Earthquake recurrence inferred from paleoseismology

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Introduction

Earthquakes threaten the United States, as illustrated by hazard maps for the 48 conterminous states (Fig. 1). Much of the threat comes from unusually large earthquakes that recur hundreds or thousands of years apart. Engineering designs, insurance rates, and emergency plans depend on national maps that forecast seismic shaking at various probability levels (Fig. 1b). The study of prehistoric earthquakes – paleoseismology – provides long-term rates of earthquake occurrence to improve confidence in such forecasts.

Paleoseismology emerged in the last decades of the 20th century, after 1965. It draws on many kinds of research, including geomorphology, stratigraphy, structural geology, geochronology, paleoecology, oceanography, civil engineering, archaeology, ethnology, and documentary history. Its literature includes collected papers and workshop proceedings (Crone & Omdahl, 1987; Ettensohn *et al.*, 2002; Hancock & Michetti, 1997; Masana & Santanach, 2001; Ota *et al.*, 1992; Pavlides *et al.*, 1999; Serva & Slemmons, 1995; Shiki *et al.*, 2000; Yeats & Prentice, 1996), national and regional overviews (Camelbeeck, 2001; Clague, 1996; Grant & Lettis, 2002; Ota & Okumura, 1999; Research Group for Active Faults of Japan, 1992; Talwani & Schaeffer, 2001), topical reviews (Jacoby, 1997; Obermeier, 1996), textbooks (McCalpin, 1996; Noller *et al.*, 2000; Yeats *et al.*, 1997), and narratives intended for general audiences (Nance, 1988; Sieh & LeVay, 1998).

This chapter describes three North American examples of earthquake history inferred from Quaternary geology. The examples resemble one another by providing long-term perspectives unavailable from traditional seismological records. Each example includes multiple earthquakes inferred from widespread paleoseismic evidence. These earthquakes suggest rates and patterns of recurrence that help define earthquake hazards. The examples differ in tectonic setting, in the kinds of features that record prehistoric earthquakes, and in overlap with instrumental and written records.

Described first is evidence for infrequent surface rupture on faults in a small part of California's diffuse boundary between the Pacific and North America plates. The faults form a 50-km-wide shear zone east of the San Andreas fault. Collectively termed the eastern California shear zone, these faults accommodate lateral motion not absorbed by the San Andreas. Movement on some of them produced surface ruptures and two large, instrumentally recorded earthquakes in the 1990s. Prehistoric offsets exposed in trenches show that thousands of years probably separated such ruptures in the Holocene. Age ranges of the prehistoric ruptures overlap among the faults. These findings suggest that the shear zone produces large earthquakes in infrequent series.

Next we discuss earthquakes in the interior of the North America plate – in the New Madrid seismic zone of Missouri, Arkansas, and Tennessee. This region's low relief and slow rates of modern deformation belie a late Holocene history of large earthquakes more frequent than those in the eastern California shear zone. A series of three large earthquakes in 1811 and 1812, known from historical accounts, produced thousands of sand blows in an alluvial area at least 200 km by 80 km. Sand blows similarly record earlier series of New Madrid earthquakes in A.D. 800–1000 and 1300–1600.

Our final example comes from the Cascadia subduction zone, where oceanic lithosphere descends beneath the North America plate in California, Oregon, Washington, and British Columbia. Though unknown from this region's written history, great subduction earthquakes repeatedly lowered much of its Pacific coast by at least 0.5 m, most recently in A.D. 1700. The subsidence is marked by buried soils at estuaries. Such soils from the past 3500 years in Washington imply that the earthquakes recur at irregular intervals ranging from a few hundred years to about one thousand years.

Eastern California Shear Zone

The eastern California shear zone, centered about 150 km northeast of Los Angeles (Fig. 2), exhibits geologic evidence for prehistoric surface ruptures during episodes thousands of years apart.

Modern Deformation and Earthquakes

According to geodetic measurements, the shear zone absorbed right-lateral slip at 11–14 mm/yr during the 1990s (McClusky *et al.*, 2001; Miller *et al.*, 2001; Sauber *et al.*, 1994). This slip rate accounts for a quarter of the interplate motion, which averages 50 mm/yr (DeMets *et al.*, 1990, 1994).

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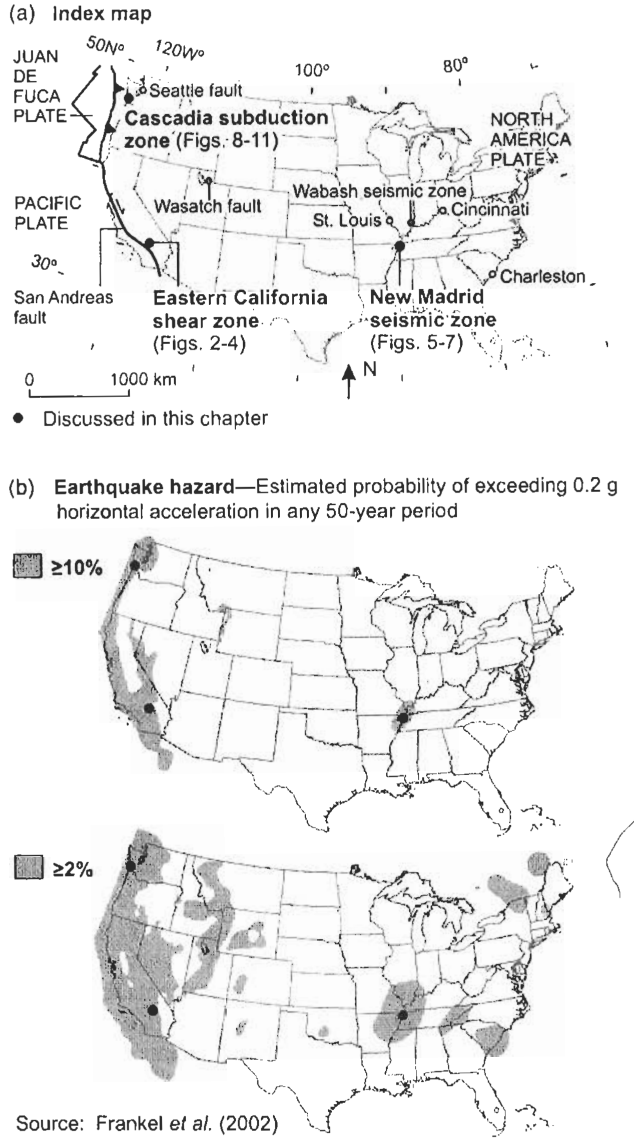


Acceleration of 0.2 g can cause partial collapse of ordinary buildings. Ground motion with a 10% probability of being exceeded in 50 years would be expected to be exceeded once in 500 years on average. For 2% in 50 years, the ground motion would be exceeded once in 2500 years on average.

Fig. 1. Overview of earthquake hazards in the conterminous United States.

The geodetic measurements coincided with a decade in which the eastern California shear zone produced two unusually large earthquakes (Figs 2 and 3). The first and largest, the 1992 Landers earthquake of moment magnitude (M) 7.3, ruptured several north- to northwest-striking right-lateral faults along a total length of 70 km (Sieh *et al.*, 1993). Within a few tens of seconds, rupture started on the southern Johnson Valley fault, progressed northward, slowed at stepovers to and from the Homestead Valley fault, and finally ended along the Camp Rock fault (Cohee & Beroza, 1994; Wald & Heaton, 1994). Coseismic dextral slip at the ground surface commonly exceeded 3 m; it reached a maximum of 6 m along the northern Emerson fault. Seven years later on a parallel trend 30 km to the northeast, the M 7.1 Hector Mine earthquake produced as much as 5 m of surface dextral slip on the Lavic Lake and Bullion faults (Treiman *et al.*, 2002). This 1999 earthquake also triggered small earthquakes over much of southern California (Hauksson *et al.*, 2002; Rymer *et al.*, 2002a).

The 1992 Landers and 1999 Hector Mine earthquakes have few historical precedents in eastern California. Before 1992, no large ($M > 7.0$) earthquakes had ruptured eastern California faults since the 1872 Owens Valley earthquake, centered 200 km north-northwest (Fig. 2a). Instead, the shear zone's largest events were moderate earthquakes of M_L 6.1 (1947 Manix), M_L 5.5 (1975 Galway Lake and 1979 Homestead Valley), and M 6.2 (1992 Joshua Tree; Fig. 2b and c). (M_L, local Richter magnitude, is similar to M in this size range.) All these earthquakes were exceeded in size by the 1992 M 6.5 Big Bear earthquake, an aftershock to the Landers earthquake.

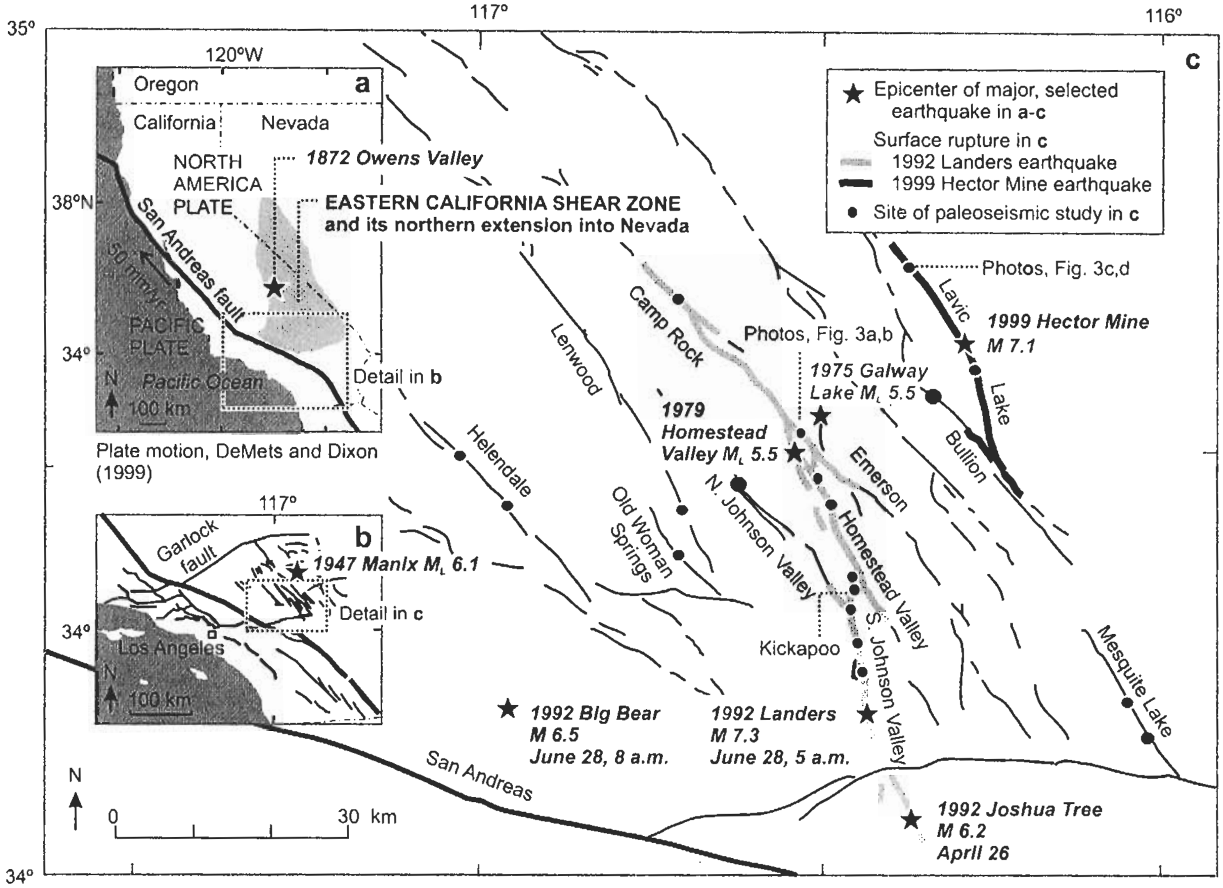
Earthquakes of the 1990s thus define an uncommon episode of seismic activity in the eastern California shear zone. An earlier seismic episode, farther north in east-central California and western Nevada, occurred between 1872 and 1954 in a shear zone 500 km long (Wallace, 1978, 1984). These historical examples raise the question, do M 6–7 earthquakes in the eastern California shear zone typically come in clusters?

Prehistoric Earthquakes

Paleoseismic studies of the eastern California shear zone began a few weeks after the 1992 Landers earthquake and eventually involved more than 17 trenches across eleven faults (Fig. 2c). The studies focused on playas where the vertical component of slip produced stratigraphic offsets in fine-grained, stratified deposits of Holocene age (Fig. 3). Evidence for surface rupture includes faults and fissures that terminate at buried land surfaces, folding and warping of beds, deposits that resulted from ponding against fault scarps, and scarp-derived colluvium. Laminated lacustrine deposits allow detection of vertical separation as small as several centimeters. Prehistoric faulting also produced noticeable offsets in alluvium, colluvium, and buried soils. Detrital charcoal and peat beds have yielded radiocarbon ages that limit inferred times of the prehistoric ruptures and related earthquakes.

The paleoseismic studies confirm that the Landers and Hector Mine earthquakes were rare events (Fig. 4). Few of the faults trenched show evidence for more than two surface ruptures between 10,000 years ago and A.D. 1992. No prehistoric rupture of Holocene age has been found where the Lavic Lake fault ruptured in 1999, with the possible exception of minor slip after than A.D. 260 (Rymer *et al.*, 2002b).

The inferred earthquakes can be grouped into three Holocene episodes on the basis of overlapping radiocarbon ages (episodes A, B, and C in Fig. 4). The episodes are loosely defined because uncertainties in dating the prehistoric ruptures commonly span centuries (Fig. 4a). Episode A, about 8000–9000 cal yr B.P., includes the most recent



Faults from Dokka and Travis (1990) and Jennings and Saucedo (1994)

Fig. 2. Faults of the eastern California shear zone and vicinity. Faults in (c) have documented or presumed evidence for surface rupture within the past 10,000 years.

pre-1992 events on the Kickapoo and northern Emerson faults. These events produced fault scarps similar in height to the 1992 scarp. Episode A may also include the penultimate prehistoric surface rupture on the Helendale and Mesquite Lake faults, as well as ruptures on the Lenwood, Camp Rock, and southern Johnson Valley faults. Episode B, about 5000–6000 cal yr B.P., followed several thousand years of apparent quiescence. Surface ruptures occurred on the Lenwood, Johnson Valley, Bullion, and Mesquite Lake faults. The shear zone became active again in the past 1000 years, during episode C. This latest series of earthquakes, which continued into the 1990s, produced surface rupture on many faults in the shear zone. Though represented by a single rupture at most sites, episode C includes both a prehistoric rupture and the 1992 Landers rupture on the Camp Rock fault (Fig. 4, site 4).

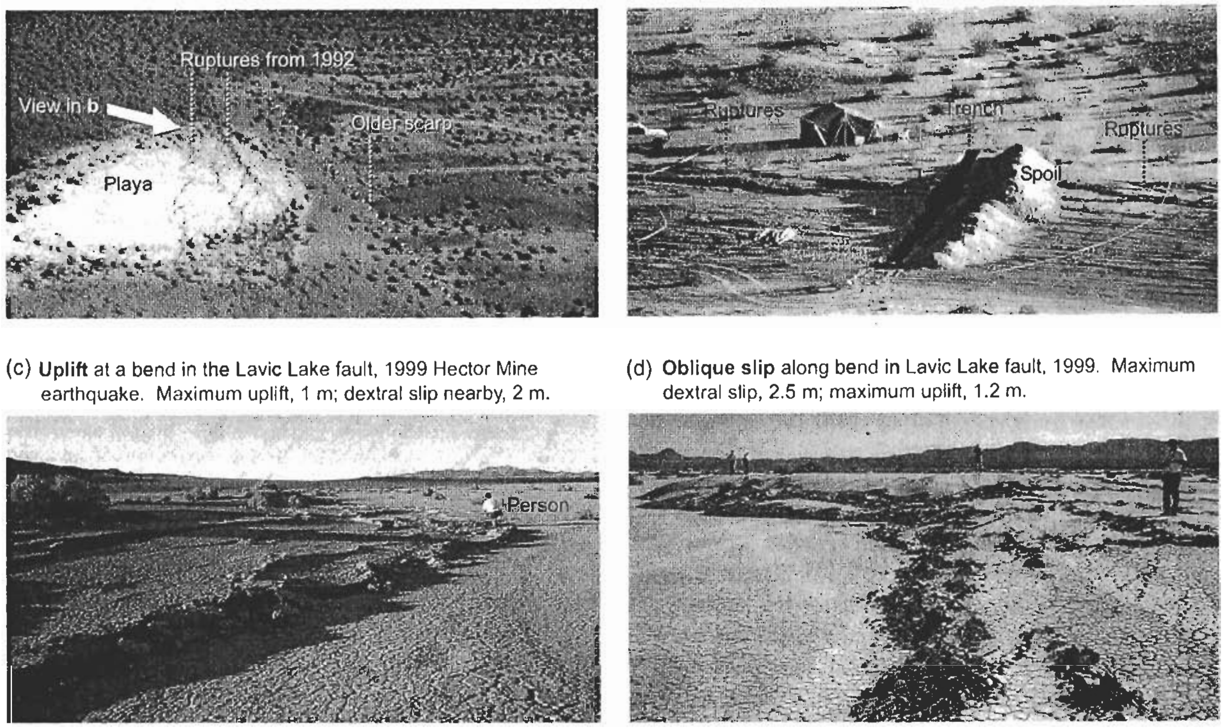
Implications and Challenges

The long intervals between episodes imply that the earthquakes of the 1990s represent an unusual peak in seismic activity in the eastern California shear zone. However, episode C differs too much from A and B for any of the episodes to define the likely size and pattern of future surface ruptures (Rockwell *et al.*, 2000). Viewed as part of episode C (Fig. 4), the shear zone's 20th-century earthquakes imply either that additional earthquakes are likely, or that episode C is drawing to a close. The Working Group of California Earthquake Probabilities (1995), without much paleoseismic information about the eastern California shear zone, presumed that in coming decades, the zone would continue producing earthquakes like those since 1970 (Fig. 2c).

Geophysicists have proposed various triggers for swift series of earthquakes in the eastern California shear zone (Freed & Lin, 2001; Harris & Simpson, 2002; Hudnut *et al.*, 2002; Pollitz & Sacks, 2002; Zeng, 2001). The zone's history of episodic Holocene earthquakes suggests that a realistic trigger will permit thousands of years to elapse between earthquake series.

New Madrid Seismic Zone

Paleoseismology can clarify fault location and earthquake recurrence far from plate boundaries, in continental regions where tectonic activity has less geomorphic or seismological



(a) **Ruptures** on the Emerson fault from the 1992 Landers earthquake. A degraded older scarp runs parallel to them.

(b) **Trench** across 1992 ruptures on playa in **a**. Maximum dextral slip, 2.3 m; maximum uplift, 0.8 m. Site 6 in Figure 4a.

(c) **Uplift** at a bend in the Lavic Lake fault, 1999 Hector Mine earthquake. Maximum uplift, 1 m; dextral slip nearby, 2 m.

(d) **Oblique slip** along bend in Lavic Lake fault, 1999. Maximum dextral slip, 2.5 m; maximum uplift, 1.2 m.

Fig. 3. Surface ruptures of the 1992 Landers and 1999 Hector Mine earthquakes (locations, Fig. 2c).

expression than in the eastern California shear zone. One such region is the lower Mississippi River valley (Schweig *et al.*, 2002). This valley contains the New Madrid seismic zone (Fig. 5a), which during the winter of 1811–1812 produced some of the most widely felt earthquakes in the written history of the United States. Studies of prehistoric earthquakes in the New Madrid region have shown that the 1811–1812 earthquakes were not freak, one-time events.

The three largest shocks of the 1811–1812 sequence, of *M* 7.5–8.0 (Atkinson *et al.*, 2000; Hough *et al.*, 2000; Johnston, 1996), rank among Earth's largest intraplate quakes (Johnston & Kanter, 1990). They destroyed settlements along the Mississippi River, damaged buildings as far away as Cincinnati and St. Louis (Fig. 1), and were felt at distances as great as 1,800 km (Nuttli, 1973). They induced severe liquefaction and related ground failure throughout the New Madrid region (Fig. 5b; Fuller, 1912; Obermeier, 1989; Saucier, 1977) and locally as far as 250 km from inferred epicenters (e.g. Johnston & Schweig, 1996; Street & Nuttli, 1984).

Although few faults have geomorphic expression in the New Madrid region, numerous small modern earthquakes illuminate several interseting faults (Fig. 5a; Chiu *et al.*, 1992; Pujol *et al.*, 1997). Most of these earthquakes occur beneath Late Wisconsin and Holocene deposits of the Mississippi River and its tributaries. Many of the fluvial deposits liquefied during the A.D. 1811–1812 earthquakes, venting water and sand that formed sand blow deposits across about 10,000 km^2 (Figs 5 and 6). Prehistoric sand blows in this area provide the main evidence for two earlier episodes of New Madrid earthquakes during the past 1200 years.

Paleoseismic Evidence

According to oral traditions of Native Americans in the Mississippi River valley, a great earthquake devastated the region centuries before 1811 (Lyell, 1849). Geologic evidence for such an earthquake was first reported by Fuller (1912), who noted liquefaction-related ground failures and a history of uplift and erosion predating 1811. He inferred that the region had experienced "early shocks of an intensity equal to if not greater than that of the last."

Detailed study of pre-1811 earthquakes in the region began at the Reelfoot scarp (Fig. 5a). This landform coincides with a northwest-trending zone of microseismicity and may be a monocline above a blind thrust fault (Russ, 1982). As inferred from deformed sediments exposed in trenches across the scarp, prehistoric folding and earthquake-induced liquefaction occurred at least twice in the past 2000 years (Russ, 1979), probably in A.D. 780–1000 and A.D. 1260–1650 (Kelson *et al.*, 1992, 1996).

Archeological studies contributed to the recognition that many sand blows in the New Madrid region predate 1811–1812. For example, 30 km northeast of Reelfoot scarp at Towosahgy State Park (Fig. 5a), sand-filled fissures and two related sand blow deposits underlie a Native American mound

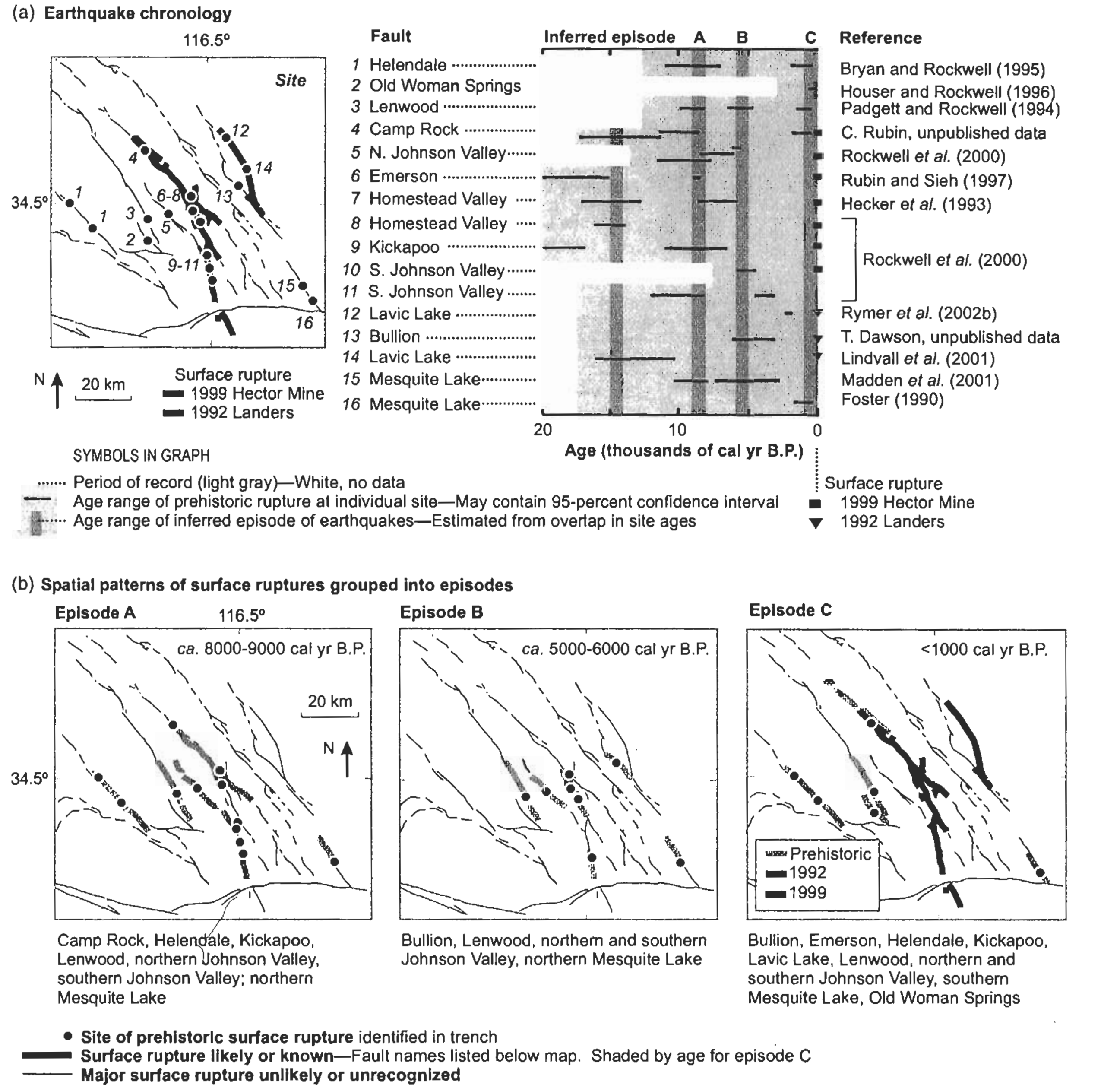


Fig. 4. Chronology and spatial patterns of earthquakes that produced surface ruptures in the eastern California shear zone.

(Saucier, 1991). The liquefaction features were attributed to two large earthquakes between about A.D. 400 and A.D. 1000.

Since the 1980s, hundreds of liquefaction features have been examined in the New Madrid region. These include more than 50 sand blows that have been studied in detail, many at archeological sites (Broughton *et al.*, 2001; Li *et al.*, 1998; Tuttle *et al.*, 1996, 1999, 2002; Vaughn, 1994; Wesnousky & Leffler, 1992). The combination of regional reconnaissance and detailed investigations has advanced the dating of the region's prehistoric earthquakes and the assessment of its earthquake potential (Tuttle *et al.*, 2002).

The challenge has not been finding sand blows, which abound in the region (Figs 5 and 6), but rather finding sand blows that can be dated well. In this agricultural region, plowing and grading have disturbed the upper 15–20 cm of soils at most sites. Soils developed on 1811–1812 sand blows are commonly thin enough to have been completely reworked by plowing. Soils developed on prehistoric sand blows, however, can be thick enough to retain cultural materials below the plow zone (Fig. 7). The New Madrid region contains thousands of Native American sites occupied at various times during the past 2000 years (Morse & Morse, 1983). Remains of these sites – including fire pits, storage pits, post molds, and trench fills – have been found on or beneath sand blows. The cultural horizons contain wood, charcoal, and plant remains that yield minimum and

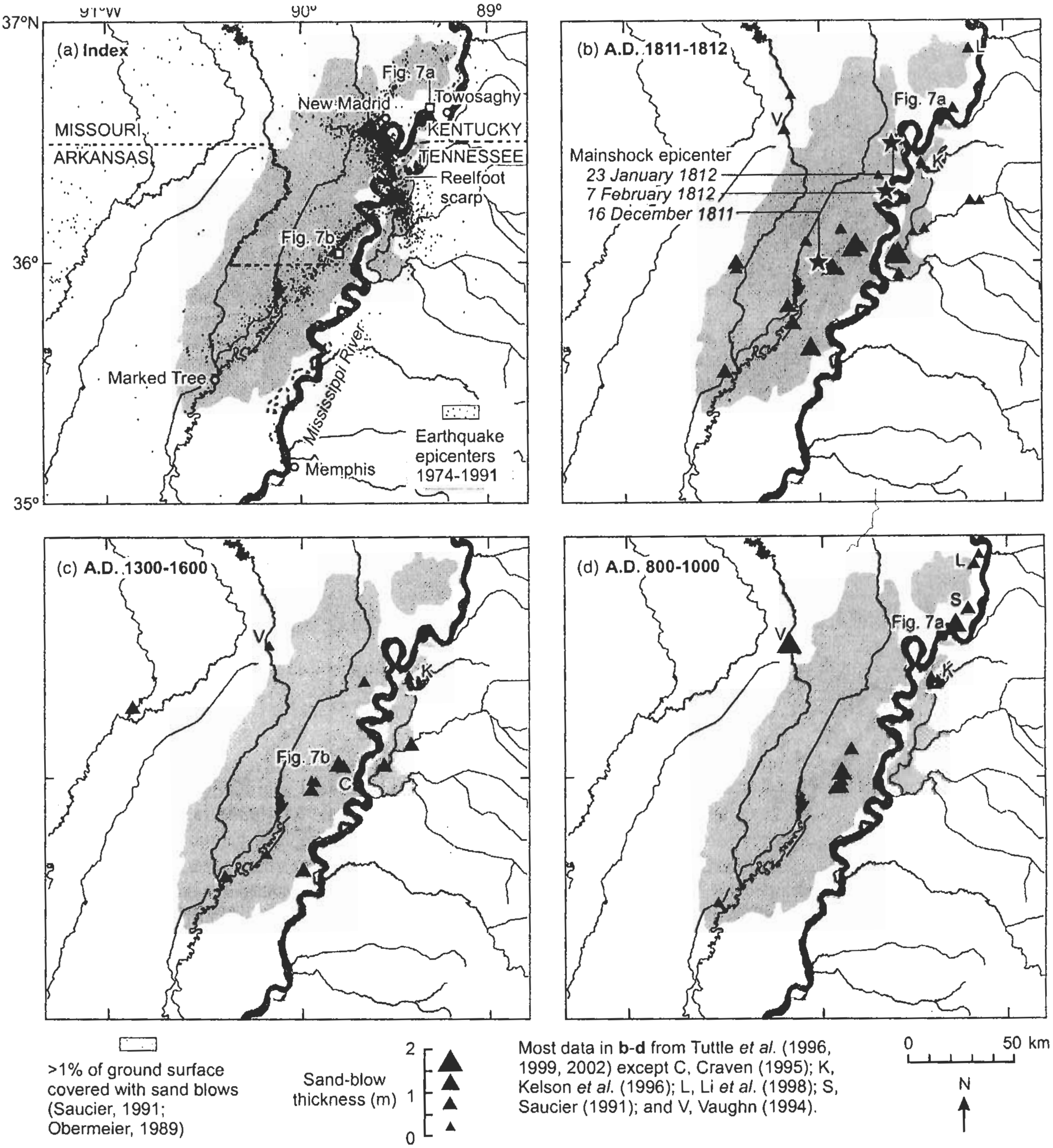


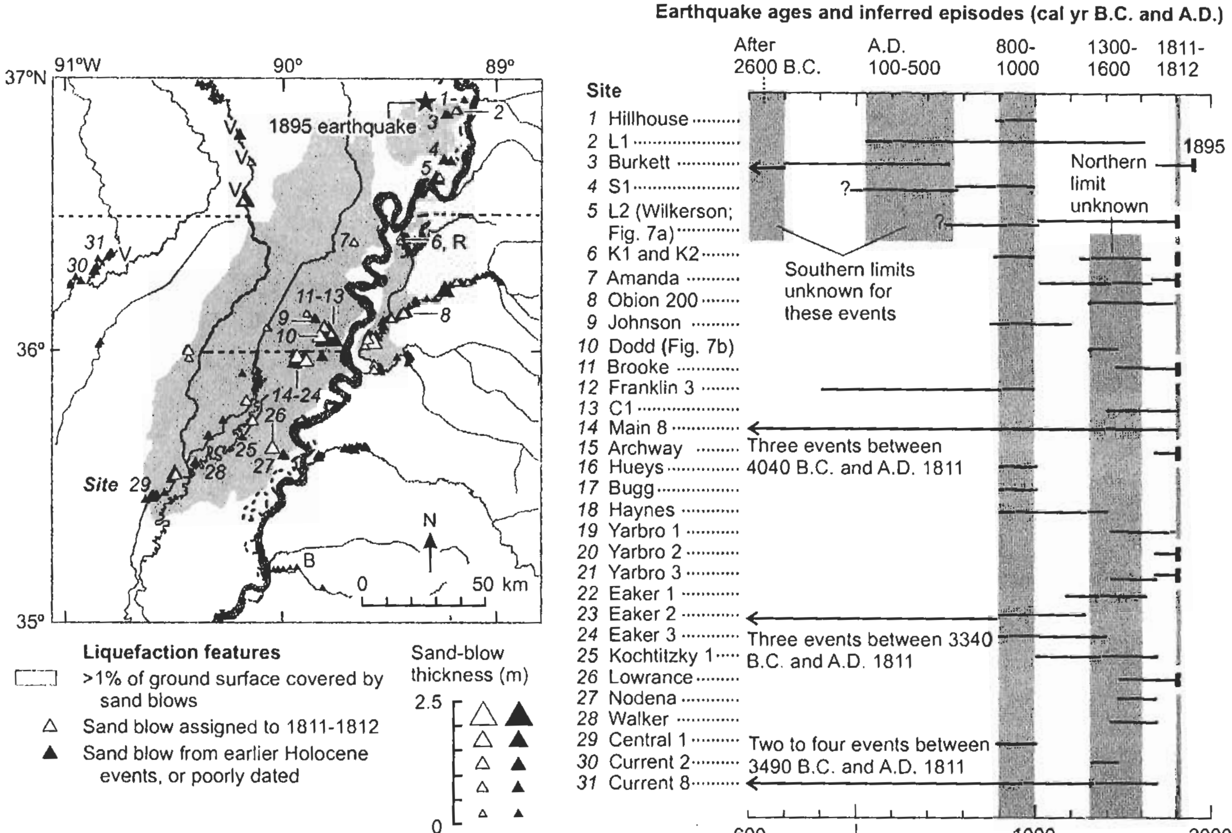
Fig. 5. Index map (a) and distribution and sizes of sand blows (b–d) at the New Madrid seismic zone.

maximum ages for earthquake-induced liquefaction features (Tuttle, 2001).

In addition to archaeological sites, the New Madrid region contains natural and artificial drainages that expose cross sections through historic and prehistoric sand blows. Reconnaissance of river and ditch banks has yielded some of the information about the size and spatial distribution of liquefaction features summarized in Figs 5 and 6.

Prehistoric Earthquakes

In the New Madrid region, prehistoric liquefaction features commonly date to A.D. 800–1000 or 1300–1600. In size and distribution, features in these age ranges resemble the sand blows from the earthquakes of 1811–1812 (Fig. 5b–d). Additional liquefaction features date from at least two earlier time intervals since 3000 B.C., but too few sites have been



Most features on map and ages in graph from Tuttle *et al.* (1996, 1999, 2002). Exceptions: sand-blow cover area (Saucier, 1977; Obermeier, 1989); localities B (Broughton *et al.*, 2001), C1 (Craven, 1995), K1 and K2 (Kelson *et al.*, 1996), L1 and L2 (Li *et al.*, 1998), R (Russ, 1982), S1 (Saucier, 1991), V (Vaughn, 1994).

Fig. 6. Chronology of earthquakes at the New Madrid seismic zone.

studied to estimate the locations and sizes of the earthquakes that produced them (Fig. 6).

The episodes of A.D. 800–1000 and 1300–1600 each contained earthquakes in swift series. The serial earthquakes of 1811–1812 produced multiple, upward-fining depositional units, each of which probably represents an individual earthquake (Saucier, 1989). Most prehistoric sand blows also contain such multiple units, both from the years 800–1000 and from 1300–1600 (Tuttle *et al.*, 2002).

Implications and Challenges

The New Madrid events of A.D. 800–1000, 1300–1600, and 1811–1812 together indicate recurrence intervals as short as 200 years or as long as 800 years, with a two-interval average of about 500 years (Fig. 6). This average has been incorporated into the latest national earthquake hazard maps as the recurrence interval for New Madrid earthquakes like those in 1811–1812 (Fig. 1; Frankel *et al.*, 2002, p. 3). In previous mapping of the region's earthquake hazards, the interval used was 1000 years.

Improved estimates of earthquake recurrence may be obtained by further studying the liquefaction features older than A.D. 800. These efforts may also help address other issues at the New Madrid seismic zone, such as long-term fault behavior (Tuttle *et al.*, 2002), causes for large earthquakes in a mid-plate region (Grollimund & Zoback, 2001; Kenner & Segall, 2000; Pollitz *et al.*, 2001; Stuart, 2001), and slowness of present-day deformation (Newman *et al.*, 1999).

Cascadia Subduction Zone

In our eastern California and New Madrid examples, geologic records of prehistoric earthquakes resemble those produced by historical earthquakes known from instrumental records and eyewitness accounts. In some other places, paleoseismic evidence has no local analog in written history. Paleoseismology provides the only detailed knowledge of surface ruptures on Utah's Wasatch fault (Gori & Hayes, 1992, 2000; McCalpin & Nishenko, 1996), as was anticipated by Gilbert (1883). Prehistoric liquefaction features record

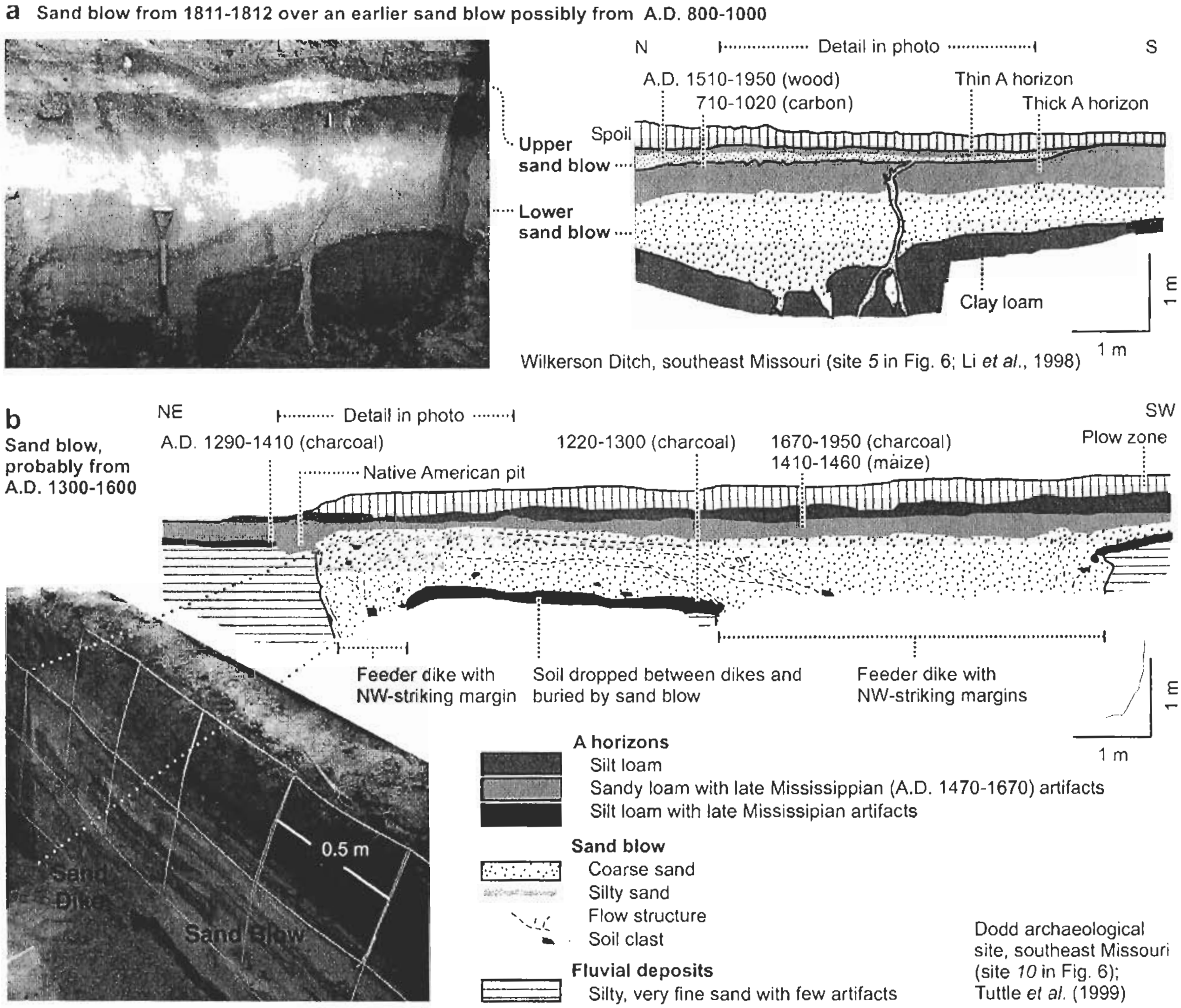


Fig. 7. Examples of sand blows at the New Madrid seismic zone.

the most recent large earthquake at the Wabash seismic zone of Illinois and Indiana, north of the New Madrid seismic zone (Obermeier *et al.*, 1991). Coastal geology shows that Washington's Seattle fault produced its most recent large earthquake about A.D. 900 (Bucknam *et al.*, 1992; fault locations in Fig. 1a).

Likewise at the Cascadia subduction zone (Fig. 8), all earthquakes of M 8–9 predate the region's written history. These great earthquakes ruptured the boundary between the subducting Juan de Fuca plate and the overriding North America plate. Although few earthquakes attain M 9 – the 20th century had no more than three or four examples (Kanamori, 1977; Ruff, 1989, p. 273) – the Cascadia earthquake in A.D. 1700 probably did. In the 1990s, this and other great earthquakes inferred from paleoseismology elevated the hazard mapped along the Pacific coast from northern California to southern British Columbia (Petersen *et al.*, 2002; Fig. 1b).

Coseismic Subsidence

Cascadia's great-earthquake hazard escaped detection until the last two decades of the 20th century. Geophysicists deduced that Cascadia can produce great earthquakes (Heaton & Kanamori, 1984; Savage *et al.*, 1981). Geologists then began finding evidence that great Cascadia earthquakes have happened (reviewed by Clague, 1997). Much of the geologic evidence consists of the buried soils of former forests and marshes that subsided into estuaries during earthquakes (Fig. 9).

Such subsidence can lower entire regions. During great thrust earthquakes at subduction zones, the upper plate lurches seaward above the rupture. Where this motion elastically stretches and thins the upper plate, the land surface drops. The grandest modern examples of coseismic subsidence come from earthquakes in Chile (1960, M 9.5) and Alaska (1964, M 9.2). Each of these earthquakes produced a largely coastal

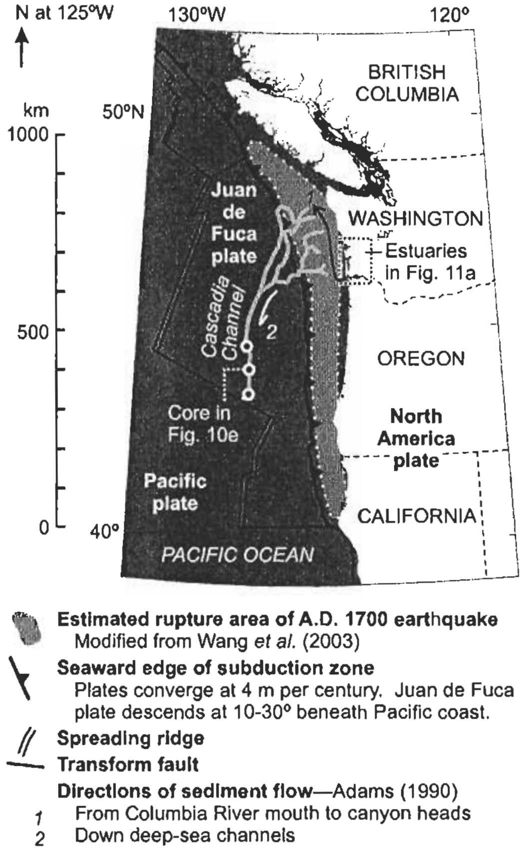


Fig. 8. Cascadia subduction zone.

downwarp more than 800 km long, many tens of kilometers wide, and as much as 2.3 m deep (Plafker, 1972). The Alaskan earthquake quickly entered the stratigraphic record at the head of a macrotidal estuary, where post-earthquake tides killed subsided forests and meadows while burying their soils with silt (Atwater *et al.*, 2001; Ovenshine *et al.*, 1976).

Estuarine stratigraphic records of coseismic subsidence can commonly be distinguished from those of other kinds of coastal change, such as gradual rise in sea level, sudden breaching of sand spits, and anomalous deposition by storms or floods (Nelson *et al.*, 1996b). To be considered evidence for coseismic subsidence, the top of a buried soil must mark a change from a relatively high environment (such as a forest or the upper part of a tidal marsh) to a relatively low one (such as an unvegetated tidal flat). Growth-position fossils of vascular plants can record such a drop (Atwater & Hemphill-Haley, 1997, p. 44), as can assemblages of diatoms, foraminifers, and pollen (Guilbault *et al.*, 1996; Hemphill-Haley, 1995; Hughes *et al.*, 2002; Kelsey *et al.*, 2002; Nelson *et al.*, 1996a; Shennan *et al.*, 1996). The change, moreover, must have happened suddenly. Sediment texture and fossils differ across a sharp contact, wide outer rings show trees healthy until their last year or two, and growth-position stems and leaves of herbaceous plants imply rapid burial (Atwater & Yamaguchi, 1991; Jacoby *et al.*, 1995; Fig. 9c).

If an earthquake produces a tsunami or liquefaction, the earthquake may be further marked by sand that mantles a buried soil in a stratigraphic section otherwise free of sand. At many Cascadia estuaries, a sand sheet suggests that burial of a freshly subsided soil began with a tsunami (e.g. Clague *et al.*, 2000). Marine diatoms within and landward of such sand sheets strengthen the case for tsunami inundation (Fig. 9d; Hemphill-Haley, 1996). In this same stratigraphic position at a few Cascadia estuaries, sand lenses fed by sand dikes show that soil burial began with venting of water and sand in response to earthquake-induced liquefaction that happened about the time the soil subsided (Kelsey *et al.*, 2002, p. 309; Obermeier, 1996, p. 43).

Prehistoric Earthquakes

In the late Holocene, coseismic subsidence in coastal Washington and Oregon has recurred at intervals mostly 300–800 years long (Kelsey *et al.*, 2002). Because of uncertainties in correlations based solely on numerical ages, little is known about the coastwise extent of individual subsidence events before A.D. 1700. However, at least three estuaries of southern Washington probably share a 3000-year history of repeated coseismic subsidence at irregular intervals (Figs 10 and 11).

This earthquake history is based on a widely correlative stack of buried soils exposed in low-tide outcrops (Fig. 10a and b). The stacked soils consistently differ from one another in organic-matter preservation and fossil-forest extent, in ways that imply differing lengths of time between earthquakes (Fig. 10c and d). The better preserved a buried organic horizon and its herbaceous fossils, the shorter the time when this buried organic matter remained subject to degradation in the profile of the next soil. The farther downstream a forested site, the longer the interseismic time when gradual uplift and sedimentation allowed forests to spread seaward along estuarine salinity gradients (Fig. 10d; Atwater & Hemphill-Haley, 1997, pp. 95–99; Benson *et al.*, 2001).

These relative measures of interseismic time agree with numerical estimates from radiocarbon dating (Fig. 10d). Radiocarbon ages from estuaries of southwest Washington anchor an earthquake chronology of uncommon precision – not only because most of the ages have reported errors of just 10–20 ^{14}C yr, but also because many ages were measured on the rings of earthquake-killed trees (Fig. 11; Appendix 1). Such tree-ring samples allow exact correction for the age of dated material relative to the time of an inferred earthquake (Nelson *et al.*, 1995). Other materials set only limiting ages for the earthquakes: maximum ages from detritus in pre-earthquake soils, minimum ages from rhizomes (below-ground stems) of plants that colonized post-earthquake tidal flats.

The individual ages are grouped in Fig. 11 by stratigraphic position defined by soil preservation and paleoecology – by field correlation of seven buried soils named J, L, N, S, U, W and Y (Fig. 10a–c; Atwater & Hemphill-Haley, 1997). The individual ages, many previously unpublished, yield combined age ranges for the field-correlated events (gray columns, Figs 10d and 11c and d). Most of these event age

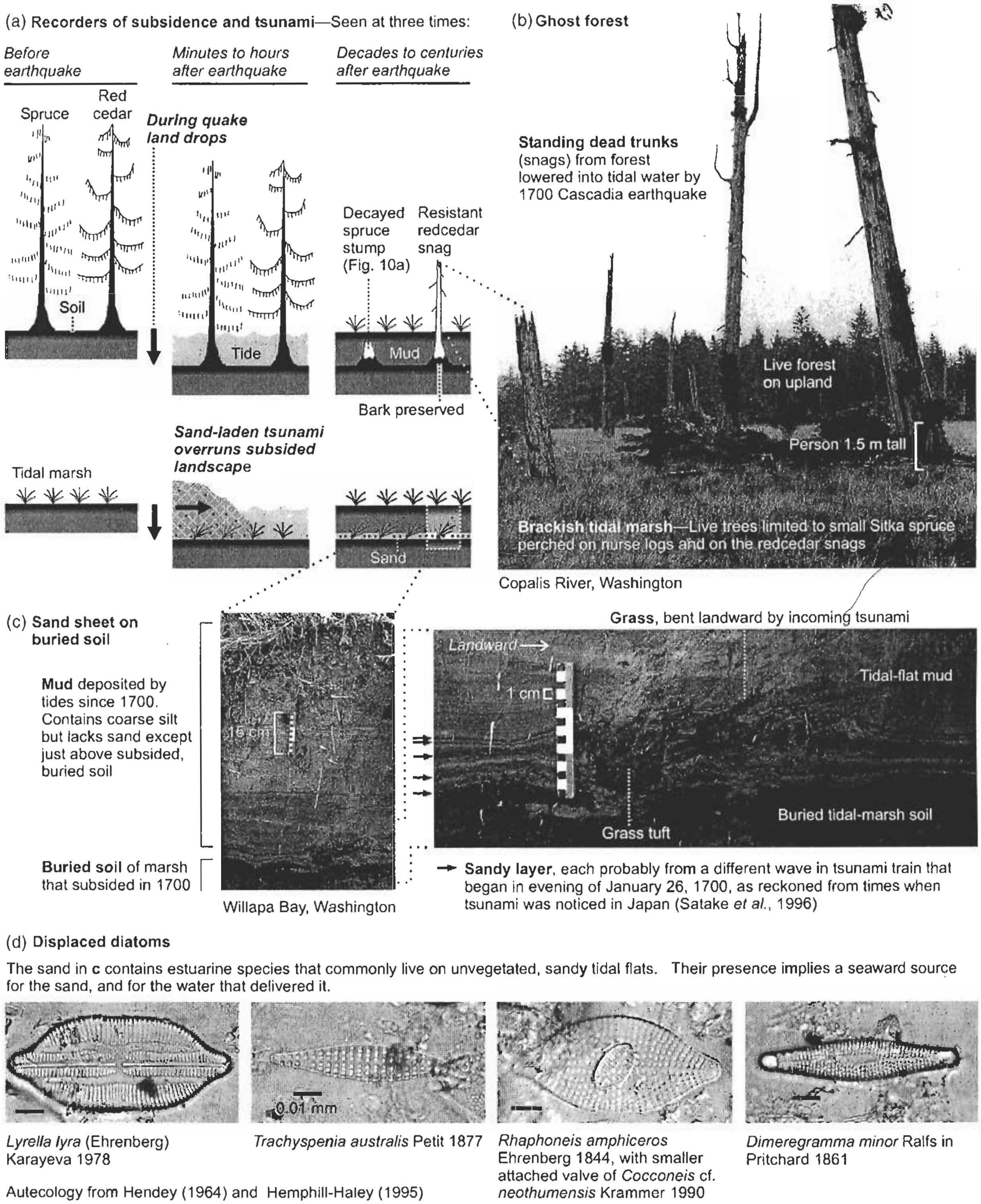


Fig. 9. Evidence for coseismic subsidence and tsunami occurrence at the Cascadia subduction zone (locations, Fig. 11a).

ranges are governed by times of tree death (black bars in Fig. 11d); some are limited also by ages from pre-event detritus or from post-event rhizomes (arrows in Fig. 11d).

To derive the age range for each event, its field-correlated individual ages were combined under the key assumption that they all refer to the same event – either an earthquake or a swift series of earthquakes. The combining is based on Bayesian statistics, which applied to an ordered sequence of ages can yield event ages with narrowed confidence limits (Biasi & Weldon, 1994; Ramsey, 2000). The ranges include generous

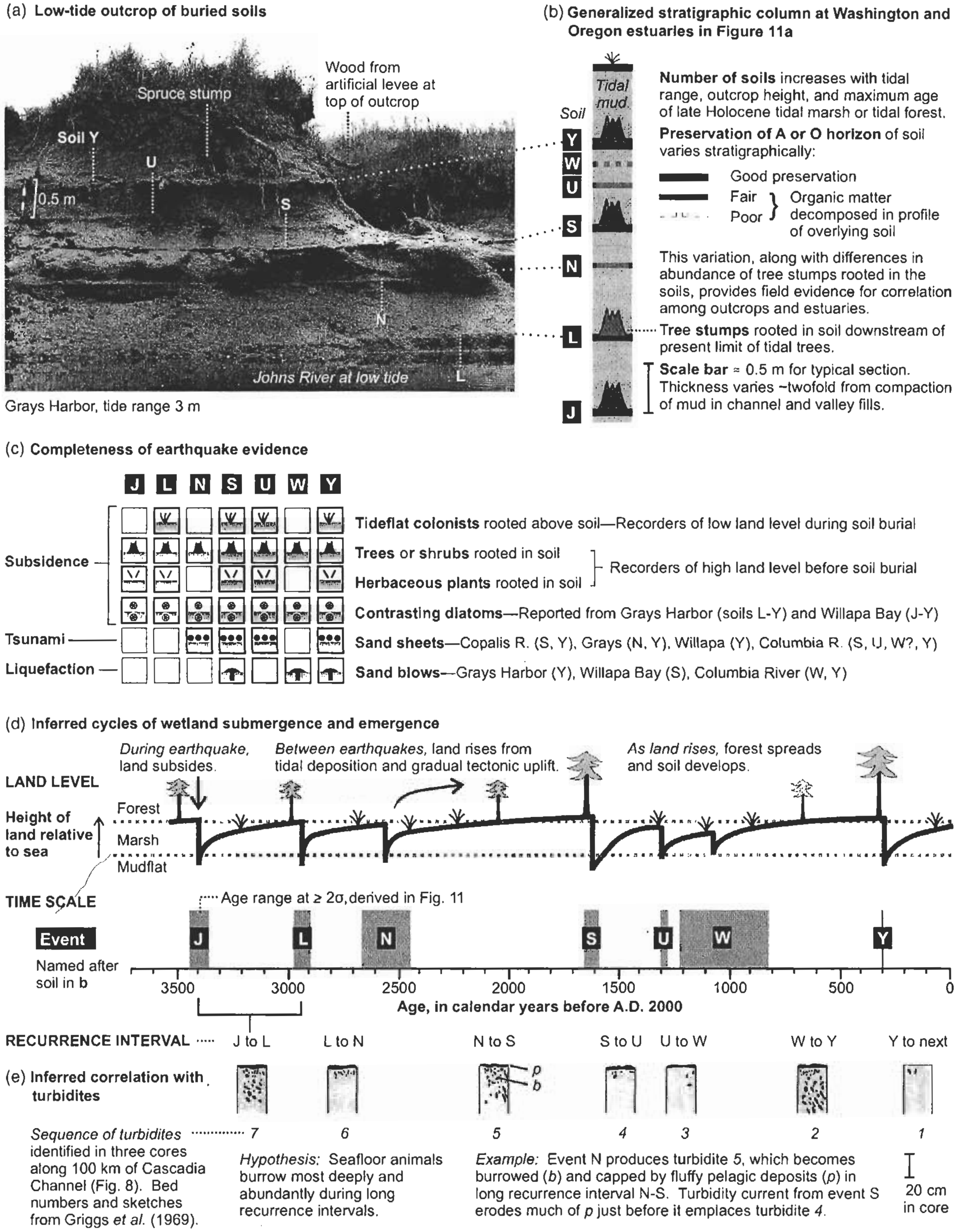


Fig. 10. Evidence for recurrent earthquakes in southwest Washington and northwest Oregon.

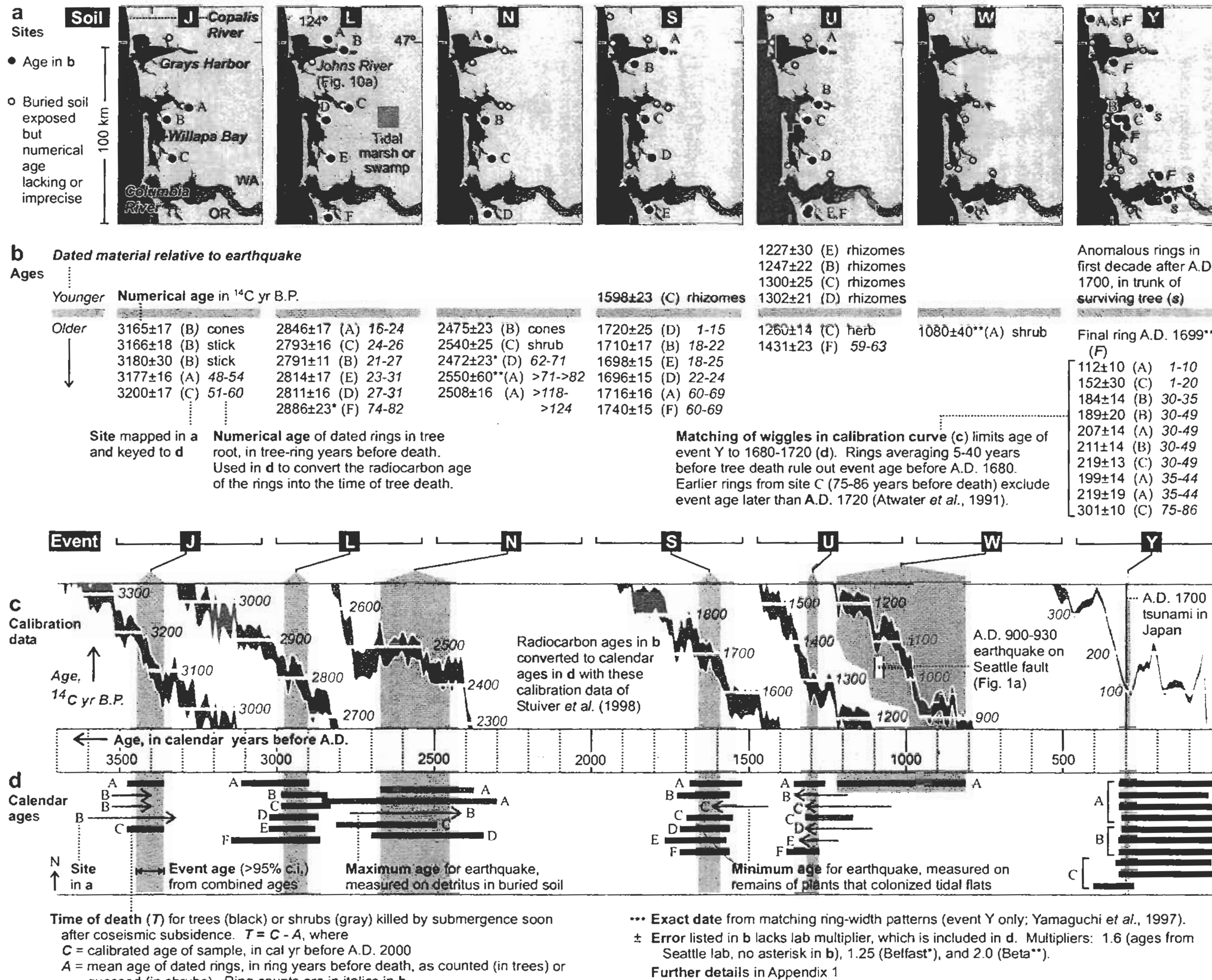


Fig. 11. Chronology of prehistoric great earthquakes in southwest Washington and northwest Oregon.

estimates of uncertainty in radiocarbon analysis (augmented by error multipliers listed at bottom right in Fig. 11). These procedures, in the calibration program Oxcal, yield event age ranges that probably include 95% confidence intervals.

The age ranges for the seven events in Figs 10 and 11 define six recurrence intervals that vary in length from a few centuries (intervals S–U and U–W) to one millennium (N–S). Though the intervals average 500–540 years, only one of the recurrence intervals (J–L) is close to this average. During the longest intervals, which exceeded this average by several centuries (N–S and W–Y), tidal forests advanced seaward as the shallowest buried soils decomposed (Figs 9b and 10a–d).

This history of aperiodic earthquakes probably correlates with turbidity-current deposits off the Oregon coast in Cascadia deep-sea channel (Atwater & Hemphill-Haley, 1997, pp. 102–103). The deposits, derived from Columbia River sediment on the continental shelf and slope, apparently originated at submarine canyon heads above the fault ruptures that caused coseismic subsidence in coastal Washington (Fig. 8). The turbidity currents repeated at intervals that averaged close to 600 years in the past 8000 years. Because eroded pelagic deposits between turbidites are similar in thickness (*p* in Fig. 10e), the repetition was first interpreted as periodic (Adams, 1990; Griggs *et al.*, 1969). However, the successive turbidites vary in their depth and abundance of animal burrows (*b* in Fig. 10e). This variability links the turbidites with the aperiodic earthquakes inferred from estuarine stratigraphy in coastal Washington and adjacent Oregon.

The most recent great Cascadia earthquake was dated by radiocarbon methods to the decades around A.D. 1700 (Nelson *et al.*, 1995; event Y in Figs 10 and 11). This era precedes, by almost a century, the Spanish and English exploration that marks the beginning of written history at Cascadia (Hayes, 1999). Along nearly 1000 km of Japan's Pacific coast, however, government officials and merchants noted a puzzling tsunami in A.D. 1700 that lacked a nearby earthquake. The time of this orphan tsunami suggests that a great Cascadia earthquake occurred on the evening of January 26, 1700. The tsunami's height of several meters further suggests that this earthquake attained *M* 9 (Satake *et al.*, 1996).

Tree-ring studies in southwest Washington and adjacent Oregon support these inferences from Japan. Death and stress in subsided trees date to the first few years after the 1699 growing season (Jacoby *et al.*, 1997; Yamaguchi *et al.*, 1997). Except for a few dozen survivors of the earthquake (Jacoby *et al.*, 1997), all trees in the region's modern tidal forests postdate 1700 (Benson *et al.*, 2001).

Implications and Challenges

The latest version of the national earthquake hazard map gives equal weight to two patterns of great-earthquake recurrence at Cascadia (Frankel *et al.*, 2002, p. 11). In one, *M* 9.0 earthquakes rupture the full 1100-km length of the subduction zone every 500 years. In the other, the subduction zone breaks in segments 250 km long that behave independently of one another. This style or rupture produces one earthquake of *M* 8.3 every 110 years somewhere along the zone (Frankel *et al.*, 1996). The shorter recurrence intervals for the independent *M* 8.3 earthquakes yield higher probabilistic ground motions than does the 500-year interval for *M* 9.0 events.

Such hazard estimates are likely to improve as great-earthquake history becomes better documented along the Cascadia subduction zone. Does the zone contain segments that sometimes rupture independently, decades or centuries out of phase with other segments? Along a single part of the subduction zone, do long recurrence intervals commonly precede short ones, much as long interval N–S preceded short intervals S–U and U–W (Fig. 10d)? Does long interval W–Y thereby justify increasing the probabilistic hazard on the national map (Fig. 1b)? Paleoseismic studies at Cascadia are just beginning to address such questions.

Also needed at Cascadia – and elsewhere – are estimates of the smallest earthquake and shortest recurrence interval that paleoseismic records resolve. Such estimates are likely to affect recurrence intervals and the probabilistic hazard inferred from them.

Summary

Paleoseismology has provided engineers and public officials with long histories of recurrent earthquakes (or histories of recurrent series of earthquakes). Typical intervals between the earthquakes (or series) span hundreds of years in our New Madrid and Cascadia examples and thousands of years in our eastern California example. In addition to enabling such estimates of recurrence intervals, paleoseismology can provide evidence for regional clustering of earthquakes in seismic zones (eastern California, New Madrid) and for aperiodic rupture along the same part of a fault (Cascadia). Such findings have made paleoseismology an essential part of earthquake-hazard assessment in the United States.

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Appendix 1 *Radiocarbon ages in Figure 11.*

Event	Estuary	Site (Fig. 11a)	Lab no.	Sample age (radiocarbon yr B.P.) One-standard deviation counting error excludes [E] or includes [I] the multiplier at lower right in Fig. 11			Material dated (Fig. 11b: ring number 1 adjoins bark; spruce, *Picea sitkensis*; redcedar, *Thuja plicata*)	Sample age with respect to time of earthquake (–, not applicable)		Earthquake age not shaved in Oxcal (individual ages in Fig. 11d; calibrated with data of Stuiver *et al.*, 1998; curves in Fig. 11c)[a]		Earthquake age shaved in Oxcal (event age ranges plotted in Figs 10d and 11d; italicized, individual ages not plotted)[a]		Location (AHH, Atwater & Hemphill-Haley, 1997; A96, Atwater, 1996; A92, Atwater, 1992, Table 1)
				Age	E	I		Numerical mean or range (ring years before plant death)	Qualitative (loosely limiting ages)					
Y[b]												1698	1715	Combined age for event
	Copalis River	A	QL-4408	112	11	17.6	Spruce root in soil, rings 1–10	5.5	–	1680	1960	*1698*	*1715*	47°07.08′N, 124°10.01′W
	Willapa Bay	C	QL-4405	152	30	48	Spruce root in soil, rings 1–20	10.5	–	1670	1970	*1697*	*1715*	Niawiakum River near Pool locality (AHH, p. 12)
	Willapa Bay	B	QL-4401	184	14	22.4	Spruce root in soil, rings 30–35	32.5	–	1690	1990	*1697*	*1716*	Bay Center beach (A96 p. 83)
	Willapa Bay	B	QL-4403	189	20	32	Spruce root in soil, rings 30–49	39.5	–	1680	2000	*1697*	*1716*	Bay Center beach (A96 p. 83)
	Copalis River	A	QL-4400	207	14	22.4	Spruce root in soil, rings 30–49	39.5	–	1680	1990	*1697*	*1716*	47°07.08′N, 124°09.96′W
	Willapa Bay	B	QL-4402	211	14	22.4	Spruce root in soil, rings 30–49	39.5	–	1680	1990	*1697*	*1716*	Bay Center beach (A96 p. 83)
	Willapa Bay	C	QL-4404	219	13	20.8	Spruce root in soil, rings 30–49	39.5	–	1680	1990	*1697*	*1716*	Niawiakum River near Pool locality (AHH, p. 12); 46°36.68′N, 123°53.61′W
	Copalis River	A	QL-4410	199	15	24	Spruce root in soil, rings 35–44	39.5	–	1690	1990	*1697*	*1716*	47°07.11′N, 124°09.96′W
	Copalis River	C	QL-4409	219	19	30.4	Spruce root in soil, rings 35–44	39.5	–	1680	1990	*1697*	*1716*	47°07.08′N, 124°10.01′W
	Willapa Bay	C	QL-4406	301	10	16	Spruce root in soil, rings 75–86	80.5	–	1600	1730	*1697*	*1716*	Niawiakum River near Pool locality (AHH, p. 12); 46°36.69′N, 123°53.74′W
W												780	1190	Based on single age
	Columbia River	A	Beta-121421	1080	40	80	Shrub root in soil	10–30	–	780	1190	*780*	*1190*	Lewis and Clark River, 46°07.90′N, 123°52.52′W
U												686	721	Combined age for event
	Columbia River	E	QL-4924	1227	30	48	*Triglochin* rhizomes above soil	–	Younger than earthquake	Before 680	Before 950	*Before 690*	*Before 940*	Lewis and Clark River, 46°09.37′N, 123°51.28′W
	Willapa Bay	B	QL-4822	1247	22	35.2	*Triglochin* rhizomes above soil	–	Younger than earthquake	Before 680	Before 890	*Before 700*	*Before 890*	Willapa River, Airport locality of AHH (p. 12, 78)
	Willapa Bay	D	QL-4827	1300	25	40	*Triglochin* rhizomes above soil	–	Younger than earthquake	Before 650	Before 810	*Before 690*	*Before 860*	Naselle River, 46°24.25′N, 123°50.42′W
	Willapa Bay	C	QL-4798	1302	21	33.6	*Triglochin* rhizomes above soil	–	Younger than earthquake	Before 660	Before 780	*Before 690*	*Before 810*	Niawiakum River, Oyster locality (AHH, p. 12, 44)
	Willapa Bay	C	QL-4795	1260	14	22.4	*Salicornia* stems in growth position within and <5 cm above soil	5–15	–	680	830	*686*	*721*	Niawiakum River, Oyster locality (AHH, p. 12, 44)
	Columbia River	F	UB-4499	1431	28	28	Spruce root in soil, rings 59–63	61	–	620	725	*685*	*721*	Lewis and Clark River, 46°07.05′N, 123°52.35′W
	Grays Harbor	A	QL-4913	1449	14	22.4	Spruce root in soil, rings 85–89	87	–	645	745	*687*	*721*	Chehalis River, 46°58.70′N, 123°46.87′W
S												340	410	Combined age for event
	Willapa Bay	C	QL-4797	1598	23	36.8	*Triglochin* rhizomes above soil	–	Younger than earthquake	Before 380	Before 560	*Before 390*	*Before 550*	Niawiakum River, Oyster locality (AHH, p. 12, 44)
	Willapa Bay	D	QL-4826	1720	25	40	Spruce root in soil, rings 1–15	8	–	240	430	*340*	*410*	Naselle River, 80 m upstream from locality 20 of A92
	Grays Harbor	B	QL-4882	1710	17	27.2	Spruce root in soil, rings 18–22	20	–	270	440	*340*	*410*	Johns River, 1.8 m depth at locality 14 of A92, site JR-1 of Shennan and others (1996)
	Columbia River	E	QL-4922	1698	15	24	Spruce root in soil, rings 18–25	21.5	–	280	440	*340*	*410*	Lewis and Clark River, 46°07.07′N, 123°51.65′W
	Willapa Bay	D	QL-4915	1696	15	24	Spruce root in soil, rings 22–24	23	–	280	440	*340*	*410*	Naselle River, locality 20 of A92, 46°23.23′N, 123°49.71′W

Appendix 1 (Continued)

Event	Estuary	Site (Fig. 11a)	Lab no.	Sample age (radiocarbon yr B.P.). One-standard deviation counting error excludes [E] or includes [I] the multiplier at lower right in Fig. 11: Age	E	I	Material dated (Fig. 11b; ring number 1 adjoins bark; spruce, *Picea sitkensis*; redcedar, *Thuja plicata*)	Sample age with respect to time of earthquake (–, not applicable): Numerical mean or range (ring years before plant death)	Qualitative (loosely limiting ages)	Earthquake age not shaved in Oxcal (individual ages in Fig. 11d; calibrated with data of Stuiver *et al.*, 1998; curves in Fig. 11c)[a]		Earthquake age shaved in Oxcal (event age ranges plotted in Figs 10d and 11d; italicized, individual ages not plotted)[a]		Location (AHH, Atwater & Hemphill-Haley, 1997; A96, Atwater, 1996; A92, Atwater, 1992, Table 1)
	Grays Harbor	A	QL-4912	1716	16	25.6	Spruce root in soil, rings 60–69	64.5	–	310	480	*340*	*410*	Chehalis River, 46°58.70′N, 123°46.87′W
	Willapa Bay	C	QL-4796	1740	15	24	Redcedar root in soil, rings 60–69	64.5	–	300	450	*340*	*410*	Niawiakum River, Pool locality of AHH (1997, p. 12, 64)
N												−670	−470	Combined age for event
	Willapa Bay	C	QL-4824	2540	25	40	Shrub root in soil	10–30	–	−800	−500	*−670*	*−470*	Naselle River, 46°23.34′N, 123°50.19′W
	Columbia River	D	UB-4497	2472	29	29	Spruce root in soil, rings 62–71	67	–	−700	−340	*−670*	*−470*	Lewis and Clark River, 46°07.05′N, 123°52.35′W
	Grays Harbor	A	Beta-113267	2550	60	120	Spruce root in soil, rings ≥ (71–82)	60–100	–	−850	−300	*−670*	*−470*	East Fork Hoquiam River, 47°01.07′N, 123°52.51′W
	Grays Harbor	A	QL-4930	2508	16	25.6	Spruce root in soil, rings ≥ (118–124)	120–160	–	−670	−390	*−670*	*−470*	East Fork Hoquiam River, 47°01.07′N, 123°52.51′W
	Willapa Bay	B	QL-4715	2475	23	36.8	Spruce cones on soil	–	Older than earthquake	After 770	After 410	*After −790*	*After−530*	Niawiakum River, Redtail locality (AHH, p. 12, 28)
L												−975	−895	Combined age for event
	Grays Harbor	A	QL-4916	2846	17	27.2	Spruce root in soil, rings 16–24	20		−1110	−890	*−975*	*−895*	East Fork Hoquiam River, 47°01.07′N, 123°52.51′W
	Willapa Bay	C	QL-4917	2793	16	25.6	Spruce root in soil, rings 20–26	23		−980	−820	*−975*	*−895*	Willapa River, Jensen locality at horizontal coordinate 52 m (AHH, p. 12, 70)
	Grays Harbor	B	QL-4883	2791	11	17.6	Spruce root in soil, rings 21–27	24		−980	−830	*−975*	*−895*	Blue Slough, 2.9 m depth at locality 10 of A92, 46°56.85′N, 123°43.42′W
	Willapa Bay	E	QL-4914	2814	17	27.2	Spruce root in soil, rings 23–27	25		−1020	−870	*−975*	*−895*	Naselle River, 46°23.34′N, 123°50.19′W
	Willapa Bay	D	QL-4923	2811	16	25.6	Spruce root in soil, rings 27–31	29		−1020	−860	*−975*	*−895*	Niawiakum River, Pool locality at horizontal coordinate 7 m (AHH, p. 12, 64)
	Columbia River	F	UB-4496	2886	29	29	Spruce root in soil, rings 74–82	78		−1140	−860	*−975*	*−895*	Lewis and Clark River, 46°07.05′N, 123°52.35′W
J												−1440	−1355	Combined age for event
	Willapa Bay	A	QL-4919	3177	16	25.6	Spruce root in soil, rings 48–54	51		−1470	−1350	*−1440*	*−1355*	South Fork Willapa River, 3.2 m depth at locality 15 of A92, 46°40.34′N, 123°59.18′W
	Willapa Bay	C	QL-4884	3200	17	27.2	Spruce root in soil, rings 51–60	54.5		−1470	−1350	*−1440*	*−1360*	Naselle River, 46°23.34′N, 123°50.19′W
	Willapa Bay	B	QL-4718	3165	17	27.2	Spruce cones on soil		Older than earthquake	After 1520	After 1390	*After −1520*	*After −1400*	Niawiakum River, Redtail locality (AHH, p. 12, 28)
	Willapa Bay	B	QL-4717	3166	18	28.8	Twigs on soil		Older than earthquake	After 1520	After 1390	*After −1520*	*After −1400*	Niawiakum River, Redtail locality (AHH, p. 12, 28)
	Willapa Bay	B	QL-4716	3180	30	48	Stick on soil		Older than earthquake	After 1600	After 1310	*After −1600*	*After −1380*	Niawiakum River, Redtail locality (AHH, p. 12, 28)

[a] Probably contains 95-percent confidence interval. In cal yr A.D. [+] and cal yr B.C. [−] (converted to cal yr before A.D. 2000 in Figs 10d and 11d).
[b] Ages reported by Atwater *et al.* (1991), recalibrated in this table.