breakage of the cap by surface oscillations is localized away from the meizoseismal zone, even for a very large earthquake.

No data are available concerning the role of thickness of liquefied sediment in development of surface oscillations, although we suspect that 1 m or more suffices for typical fluvial sands, at least near the meizoseismal region of a very large earthquake. This suggestion is based on field observations in the Wabash Valley of Indiana and Illinois (Fig. 1). Preliminary data (Obermeier) also indicate that parallel joints in the cap can develop from seismic shakings, even where no liquefaction has occurred, providing that shaking has been strong enough. Joints from other mechanisms such as weathering and desiccation are commonly much more haphazard and discontinuous than those of seismic origin.

In the previous section we noted that formation of liquefaction features depends on depth to the water table. The influence of water table depth seems to be very dependent on the mechanism primarily responsible for rupturing the cap. Obermeier's data indicate strongly that cap breakage by hydraulic fracturing can be much more sensitive to depth of the water than is cap breakage by lateral spreading.

Field examples of manifestations of liquefaction. Evidence of liquefaction-related ground-failure mechanisms is apparent in aerial photographs of the meizoseismal zone of the great 1811–1812 New Madrid (Missouri) earthquakes (Figs. 1, 2). Within a time span of only three months, numerous strong earthquakes struck along a more than 175-km-long fault zone. One earthquake was probably nearly M_w 8, and two more were nearly as large (Johnston and Schweig, 1996). The earthquakes were centered beneath a huge region of liquefiable deposits and caused tremendous liquefaction. Sand that vented to the surface formed a veneer more than 0.5 to 1 m thick over hundreds of square kilometers. More than 1% of the ground surface was covered by vented sand over thousands of square kilometers (Obermeier, 1989). The meizoseismal zone of the 1811–1812 earthquakes is one of the best in the world to see the effects of liquefaction in both plan and vertical views. The vertical view permits the observer to see dikes that pinch together and do not reach the surface, and also permits viewing of dikes that were later buried by sediments or have been weathered so severely as to not be observable at the surface.

Fissuring and venting during 1811–1812 took place in braid-bar deposits of latest Pleistocene age and in Holocene point-bar

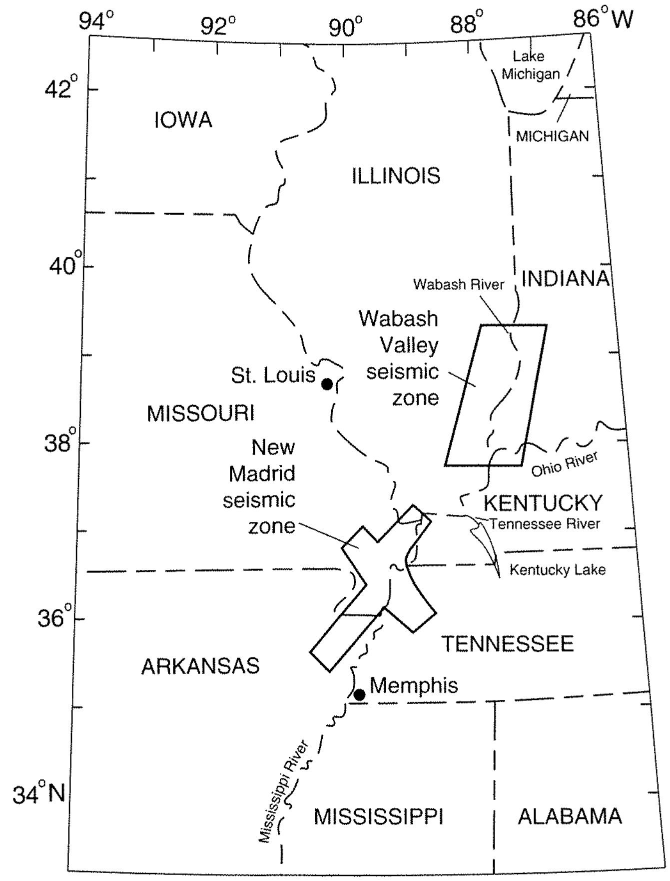

Figure 1. Approximate limits of the New Madrid seismic zone and the Wabash Valley seismic zone. The New Madrid seismic zone is the source area of the great New Madrid, Missouri, 1811–1812 earthquakes; the region continues to have many small and some slightly damaging earthquakes. The Wabash Valley seismic zone is a weakly defined zone of historic seismicity having infrequent small to slightly damaging historic earthquakes.

deposits (Fig. 2). The ground surface there is typically flat, except at stream banks that generally are only several meters high. The light-colored parts of the photos show sand vented to the surface. The dark background is the dark-colored, clay-rich cap onto which the sand vented. The light-colored linear features are long fissures through which sand vented, and the light-colored spots are individual sites of venting. Venting occurred through dikes.

Note the abundance of linear fissures that are more or less parallel to one another in the upper right side of Figure 2. These fissures are of lateral spreading origin and formed near a break in slope. The photo clearly shows that lateral spreading in the area was severe at distances farther than 0.5 km from the stream. Individual sand blows, which are particularly well expressed in the upper left part of the photo, were formed by hydraulic fracturing, clearly indicated by the random "shotgun pattern" of the sand blows.

Hydraulic fracturing can also follow geologic details, as illustrated in the lower part of Figure 2. Sand blows here developed in point bar deposits, as illustrated by the arcuate bands of vented sand. The venting occurred along the crests of scrolls of point bar deposits, where the cap is thinnest (Saucier, 1977). A venting origin for the sand is demonstrated by the irregular, jagged patterns of sand along the arcuate bands, which precludes the possibility of the sand being visible at the surface simply because of the absence of a fine-grained cap along the crest of the scroll.

The development of lateral spreads and individual sand blows is typical of that throughout the meizoseismal region of the 1811–1812 New Madrid earthquakes in that the dikes from lateral spreading commonly extend more than 1 km from any breaks in slope, and the isolated sand blows developed throughout the area, independent of proximity to a stream bank.

There is a widespread perception that wide dikes that form by lateral spreading are restricted to areas very near stream banks. However, in the meizoseismal region of the 1811–1812 New Madrid earthquakes, dikes from lateral spreading as much as 0.5 m wide are plentiful even hundreds of meters from any significant slopes. In another example, in the Wabash Valley of Indiana and Illinois, within the meizoseismal zone of a prehistoric M_w ~7.5 earthquake, dikes up to 0.5 m in width probably formed hundreds of meters from any stream banks when the earthquake struck (Munson and Munson, 1996; Pond, 1996; Obermeier, 1998a).

In both the New Madrid and Wabash regions, liquefaction susceptibility is only moderate at most places (Obermeier, 1989; Pond, 1996) and is probably typical of medium-grained, moderately well graded fluvial deposits elsewhere. Data from a worldwide compilation of historical earthquakes by Bartlett and Youd (1992) clearly show that horizontal movements of meters commonly extend hundreds of meters back from stream banks, especially where seismic shaking has been strong.

The probable explanation for the exceptional development of lateral spreading from the 1811–1812 New Madrid earthquakes is that there were very high levels of shaking, caused by large drops in stress in bedrock at depth (Hanks and Johnston, 1992). A major point of relevance, indicated in Figure 2, is that the severity of lateral spreading, including the distance of development from stream banks, can be an indicator of the severity and duration of strong shaking.

WHERE WAS THE SOURCE REGION?

Verification of a seismic origin for suspected liquefaction features typically involves demonstrating that (1) details of individual clastic dikes conform to those of known seismic origin, (2) both the pattern and location of dikes in plan view conform with a seismic origin, on a scale of tens to thousands of meters, (3) the size of dikes on a regional scale identifies a central "core region" of widest dikes, which conforms with severity of effects exposed in the energy source region (the meizoseismal zone), and (4) other possible causes for the dikes, such as artesian conditions and landsliding, are not plausible (Obermeier, 1996, 1998a). As we use the term "core region," we are referring to the region of strongest bedrock shaking. We also

Figure 2. This aerial photograph shows effects of severe liquefaction in the meizoseismal zone of the great 1811–1812 New Madrid (Missouri) earthquakes. White linear features show sand that has vented through breaks in proximity to stream breaks. Isolated white spots are sand that has vented through breaks in the cap by hydraulic fracturing. Black box on map of Arkansas indicates location of photo area.

refer interchangeably to this region throughout the paper as the "source region" or the "energy center."

Two methods have been used to estimate the source region in an excellent study example, in the Wabash Valley (Obermeier et al., 1993; Munson and Munson, 1996; Pond, 1996). Both methods have widespread applicability for paleoseismic studies. One involves direct measurement of dike widths and the other involves back-calculating the strength of shaking. Both require collecting data over a large region in order to see a clear-cut trend in the data. In practical terms, for an earthquake of M_w 6 to 7, the data must be collected over an area of several to many tens of kilometers in radius. Preferably, data are from the region of distal effects of liquefaction, where dikes are small (narrow) and sparse, and also from the area close to the source region, where dikes are much larger (wider) and more plentiful.

Dike width serves as a superior parameter to locate the source region in many field settings (Obermeier, 1996). This width generally reflects the amount of lateral spreading except where dikes are relatively small (say, less than 10 cm wide). Conceptual verification for using dike width to locate the source region is provided from a study of historical earthquakes by Bartlett and Youd (1992). Dike width works well because the development and magnitude of lateral spreading are largely independent of thickness and strength of the cap, at least for sediments that are typical of the Holocene and late Pleistocene. Maximum dike width and the sum of dike widths at a site appear to work equally well to estimate the source zone (Munson et al., 1995). A valid interpretation based on the widths of dikes obviously requires that bank erosion has not been so severe as to have destroyed dikes by lateral spreading. Problems of interpretation due to erosion are generally not serious in the meizoseismal zone of a very large magnitude earthquake because of the tendency for large lateral spreads to develop even relatively far from the stream banks.

Data from historical earthquakes in the Wabash Valley region, in the forms of modified Mercalli intensities and instrumentally located epicenters (Rhea and Wheeler, 1996), suggest that using liquefaction features to locate the source region of prehistoric earthquakes is generally accurate to within a few tens of kilometers, at least for earthquakes of moderate size. The uncertainty in location probably increases with increasing magnitude because of the tendency for the epicenter of larger earthquakes to be farther removed from the area of strongest shaking (e.g., Youd, 1991). Still, it appears that the distribution and severity of liquefaction effects can be used to reasonably estimate the region of strongest bedrock shaking (Pond, 1996).

Other parameters, such as the density of dikes per unit length and density of dikes per unit area, have been used by other researchers in their attempts to locate the region of strongest shaking. There are numerous practical problems in using such an approach to interpret the data, however, because dike density is controlled by different factors for each of the mechanisms of lateral spreading, surface oscillations, and hydraulic fracturing. In many field situations it is impossible to determine which mechanism(s) controlled the density of dikes. Interpretations can be questionable without such a differentiation.

Back-calculation of the strength of shaking at widespread sites can sometimes be used to better locate the source region where dike-width data are sparse. This back-calculation procedure has been verified by comparing this interpretation with that of the dike-width method discussed above; both yielded the same results (Pond, 1996).

A question often asked is whether paleoliquefaction features resulted from a single large earthquake or from a series of small earthquakes that were closely spaced in time. The answer is generally best resolved by analysis of the regional pattern of dike widths. The attenuation pattern of maximum dike widths around a core region should be examined in orthogonal coordinates (preferably along the suspected fault axis and perpendicular to the axis). A monotonic decrease of maximum dike width in orthogonal directions around the suspected core indicates a single large earthquake. In the Wabash Valley, this approach was verified by geotechnical back-calculations of the prehistoric strength of shaking for four prehistoric earthquakes (Pond, 1996). The use of dike widths alone to resolve the issue of the number of events requires that the liquefaction susceptability be reasonably uniform on a regional basis and also that the amplification or attenuation of bedrock motions be similar on a regional basis.

To answer the question of whether a single earthquake or multiple earthquakes caused the observed features, the methods described above usually work best for very large earthquakes, because of the tendency for the regional pattern of liquefaction to become more conspicuous with increasing magnitude. For example, the regional pattern of dike sizes and abundance, in conjunction with radiometric dating, has been used in coastal South Carolina to show that liquefaction effects from prehistoric earthquakes were caused by very large earthquakes rather than multiple small earthquakes closely spaced in time (Obermeier, 1993; Obermeier, 1996). More recently, using basically the same logic, the regional pattern of severity of venting has been used to evaluate whether paleoliquefaction features discovered within the meizoseismal region of the great 1811–1812 New Madrid earthquakes were from a few very large earthquakes rather than a series of much smaller earthquakes (Tuttle, 1999). The New Madrid region is nearly ideal for this type of analysis because the liquefaction susceptibility is remarkably uniform over a huge area, the causative fault system for major earthquakes is likely known, and the regional pattern and extent of liquefaction from the 1811–1812 earthquakes has long been known reasonably well (e.g. Obermeier, 1989).

Using the paleoliquefaction method for determining the timing and strength of shaking of various earthquakes within a relatively small region works best where the large earthquakes are spaced apart sufficiently in time to distinguish different generations of liquefaction features. The techniques for sorting these generations have been developed mainly in a classic study in the Wabash Valley region by Munson and Munson (1996). Their approach is well suited for many field settings and typically uses radiocarbon dating, depth of weathering, pedology, sediment stratigraphy, and archaeological artifacts, in conjunction with the regional pattern of sizes of liquefaction features. They also were the first to note that sand deposits that had been vented to the surface were especially valuable as sites for narrowly bracketing when liquefaction occurred; the vented sand typically formed slightly elevated, dry sections in lowland areas that were otherwise wet and muddy much of the year. The vented sand deposits were frequented by Native Americans, who commonly left behind on the vented sand hearths and artifacts that can be used for dating.

DID STRONG SHAKING OCCUR WITHOUT LEAVING LIQUEFACTION EVIDENCE?

There is a common perception that liquefaction can occur in a region but leave behind no evidence. Discovering effects of liquefaction in the field is usually easy where liquefaction has been severe throughout a region, but it may be difficult where liquefaction has been marginal or highly localized. Below we present some of the major factors that determine the severity of liquefaction.

Effects of strength of shaking and liquefaction susceptibility

Our approach of relating occurrence and severity of liquefaction effects to modified Mercalli intensity (MMI) is used because MMI correlates strongly with both severity of liquefaction and damage to human-made structures (Wood and Neumann, 1931). MMI also correlates reasonably well with peak surface acceleration (Krinitzsky and Chang, 1988). Seed et al. (1985) reported similar correlations developed in China for M_w~7.5 earthquakes. The Chinese correlations emphasize higher earthquake magnitudes than those of Krinitzsky and Chang, whose relations are for a much wider range of magnitudes (Table 2). Relations below by Krinitzsky and Chang (1988, Fig. 7) are for sites in the "far field," which are removed from the region of strongest shaking.

Throughout the meizoseismal region of a very strong earthquake, in which the MMI value is IX or higher, liquefaction features should abound even where the liquefaction susceptibility is only moderate. Any reasonable effort to locate numerous liquefaction features should be successful. Some wide dikes almost certainly exceeding 0.3 m and many small dikes should be discovered.

For moderate liquefaction susceptibility in regions of MMI VII–VIII, small liquefaction features may be sparse but still should be numerous enough that some features would be discovered during the examination of tens of kilometers of stream banks.

Moderate liquefaction susceptibility implies medium relative density (Table 1) as well as a water table within several meters of the surface and a cap thickness less than 8 or 9 m. Moderate liquefaction susceptibility is about the norm for deposits of latest Pleistocene and Holocene age that have been laid down by moderate to large streams in the central and eastern United States. This level of susceptibility applies to streams of both glaciofluvial braid-bar and Holocene point-bar origins. A lower limit of moderate susceptibility requires a bed of silty sand, sand, or gravelly sand (generally less than about 40% gravel) that is at least a few to several meters thick and is capped by at least 0.5 m of lower permeability sediment. Where a cap is underlain by medium-grained sand or coarser sediment, the water table should be at or above the base of the cap at the time of the earthquake; otherwise, unless liquefaction occurs through a large thickness of sediment and has made available a large quantity of water, the high permeability of the material beneath the cap can permit dissipation of pore-water pressure induced by shaking, leaving no evidence of liquefaction.

Effect of grain size

Tsuchida (1970) recognized that liquefaction features predominate in sands containing little or no gravel or fines; uniform fine clean sands are the most susceptible to liquefaction. Since 1970, liquefaction features have been documented in nearly all cohesionless soils, including sandy gravels, silty sands to sandy silts, cohesionless silts, and tailings sands and slimes (e.g., Obermeier, 1996).

Gravelly sand and sandy gravel (as much as 60% gravel content—perhaps even more) can liquefy and form large dikes during earthquakes in conditions of strong shaking, impeded drainage, and a water table near the ground surface (Meier, 1993; Yegian et al., 1994). It appears that the presence of a fine-grained cap controls the formation of liquefaction features in gravel-rich deposits. Both historic (Harder and Seed, 1986; Andrus, 1994; Yegian et al., 1994) and prehistoric liquefaction features have been observed in gravelly soils with caps. Even a thin, fine-grained cap can impede drainage and allow the pore-water pressure to increase during shaking, but it seems likely to us that the areal extent of the cap must also be large in order to prevent dissipation of pore-water pressure during shaking. Earthquake magnitudes of M_w~7 to 7.5 and shaking levels lower than 0.4 to 0.5 g were adequate to trigger liquefaction in many of the cases cited above. However, very gravel-rich deposits without fine-grained caps can withstand strong shaking (on the order of 0.5 to 1.0 g) without forming liquefaction features (Yegian et al., 1994).

Back-analysis of liquefaction cases involving gravelly soils is complicated because of the effect of gravel on the measurement of penetration resistance. Tokimatsu (1988) showed that the penetration resistance for soils with a small percentage of gravel can be artificially increased compared to that of a clean sand at

TABLE 2. CORRELATIONS BETWEEN MODIFIED MERCALLI INTENSITY (MMI) AND PEAK SURFACE ACCELERATION

MMI	Peak surface acceleration (g) Chinese	Krinitzsky and Chang (1988)
VII	~0.1	~0.13
VII	~0.2	~0.2
IX	~0.35	no data

the same relative density and confining pressure, because of the large size of the gravel particles relative to the size of the penetration equipment. Tokimatsu (1988) tentatively suggested a reduction factor to correct the SPT blow count of gravelly soils (based on mean grain size) to that of sandy soil for use in liquefaction analyses. However, the application of such a correction factor raises uncertainty in any back-analysis.

Cohesionless silt will also liquefy and fluidize to form dikes, sometimes extensively (Youd et al., 1989). "Dirty" sands containing as much as 85% fines (silt and clay) have been observed to liquefy (Bennett, 1989), but soils with more than 15% to 20% clay content (<0.005 mm) are unlikely to liquefy (Seed et al., 1983). The effect of fines on liquefaction susceptibility has not been completely resolved, and numerous apparently conflicting data and opinions exist in the literature. The effect of fines on the susceptibility can be separated into two categories: (1) effect on liquefaction resistance of the soil, and (2) effect on penetration resistance.

Recent studies have indicated that the effect of fines on liquefaction susceptibility depends on the nature of the fines (i.e., plasticity and cohesion). Cohesionless silts (e.g., tailings slimes) and some sands with cohesionless silt contents as high as 30% may be more susceptible to liquefaction than clean sands (Ishihara, 1993; T.L. Youd, 1997, personal commun.). In addition, Yamamuro and Lade (1998) noted that at low overburden pressures, uniformly sized sand with a low cohesionless silt content is more likely to collapse and liquefy than the same sand containing no silt; Yamamuro and Lade hypothesized that the silt grains cause the silty sand to form a more honeycombed structure during deposition compared to a clean sand, even at the same global relative density. This causes the silty sand to be more susceptible to collapse and pore-water pressure increase upon shearing.

Field observations vary concerning the influence of silt content on liquefaction effects. In the western United States, M.J. Bennett (2000, personal commun.) has observed that silty sands and sandy silts are more susceptible to liquefaction than clean sands. It has been the experience of Obermeier, however, that dikes and liquefaction-induced features involving silty sands (say, 20% to 30% fines or more) are only rarely observed in paleoliquefaction searches in the central and eastern United States, even where shaking has been very strong; yet nearby, liquefaction features involving clean sand sources are commonly abundant.

Effect of depth to water table

The depth to the water table has a profound effect on the liquefaction susceptibility of a sand deposit and can also have an important bearing on the ground-failure mechanism that develops. Where the water table is more than 4 to 5 m deep, it appears that severe effects of liquefaction, especially due to hydraulic fracturing, become greatly suppressed and can be scarce even where shaking is moderate (~0.2 g).

It is commonly observed that dikes from lateral spreading are the only ones seen in an exposure. Levels of shaking for this situation probably can be as high as 1/4 g in field settings where the source sands are fluvial in origin, medium-grained, moderately well-graded, and moderately compact.

In general, if the water table is ≥10 m below the ground surface, formation of liquefaction features from any failure mechanism is highly unlikely, unless shaking is severe and the field setting is conducive to their formation.

Locating the depth to the water table at the time of the earthquake is very important in estimating the strength of shaking. For clean sands, fine-grained and coarser, this depth can be estimated by observing the highest level of the base of dikes (i.e., at the base of the fine-grained cap at widespread sites). In field situations where the water table is much lower that the base of the cap, for low to even moderate severity of liquefaction, the high permeability of these clean sands would probably allow dissipation of excess pore-water pressure along the base of the cap, thereby precluding the formation of dikes in the cap.

Long bank exposures over a large region, at least kilometers in extent, in which the contact of the fine-grained cap with underlying sand can be observed are especially valuable for the approach discussed above. Where bank exposures are limited in length or in regional extent, confidence in the interpretation of the depth of the water table is increased by measuring the relative liquefaction susceptibilities of sand at various depths. Obermeier et al. (2000) discussed how these factors were incorporated to evaluate the depth of water table in a study area during the 1811–1812 New Madrid earthquakes.

HOW STRONG WAS THE PALEO-EARTHQUAKE?

Much progress has been made in the past few years in the development of techniques to back-calculate strength of shaking and magnitude of paleo-earthquakes. Four methods, each distinct from the rest, and which we believe are especially relevant or promising are: (1) the magnitude bound method, which uses the farthest distance of paleoliquefaction features from the tectonic source to estimate magnitude; (2) the cyclic stress method, which estimates the lower bound peak accelerations at individual sites of liquefaction, and which can be used in conjunction with the regional pattern of acceleration attenuation to estimate the actual magnitude of prehistoric earthquakes; (3) energy-based solutions, which use fundamental parameters of earthquake strength and soil susceptibility to liquefaction, and (4) the Ishihara method, which uses dike height at a site of hydraulic fracturing to estimate the peak acceleration. The first two methods are applicable to many field and tectonic settings, and though existence of these methods is known by many, their strengths and limitations are not widely appreciated. The latter two methods are still in development, but can be useful. Selection of the appropriate method(s) depends on the data available at the field sites, as noted below. Much more detailed discussion for each of the four methods is given in Obermeier et al. (2001).

The first two techniques have been used to determine the prehistoric levels of shaking in the Wabash Valley region of Indiana and Illinois (Munson et al., 1997; Pond, 1996). For comparison, we also used solution still in development, the energy-stress method of Pond (1996). The Wabash Valley region lies in an area of intraplate seismicity in which the largest historical earthquake (during the past 200 yr) has been M_w 5.8. Paleoliquefaction features clearly demonstrate, however, that numerous and much larger Holocene earthquakes have been centered in the region, on the basis of sizes of liquefaction features and regional extent of liquefaction from these earthquakes. This region is typical of many where paleoliquefaction interpretations are especially useful—i.e., there are no surface faults available for study, and the prehistoric earthquakes are spaced widely enough in time to separate their liquefaction effects from one another.

Evaluation of the prehistoric levels of shaking in the Wabash Valley presents challenges, however, because of the absence of seismological records of large earthquakes in the region. The record is available only for small earthquakes (M_w <5). The behavior of the smaller earthquakes has been extrapolated to predict the behavior of much larger events for some of the analyses we discuss in this section, but such an extrapolation may not reflect reality. Similar uncertainty exists in most regions where paleoliquefaction studies have been used as a basis for interpreting the prehistoric record (i.e., central and eastern United States). Unknown seismic factors in the Wabash Valley region include the stress drop, which can have a large effect on the strength of shaking (Hanks and Johnston, 1992) and possibly other factors, such as strength of shaking at various frequencies, in which some frequencies may be too high to induce shear strains large enough to cause liquefaction. A deep focal depth can cause the strength of shaking to be diminished at the ground surface, above the focus. Unlithified sediment of considerable thickness (hundreds to thousands of meters) above bedrock may alter the severity and/or frequency of shaking as it is transmitted from the bedrock.

A preferred orientation of strong shaking (i.e., directionality; see Appendix) can also complicate interpretations of prehistoric strength of shaking. For strike-slip faulting, the effects of directionality can be manifested as higher accelerations along the projection of the fault axis. Other types of faulting have other types of directionality effects. However, a paleoliquefaction search that encompasses a large region should clarify effects of directionality, permitting proper use of back-analysis of strength of shaking and magnitude. These procedures for back-calculations are based on techniques that provide only maximum levels of shaking as a function of earthquake magnitude and distance from the energy center, regardless of orientation from the energy center; this requires, therefore, that back-calculations for paleoseismic interpretations determine the highest level of (bedrock) shaking as a function of distance from the energy center.

The confidence in interpretations of prehistoric levels of shaking is highest where different procedures for back-calculation yield the same results. Even in this case though, there can be some uncertainty because some of the methods may depend similarly on assumed parameters such as stress drop and focal depth. Evaluations of prehistoric magnitudes for four large paleo-earthquakes in the Wabash Valley region are given in Table 3. It is obvious from the table that for each of the paleo-earthquakes, the back-calculated earthquake magnitudes using the different methods are very close to one another. The implication is strong that the magnitude has been reasonably approximated for each of the paleo-earthquakes, at least in terms of destructive potential. This example is the only major extant study for which the various methods have been used for back-analysis, and it is a landmark effort.

Magnitude bound method

The magnitude bound method estimates the magnitude of a paleo-earthquake by using relations between earthquake magnitude and the distance from the tectonic source to the farthest site of liquefaction. The method is based on increasingly stronger earthquakes causing liquefaction at increasing distances from the energy center, in a systematic manner.

Distances from energy centers to the farthest liquefaction features for many worldwide, historical earthquakes have been compiled by Ambraseys (1988). The sites of farthest effects were from locales of minor venting of sand or minor lateral spreading. Sites having only soft-sediment deformation, such as ball-and-pillow structures, load casts, or convoluted bedding, were not included in the data set. The data are from various tectonic conditions and susceptibilities to liquefaction, so it is not surprising that the maximum extent of liquefaction is highly variable for a given earthquake magnitude. This variability makes it essential that the technique be calibrated for the tectonic setting of interest, preferably by using data on the extent of liquefaction from historical earthquakes in the study area, to account for the influence of local factors such as stress drop, focal depth, and liquefaction susceptibility.

Where both the energy source and the outer limits of liquefaction of a paleo-earthquake are well defined and effects of liquefaction from historic earthquakes are available for calibration, a reasonable estimate of prehistoric magnitude can be achieved in many study areas.

TABLE 3. BACK-CALCULATED MAGNITUDES FOR FOUR LARGEST PALEO-EARTHQUAKES CENTERED IN WABASH VALLEY REGION OF INDIANA AND ILLINOIS

	Paleoearthquake			
Back-analysis method	Vincennes	Skelton	Vallonia	Waverly
Date*	6100 yr	12 000 yr	3950 yr	mid-Holocene
Magnitude bound	7.8	7.2	6.9	6.8
Cyclic stress	7.5–7.7	7.4	6.7	6.9
Energy stress	7.5–7.8	7.3	7.1	6.8–7.1

Note: Data from Pond (1996), Munson and Munson (1996), Munson et al. (1997), and Obermeier (1998a).

*Radiocarbon dates, from Munson and Munson (1996).

Cyclic stress method

Seed et al. (1985) updated a procedure originally proposed by Seed and Idriss (1971) to evaluate the liquefaction susceptibility of sandy soils. The procedure is based on case histories of sites that did or did not develop liquefaction effects during earthquakes worldwide. The occurrence of liquefaction was judged from many types of observations, such as sand blows caused by hydraulic fracturing, lateral spreading, ground cracking or settlement, and damage to structures caused by settling or tilting.

The Seed et al. (1985) method and its predecessors were originally developed as a geotechnical procedure to estimate the strength of cohesionless sediment required to prevent liquefaction during an earthquake, for a given earthquake magnitude and peak acceleration. The method is based on comparing the earthquake-induced (horizontal) cyclic shear stress to the cyclic resistance of the soil (i.e., to the strength of the soil or its resistance to pore-water pressure buildup). The earthquake-induced cyclic shear stress is related to both the strength and duration of shaking. These values, in turn, statistically relate to earthquake magnitude. The influence of the seismically induced horizontal shear stress is incorporated within the parameter of cyclic stress ratio (CSR); CSR is a function of earthquake magnitude, peak surface acceleration, the total and effective overburden stresses, and the depth of the source bed. The strength of the soil is evaluated in terms of the parameter $(N_1)_{60}$, which is the SPT blow count (*N*) normalized to account for depth of sediment and the water table, as well as for the specific type of SPT test equipment.

The cyclic stress method has been developed from a large database of historic earthquakes having magnitudes $\leq M_w$ 7.5; the database has only limited data for larger earthquakes. Using a technique for analysis developed recently by Pond (1996), we can employ the cyclic stress method to estimate a magnitude for paleoearthquakes of $M_w \leq 7.5$, which is useful for hazard assessments.

Energy-based approaches

Energy-based approaches to liquefaction analysis are inherently appealing because moment magnitude M_w is a direct measure of seismic energy. Such approaches are all the more attractive because energy is a fundamental physical parameter. Still, energy-based approaches are not at the state of development to be used for routine analysis, although some will doubtless be so in the near future. Two approaches have been used for liquefaction analysis, one based on field case histories and the other based on laboratory testing of sediment from the site of interest.

Field case histories using Gutenberg-Richter relations. Davis and Berrill (1982) first developed an energy-based approach for predicting liquefaction from field data. Similar and extended approaches attempting to relate sediment properties at a site to the Gutenberg-Richter (Gutenberg-Richter, 1956) function for energy at the site were proposed by Berrill and Davis (1985), Law et al. (1990), and Trifunac (1995). Well-defined relations were not observed throughout the distance from the tectonic source. Part of the scatter almost certainly originates from the empirical Gutenberg-Richter (1956) function as a measure of radiated energy, *E*, defined as

$$E \approx \frac{10^{1.5M}}{R^2}, \qquad (1)$$

where M_w is moment magnitude, and *R* is either the epicentral distance (Trifunac, 1995) or the distance from the energy center (Davis and Berrill, 1985). As pointed out by Trifunac (1995, 1999), much of the scatter is probably because the model does not account for seismic source mechanisms, directionality of strong motions, or local geologic conditions. It is also likely that part of the scatter is caused by use of epicentral distance rather than distance from the energy center, especially for larger earthquakes. However, we suggest that because body waves are dominant in the development of liquefaction and because surface waves (despite the fact that they can play a large role in the breakup of a cap and determine whether venting takes place at the ground surface) do not extend below shallow depths, and are unlikely to have much influence in the development of liquefaction at many places.

Near the ground surface, breakdown of the sediment grain-to-grain contacts may lead to a considerable loss of energy. This may be the case for some liquefiable deposits; if so, this energy loss could be a source of serious error when the function *E* is used for analysis.

Laboratory test results. Laboratory testing has clearly demonstrated that a direct relation exists between dissipated energy and buildup of pore-water pressure during undrained cyclic shearing of saturated sands (Nemat-Nasser and Shokooh, 1979; Simcock et al., 1983; Liang et al., 1995; Green et al., 2000). Building on this observation, several researchers have attempted to use laboratory data to correlate normalized energy capacity (i.e., capacity per unit volume, accounting for influence of initial effective confining stress) and relative density for various types of soils (e.g., Al-khatib, 1994; Ostadan et al., 1998). In these studies, normalized energy capacities were computed as the area bounded by the stress-strain hysteresis loops, up to the point of liquefaction. It is now well demonstrated that the normalized energy capacity is a fundamental parameter for evaluating the liquefaction potential of reconstituted samples that are tested in the laboratory (Green, 2001).

Reconstituted laboratory samples such as those used in the studies cited above cannot be used directly for evaluation of the liquefaction potential of naturally occurring samples in the field, because of differences in deposition, overconsolidation, preshearing, or aging (i.e., Terzaghi et al., 1996; Olson et al., 2001). This problem can be circumvented for important projects, including paleoseismic studies, by conducting the laboratory tests on undisturbed frozen samples. However, obtaining undisturbed field samples is very expensive, and therefore this technique has seen limited use.

Arias Intensity method. Kayen and Mitchell (1997) extended preliminary work of Egan and Rosidi (1991) and developed correlations to predict the occurrence of liquefaction as functions of the Arias Intensity of the earthquake motion and penetration resistance of the soil. Arias Intensity, I_h, is defined by the equation

$$I_h = I_{xx} + I_{yy} = \frac{\pi}{2g}\int_0^t a_x^2(t)dt + \frac{\pi}{2g}\int_0^t a_y^2(t)dt\ , \quad (2)$$

where I_{xx} is the intensity value in the x-direction in response to transient motions in the x-direction, I_{yy} is the intensity in the y-direction, g is the acceleration due to gravity, t is the duration of shaking, and $a_x(t)$ and $a_y(t)$ are the transient accelerations of earthquake motion in the x- and y-directions, respectively. In this approach, the energy applied to the soil (I_{hb})is:

$$I_{hb} = (I_{xx} + I_{hh})r_b\ , \quad (3)$$

where r_b is a depth reduction factor that accounts for the variation of Arias intensity with depth.

However, this technique does not explicitly account for the influence of effective confining pressure (e.g., depth of water table), and, as Trifunac (1999) noted, there are basic questions concerning whether the Arias Intensity function represents the actual energy input into an element of soil. Still, the correlations that Kayen and Mitchell (1997) developed appear very good and may be suitable for paleoliquefaction analysis in field situations where the water table was shallow at the time of the earthquake. Acceleration time-history relations that can be used to express Arias Intensity as a function of earthquake magnitude and site-to-source distance are currently available for only a few regions of the world, but this problem possibly may be avoided by using Arias Intensity attenuation relationships as suggested by Kayen and Mitchell (1997) for the western United States and as used tentatively by Schneider (1999) in the central United States.

Ishihara method. For paleoseismic analysis, the Ishihara (1985) method is a technique to estimate peak acceleration at sites of paleoliquefaction. The premise of the method is that the maximum height of dikes (accompanied by venting at the surface) is controlled by two factors: the thickness of liquefied sediment and the peak acceleration. Ishihara (1985) originally developed the bounds using data from only a few earthquakes with magnitudes on the order of ~7.5 and higher, and only limited data have since been added for such large earthquakes. Youd and Garris (1995) showed that the method is not valid for ground failures due to lateral spreading or surface oscillations. The relationships for the Ishihara method probably represent sites where surface effects of liquefaction from hydraulic fracturing were abundant—i.e., liquefaction was severe (T. L. Youd, 1998, personal commun.).

Pease and O'Rourke (1995) also critiqued the Ishihara (1985) method for the M_w 7.1 Loma Prieta earthquake of 1989. They found that the method correctly predicted occurrences of surface effects of liquefaction except at sites of lateral spreading or surface oscillation. Pease and O'Rourke (1995) did not present detailed data concerning the properties of the liquefied source sands, but most appeared to have been loosely compact and some were moderately compact.

The Ishihara (1985) method may be applicable where the cap thickness is reasonably uniform (or at least does not slope much along its base) and for source sands ranging from very loose to moderately compact, at least for M_w~7.5 or larger earthquakes. For lower magnitudes, the method likely applies only for loose deposits. More detailed data regarding site-specific parameters are needed to critique the method more fully.

The Ishihara (1985) method has great potential for paleoseismic analysis at sites where the ground failure can be attributed confidently to hydraulic fracturing. The method is ideally suited for using measurements of dike height and cap thickness, which are observable along stream banks found in many paleoliquefaction searches.

SUMMARY AND COMMENT

The extensive reliance on paleoliquefaction studies in the central and eastern United States is due partly to the abundance of stream valleys in this humid environment containing liquefiable deposits. Similarly, it has been found that such liquefiable deposits occur throughout much of the humid and rainy U.S. Pacific Northwest, revealing the paleoseismic record through at least much of Holocene time (Obermeier and Dickerson, 2000). However, adequate streams are available for paleoliquefaction studies even in many arid conditions. Overall, throughout much of the United States and in many field settings worldwide, there are adequate liquefiable deposits to reveal the record of strong Holocene seismicity.

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APPENDIX: DEFINITION OF TERMS

Directionality—The transmission of seismic energy, as manifest in parameters such as maximum acceleration, in a preferred orientation in plan view. The influence of directionality depends on the orientation and type of fault. For example, strike-slip faults tend to transmit more energy along the plane of the fault; normal faults tend to transmit more to the downthrown block; and reverse faults tend to transmit more to the upthrown block.

Meizoseismal zone—The American Geological Institute *Glossary of Geology* defines "meizoseismal" as "pertaining to the maximum destructive force of an earthquake," from which one could infer that "meizoseismal zone" means the area within, or approximately within, the highest isoseismal (as, for example, the area within the highest Modified Mercalli intensity). Others, however, use the term to refer to the region of higher intensities of the earthquake (note: not highest intensities). We use this term in the latter sense.

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