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Geological criteria for evaluating seismicity revisited: Forty years of paleoseismic investigations and the natural record of past earthquakes

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ABSTRACT

The identification of individual past earthquakes and their characterization in time and space, as well as in magnitude, can be approached in many different ways with a large variety of methods and techniques, using a wide spectrum of objects and features. We revise the stratigraphic and geomorphic evidence currently used in the study of paleoseismicity, after more than three decades since the work by Allen (1975), which was arguably the first critical overview in the field of earthquake geology. Natural objects or geomarkers suitable for paleoseismic analyses are essentially preserved in the sediments, and in a broader sense, in the geologic record. Therefore, the study of these features requires the involvement of geoscientists, but very frequently it is a multidisciplinary effort. The constructed environment and heritage, which typically are the focus of archaeoseismology and macroseismology, here are left aside. The geomarkers suitable to paleoseismic assessment can be grouped based on their physical relation to the earthquake's causative fault. If directly associated with the fault surface rupture, these objects are known as direct or on-fault features (primary effects in the Environmental Seismic Intensity [ESI] 2007 scale). Conversely, those indicators not in direct contact with the fault plane are known as indirect or off-fault evidence (secondary effects in the ESI 2007 scale). This second class of evidence can be subdivided into three types or subclasses: type A, which encompasses seismically induced effects, including soft-sediment deformation (soil liquefaction, mud diapirism), mass movements (including slumps), broken (disturbed) speleothems, fallen precarious rocks, shattered basement rocks, and marks of degassing (pockmarks, mud volcanoes); type B, which consists of remobilized and redeposited sediments (turbidites,

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homogenites, and tsunamites) and transported rock fragments (erratic blocks); and type C, entailing regional markers of uplift or subsidence (such as reef tracts, micro-atolls, terrace risers, river channels, and in some cases progressive unconformities). The first subclass of objects (type A) is generated by seismic shaking. The second subclass (type B) relates either to water bodies set in motion by the earthquake (for the sediments and erratic blocks) or to earthquake shaking; in a general way, they all relate to wave propagation through different materials. The third subclass (type C) is mostly related to the tectonic deformation itself and can range from local (next to the causative fault) to regional scale.

The natural exposure of the paleoseismic objects—which necessarily conditions the paleoseismic approach employed—is largely controlled by the geodynamic setting. For instance, oceanic subduction zones are mostly submarine, while collisional settings tend to occur in continental environments. Divergent and wrenching margins may occur anywhere, in any marine, transitional, or continental environment. Despite the fact that most past subduction earthquakes have to be assessed through indirect evidence, paleoseismic analyses of this category of events have made dramatic progress recently, owing to the increasingly catastrophic impact that they have on human society.

INTRODUCTION

Paleoseismology is a currently growing and developing earth sciences discipline. The recognition that paleoseismology has in fact become a branch of learning in itself is a rather recent development (Slemmons, 1957; Allen, 1975; Sieh, 1978; Crone and Omdahl, 1987; Vittori et al., 1991; Yeats et al., 1997; Audemard, 1998; McCalpin, 1996, 2009). This is borne out by the recent publication of dedicated textbooks (Wallace, 1986; McCalpin, 1996; Yeats et al., 1997; Keller and Pinter, 1998; Burbank and Anderson, 2001; McCalpin, 2009) and special issues of scientific journals (e.g., Serva and Slemmons, 1995; Michetti and Hancock, 1997; Pavlides et al., 1999; Pantosti et al., 2003; Michetti et al., 2005a; Reicherter et al., 2009, this volume) or conference proceedings (e.g., Michetti, 1998; Cello and Tondi, 2000; HAN2000–PALEOSIS Project; Okumura et al., 2000, 2005; Toda and Okumura, 2010). Also, special issues or proceedings, and even books, on more specific paleoseismic issues have also come to light (e.g., Maltman [1994] on soft sediment deformation; Stratton Noller et al. [2000] on Quaternary dating techniques; Comerci [2001] on seismically induced mass movements; Gilli and Audra [2001] on speleoseismology; and Tappin [2007] on the sedimentary record of tsunamis and tsunamites). Finally, concerted efforts within the International Union for Quaternary Research (INQUA) community for several years have led to the recent proposal and acceptance (currently under validation) of a new macroseismic scale based exclusively on earthquake environmental effects, named the Environmental Seismic Intensity Scale 2007, known under the acronym ESI 2007 (Michetti et al., 2007).

Most researchers agree that the main aim of paleoseismology is the identification and characterization of past earthquakes from geo-archives. In a broader sense, this type of research intends to establish the seismic history of a given fault or given region from deformed sediments or rocks, beyond the limited temporal resolution of instrumental and historical seismicity. However, this does not exclude major temporal overlapping with these two latter disciplines, as proposed by many authors (e.g., Gürpinar, 1989; Audemard, 1998, 2005; Levret, 2002; Carrillo et al., 2009). In regions of the world where historical and/or instrumental seismicity studies are really scarce, or where earthquake recurrence is too long to allow any detection of large past events, the study of past seismic behavior of faults has to rely exclusively on paleoseismology. Furthermore, paleoseismic research can even reveal very distinct trends of seismic activity over longer periods of time (longer than the current interglacial period), as is the case in large regions of the world subject to glaciation-deglaciation cycles (e.g., Adams, 1996b; Mörner et al., 2003; Mörner, 2005). On the other hand, a common misconception about paleoseismology relates this discipline to the study of prehistoric earthquakes only, as originally defined by Wallace (1981). This surely used to be, and still is, its main focus, but it does not cover all current potential spectra and unnecessarily restrains paleoseismology to a strict time window. In fact, paleoseismology should be understood as the study of the ground effects from past earthquakes preserved in the geologic and geomorphic record (Michetti et al., 2005b), regardless of the time of occurrence. In a more general way, it aims at studying any geologic deformation related to earthquakes (Audemard, 2005). In that sense, the separation among disciplines studying past individual earthquakes, i.e., instrumental seismology, historical seismology, archaeoseismology, and paleoseismology, should not be based on strict time windows of observation, but on the type of source information that is being used and managed by the researchers. In other words, the distinction between these four different disciplines is fundamentally methodological, both in terms of applied tools and research objects. Consequently, we share the proposal of Caputo and Helly (2008), in which each discipline focuses on different objects: (1) analogue or digital instrumental records, in modern seismology; (2) oral or written witnesses, in historical seismology; (3) artifacts, in archaeoseismology, where artifact is defined as any product of human activity; and (4) natural features, in paleoseismology. Particularly, "natural object" in this paper is a generic name given to all geologic and geomorphologic features, indicators, evidence, effects, and markers pinpointing the occurrence of an earthquake.

The methods and techniques employed for paleoseismic characterization of source parameters are diverse and numerous (for a thorough review, refer to reference textbooks such as McCalpin, 1996; Yeats et al., 1997; McCalpin, 2009) but are still evolving. Methodologies are continuously changing and adapting to encompass very broad and multidisciplinary approaches for the characterization of past earthquakes. Fault trenching investigations have become a cornerstone for paleoseismic analysis because they have the potential to provide a direct assessment of the amount and timing of recent coseismic and total (coseismic + aseismic) fault slip. This has led to a common misconception that trenching and paleoseismology are almost synonymous or equivalent (e.g., Caputo and Helly, 2008; Carrillo et al., 2009). As a matter of fact, trenching is only one technique among many other paleoseismic approaches, with its own specific advantages and limitations. For further details on this issue, we refer to McCalpin (1996), Yeats et al. (1997), Audemard (2005), Michetti et al. (2005b), and McCalpin (2009), among other reviews.

The study objects that are the focus of paleoseismic investigations have also evolved with time and have particularly increased in number as a response to new demands in the field of seismic hazard assessment (SHA). From the understanding that a tectonic deformation may induce a sedimentological response, which can then be fossilized and preserved in the geologic record, paleoseismologists, and particularly geologists and tectonic geomorphologists, have continuously developed new paleoseismic objects to derive the timing and magnitude of hazardous earthquakes. These developments in paleoseismology have led to the incorporation of an ever-growing number of earth sciences disciplines, such as seismology, Quaternary geology, tectonics, structural geology, sedimentology, (sequence) stratigraphy, pedology, geomorphology, geochronology, remote sensing and geophysical prospecting methods, such as seismic reflection, side-scan sonar, ground-penetrating radar (GPR), and electric tomography, as well as geodetic surveying and monitoring techniques, such as global positioning system (GPS) and electron distance meter (EDM), among several others. In fact, the present contribution aims at answering the following question that any researcher may pose himself or herself anytime they are requested to assess the seismic hazard of a given fault or region: What object(s) can I study in this region in order to know the level of seismic hazard to which the planned construction project will be subjected?

MAIN POTENTIAL STUDY OBJECTS

The study and understanding of contemporary earthquakes and their associated effects have provided the basis for the search, recognition, and characterization of the types of evidence that paleoseismology needs to unravel from the geologic record. Owing to the physics of earthquakes, two major types of deformations typically occur at ground or sea-bottom surfaces. One set of features relates directly to the earthquake rupture process itself, while other features are the result of the accompanying seismic shaking. Serva (1994), McCalpin (1996), and Michetti et al. (2007) referred to these as primary and secondary evidence, respectively. Depending on their location (or proximity to the causative fault or surface break), McCalpin (1996) subdivided these mechanisms into on-fault (near-field) and off-fault (far-field) evidence. Since the secondary features can occur throughout the entire affected region, including on or across the surface break, Audemard (2005) preferred to subdivide the coseismically induced permanent ground deformations into two groups based on their genesis and degree of association with the earthquake, namely direct and indirect ground deformations. The first set of features is directly related to the fault plane and its kinematics. The second group of features includes both the seismically induced effects (mass instability and soft-sediment deformation, including liquefaction) and all other ground modifications at the local or regional scale, such as uplift and subsidence, warping, buckling, bulging, etc. In order to incorporate the many newly employed paleoseismic study objects or targets, this latter classification needs to be given a wider meaning that does not restrict it to ground deformation only and that provides room to include other indirect types of evidence of earthquakes, such as seismites, which do not imply any ground-surface deformation, but refer to sediment remobilization and redeposition, broken or affected speleothems, truncated bioherms, tree growth changes recorded in dendrochronology samples, tsunamites, and even some types of soft-sediment deformations that include sills and convoluted bedding. McCalpin (1996) added a third level of subdivision based on the timing of the feature with respect to the time of occurrence of the earthquake, namely instantaneous (coseismic) and delayed response (postseismic). The need for this third level of classification does not appear to have any practical application because paleoseismic studies typically can make no distinction between coseismic and postseismic processes. Furthermore, the time taken by delayed processes (e.g., deposition of colluvial wedges, filling of open cracks, remobilization and redeposition of sediments, tree growth recovery or healing, among others) is negligible compared to the recurrence interval between large earthquakes in most faults.

We herein propose to subdivide the indirect or off-fault objects into three subgroups or types: Type A encompasses seismically induced effects, including soft-sediment deformation (soil liquefaction, mud diapirism), mass movements (including slumps), broken or disturbed speleothems, fallen precarious rocks, shattered basement rocks, and marks from degassing (mud volcanoes, seeps, pockmarks). Type B consists of remobilization and redeposition of sediments (turbidites, homogenites, and tsunamites) and rocks (e.g., large boulders launched by tsunami waves on rocky coasts, i.e., erratic blocks). Type C includes regional markers of tectonic uplift or subsidence, such as flexural-slip thrust faults, reef tracts, (truncated) bioherms, micro-atolls, terrace risers, river channels, progressive unconformities, and mountain fronts.

ON-FAULT PALEOSEISMIC OBJECTS

On-fault features, i.e., those related to fossil or fresh surface rupture of earthquakes, attracted the attention of researchers from the early days of paleoseismology. Owing to the scarcity of natural exposures of these features (with very rare exceptions, such as the case studied by Audemard et al. [1999] and also the spectacularly preserved surface ruptures of normal fault earthquakes in Nevada [e.g., Slemmons, 1957; Wallace, 1984]), or the rare availability of anthropogenic cuts across them (e.g., Nicol and Nathan, 2001; Mörner et al., 2003; Kuhn, 2005; López, 2006; López C. and Audemard M., this volume), paleoseismic trenching has become the most widespread technique of viewing these ground or sedimentary disturbances in the late Quaternary geologic record. We could actually state that paleoseismology at its early stages essentially developed around trenching studies. Trenching is typically practiced across active fault traces that are currently well identified by means of their geomorphic expression via fault-related landforms (D in Fig. 1A and I in Fig. 1B). Several hundreds of these studies are reported in the literature (e.g., Sieh, 1978; Machette et al., 1987; Crone and Luza, 1990; Van Dissen et al., 1992; McCalpin et al., 1993; Okumura et al., 1994; Audemard, 1996; Michetti et al., 1996; Townsend, 1998; among many others), but even larger numbers of paleoseismic studies are of a confidential nature or can only be found in unpublished reports, particularly in the United States, as many of these studies have been performed in compliance with regulations or legislation. This volume includes several new cases of these types of studies (Audemard, Clark, Lalinde P. et al., López C. and Audemard M., McCalpin, Olig et al.). Paleoseismic trenching strongly relies on a previous very thorough and detailed neotectonic assessment of the fault or region (Audemard and Singer, 1996, 1997, 1999; Audemard, 2005; Michetti et al., 2005b) to identify the best potential trenching sites. The main aim of this technique is to expose the interplay of fault activity with recent sedimentation in trench walls. Young (late Pleistocene to Holocene) sediments record perturbations that are directly related to the fault interference with the sequence. We emphasize that successful results from trench assessments heavily rely on the recognition of active or capable fault traces and the understanding of the interaction between tectonics and sedimentation at the chosen trench site, prior to excavation. The only way of bypassing this lengthy procedure of trench-site selection is when a trench can be directly excavated across a newly ruptured fault surface break (i.e., the Cariaco 1997 earthquake surface rupture studied by Audemard [1999a, 1999b, 2007, this volume], and the Chi-Chi 1999 rupture studied by Chen et al. [2004]), in which case an accurate location of the seismogenic fault trace(s) and the tectonic style of the active fault are provided. Finally, the trenching approach is not only applied to on-fault features but also to all those perturbations indirectly induced by active faulting or by seismic shaking that may be suitably recorded in the sedimentary sequence (Fig. 1). This issue shall be dealt with in more detail later in this paper.

In particular settings, geophysical investigations have been of great help in trench studies. The most commonly applied geophysical methods are ground-penetrating radar, which is widely known by its acronym GPR (e.g., Cai et al., 1996; Jol et al., 2000; Reicherter et al., 2003; Ollarves et al., 2004; Maurya et al., 2005), and electrical resistivity tomography (ERT; e.g., Silva et al., 2001; Caputo et al., 2003; Colella et al., 2004; Fazzito et al., 2009). These methods have helped to locate the active fault trace with a much higher precision; these traces often exhibit subdued or very subtle earthquake-related geomorphic features or are situated in land that has been heavily modified by anthropogenic activities (e.g., Cushing et al., 2000; Dost and Evers, 2000; Jongmans et al., 2000; Lehmann et al., 2000; Meghraoui et al., 2000; Verbeeck et al., 2000; Demanet et al., 2001). In the particular case of the surface trace of a gently dipping thrust fault that tends to parallel and mimic the topographic contours, the localization of the fault trace solely on the basis of morphologic expression is very difficult and requires the support of these geophysical surveys prior to trenching (e.g., Chow et al., 2001; Fazzito et al., 2009). In addition, GPR and/or ERT have also been applied to onshore strike-slip faults, where often the fault splays into several branches, and these techniques serve to determine the branches that are the most active (e.g., Gross et al., 2000; Wise et al., 2003; Audemard et al., 2006; Kürçer et al., 2008).

The most common on-fault geomorphic features used as paleoseismic indicators, from a geomorphologic viewpoint, are fault scarps or counterscarps, sag or fault ponds, pop-ups or pressure ridges (at smaller scale, these are known as mole tracks), shutter ridges, and open fissures, among others. Regardless of the tectonic style (thrust, normal, or strike-slip faulting), fault scarps are definitely the most widely assessed feature due to their common occurrence (D and G in Fig. 1A and I in Fig. 1B). Furthermore, they constitute a perturbation at the ground surface, which triggers morphodynamic processes such as erosion, redeposition, and surface smoothing that have a sedimentary signature and are prone to fossilization. Depending on the balance between tectonic and sedimentation/erosion rates at the scarp, three different situations are recognized that are recorded distinctively: (1) When tectonic rate is higher than sedimentation rate, the scarp grows through time and sedimentation is bounded to the "low-topography" block; (2) when tectonic and sedimentation rates are similar, available space is filled by sediments and the fault-related landform is leveled; and (3) when sedimentation rate is higher than the tectonic rate, the landform is buried by sediments. This last condition leads to blind normal or thrust faulting, which in turn induces growing fault-propagation folding. Depending on whether the difference between sedimentation and tectonic rates is small or large, this deformation ought to be assessed either as on-fault evidence or an off-fault feature, respectively. In regions where event dating is difficult to impossible due to lack or scarcity of datable materials (K in Fig. 1B), techniques based on the diffusion equation (e.g., Begin, 1992; Hanks, 2000) have been applied to date scarps by estimating the degree of scarp degradation. This technique experienced a boom in the western United States in the 1970s and 1980s (e.g., Bucknam and Anderson, 1979; Nash, 1980; Colman and Watson, 1983), but its generalized application on a worldwide scale has become very limited because it is strongly dependent on local climate, which may vary quite rapidly within a basin or along a mountain front. It definitely depends on a local diffusivity (Phillips et al., 2003), which is different from place to place. In recent years, cosmogenic dating (radionuclides ^{10}Be, ^{26}Al, and ^{36}Cl, and stable nuclides ^{3}He and ^{21}Ne) of fault scarps by exposure to cosmogenic radiation has largely replaced the scarp degradation technique. This approach has been applied to fault scarps both in Quaternary alluvial deposits (e.g., Phillips et al., 2003; Siame et al., 2006) and bedrock (e.g., Zreda and Noller, 1998, 1999; Mitchell et al., 2001; Siame et al., 2006). Most commonly, this technique provides data on tectonic slip rates during the time span of scarp exposure to cosmogenic radiation (e.g., Ritz et al., 1995; Van der Woerd et al., 1998; Palumbo et al., 2004). In a similar manner, it has been used for dating nontectonic scarps, such as sackung uphill-facing scarps (B in Fig. 1A) by Hippolyte et al. (2006), which could also eventually provide pertinent paleoseismic information in the cases where these large mass instabilities were seismically triggered (refer to section "Seismically Induced Effects" later herein).

As mentioned earlier, scarps are the features most commonly trenched for paleoseismic purposes. The success of this type of investigation is enhanced if sag- or fault-bounded ponds form against them. However, tectonic scarps are seldom preserved in the case of thrust faults, unless the fault cuts across very competent rocks. Most Quaternary thrust faults cut across soft or unconsolidated deposits, since they tend to crop out at the foot of the mountain front. It is frequently the case that tectonic slip rates of such thrust faults are much less than the sedimentation rate, which can lead to an almost complete masking of the active thrust fault by younger sediments, creating the situation of a blind thrust fault. If no other option is available, paleoseismologists have overcome this situation by trenching sympathetic faults—secondary faults that form in mechanical and kinematical association with the main thrust fault, such as scarps of flexural-slip thrust faults (Fig. 1B; e.g., Yeats, 1986a, 1986b; Costa et al., 1999; Livio et al., 2009). Back-thrust faults are also a potential trenching target. They happen to exhibit a higher dip than the main thrust and very frequently face in an upstream direction, which offers a better scenario for the interaction between tectonics and sedimentation. The major limitation in these situations is the low degree of certainty the researcher has on the actual kinematic activation of the sympathetic and/or antithetic faults in relation to the coseismic slip of the seismogenic thrust fault. In other words, the sympathetic fault does not necessarily move each time the main thrust fault does, which converts it into a poorly reliable chronometer. Moreover, although mechanically connected, there are as yet no published cases of the relationship between actual slip on the major fault and the induced slip on the flexural-slip faults. Earthquake size estimates in this case should be based on: (1) intensity data (Serva, 1994), which are then converted into magnitude values, when available, and (2) the application of the concept of seismic landscape (Michetti et al., 2005b). The definition of the relationship between macroseismic intensity and earthquake ground effects was in fact the rationale for the introduction of the ESI 2007 Intensity Scale proposed by INQUA (Michetti et al., 2007; Reicherter et al., 2009).

Generally speaking, all the previous geomorphic objects, which favor the continuous fine-grained, thinly bedded sedimentation against the fault plane, constituting the ideal setting for the recording of future ground-surface deformations, have proved to be really useful for paleoseismic studies. After overcoming water problems (through draining/water pumping in tropical regions and digging frozen ground in temperate regions), sag and fault ponds have generally provided the best and most complete paleoseismic records (e.g., Audemard, 1997; Lindvall et al., 2002; Audemard, this volume). Several favorable geologic conditions (for more details, refer to Audemard, 2005) have to come together at these "trenchable" sites in order to increase the chances of successful seismic hazard assessment. The paleoseismic assessment of any of the aforementioned objects can yield information on most or all of the following aspects (Audemard, 2005): confirmation of Holocene fault activity, slip per event and average slip rate of a given fault or fault segment, slip vector decomposition from fault-plane kinematic indicators, recurrence intervals and magnitude of the larger earthquakes (seismic potential) on known faults, coseismic rupture length, fault segmentation, fault interaction as a consequence of stress loading by stick-slip on contiguous faults, time-space distribution of seismic activity along a given tectonic feature, elapsed time since latest event and estimated time to the next event, which determines the likelihood of occurrence of a future earthquake, seismotectonic association of historical earthquakes, and short- and long-term landscape evolution.

In cut or trench exposures, the most frequent kinds of geologic evidence of the interaction of sedimentation and faulting (Amit et al., 1995; Audemard, 2005), which are interpreted as the result of individual earthquakes, are: (1) filling of ground-surface open fissures (tension or T cracks, synthetic Riedel or R shears, or their combination); (2) fault-scarp–bounded deposition on a downthrown block; (3) a colluvial wedge derived from scarp degradation/subduing, occasionally interspersed with sedimentation on the downthrown fault compartment; (4) event horizon (faulting sealed by a leveling depositional episode); and (5) normal-fault propagation folding, later buried by overlying beds.

Trees have been excellent time recorders—chronometers—of the occurrence of large contemporary earthquakes, in association with, as well as away from, their surface breaks (e.g., Sheppard and Jacoby, 1989). The study of the annual growth rings of certain particular plant species, known as dendrochronology, which is somewhat equivalent to varve chronology in lakes of temperate regions (which has applications in higher-latitude regions of the globe that exhibit a marked seasonality),

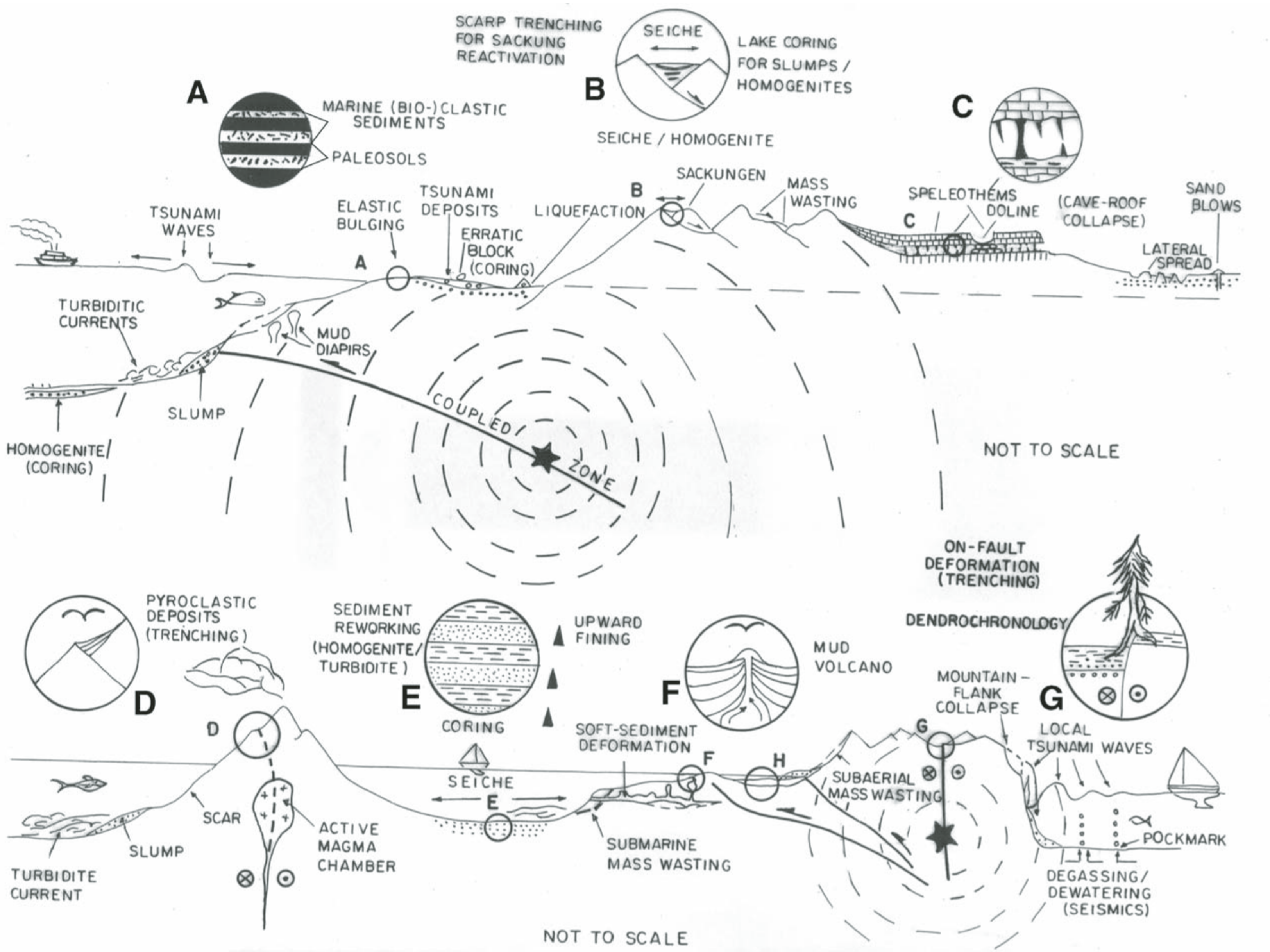

Figure 1. Generalized sketches of different potential settings of coexisting natural indicators of seismic activity (see text for explanation). (A–G) Upper cartoon depicts a subduction setting, where the paleoseismic geomarkers in relation to the marine environments are to the left, while those related to the orogen are to the right. The lower cartoon emphasizes the objects related to several subaqueous subenvironments. (H–L) The upper diagram tries to show the potential paleoseismic objects existing in a compressional environment, dominated by either outcropping or blind thrust faults. The lower diagram summarizes the paleoseismic objects to seek in a cratonic environment subject to glaciation-deglaciation processes, such as Scandinavia. LGM—Last Glacial Maximum.

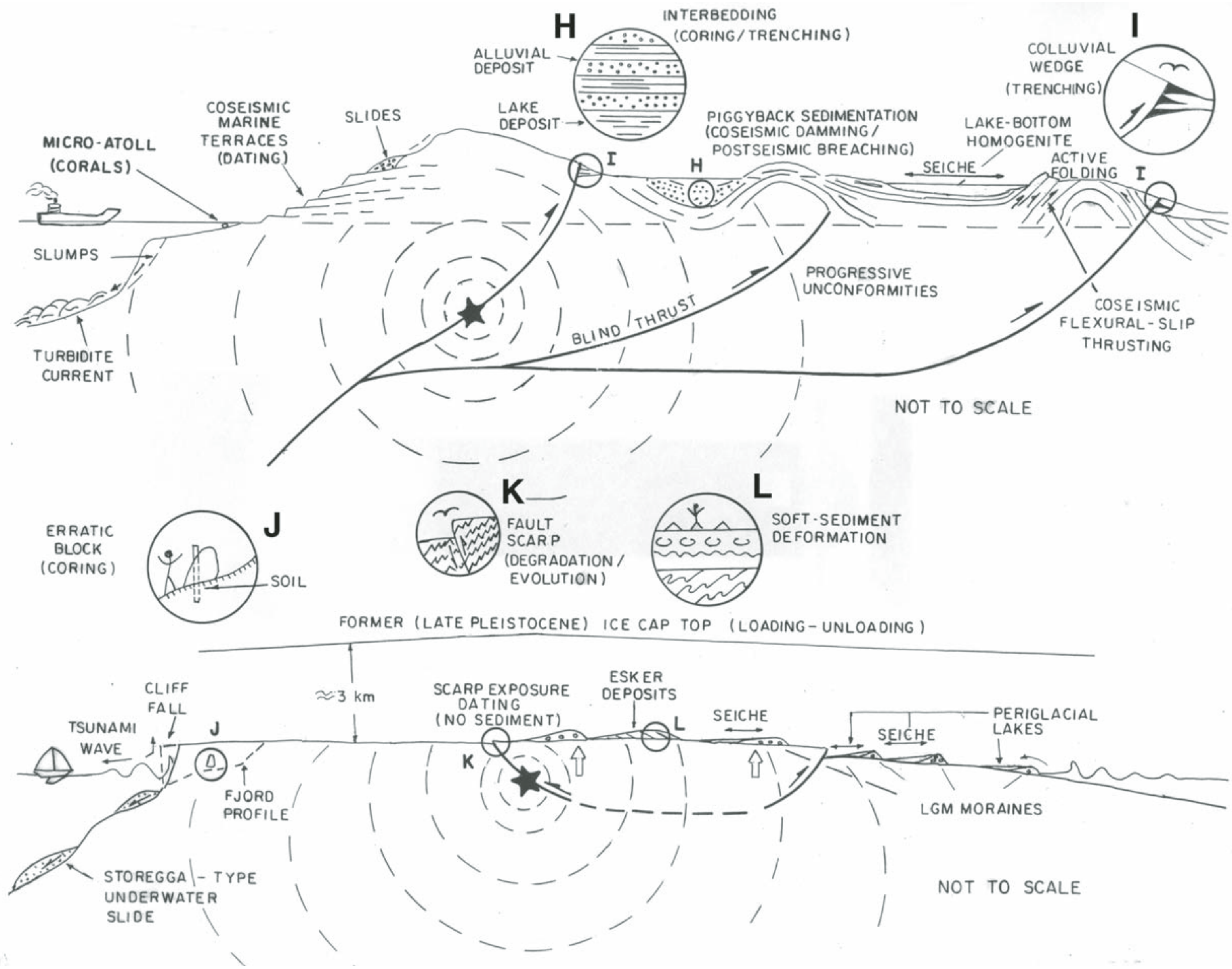

Figure 1. (*Continued.*)

has revealed that certain growth changes (ring eccentricity, dead and healing zones, among others) that are the result of seismically induced perturbations can be recorded in big, long-lived tree trunks (e.g., Meisling and Sieh, 1980; Yamaguchi et al., 1997). In response to the natural positive phototropism exhibited by most large trees, loss of tree verticality—tilting of a tree—is soon naturally repaired in the years directly after the perturbing event, which is nicely recorded in the growth of annual rings. This has been shown in localities close to or right on top of contemporary earthquake ruptures (e.g., Page, 1970; Sheppard and White, 1995; Jacoby et al., 1997; Lin and Lin, 1998; Carver et al., 2004). In some cases, disruption of tree roots by ground cracks (e.g., Tasdemiroifelgülu, 1971; Carver et al., 2004) is marked by a deceleration of ring growth, which records the partial or total death of the tree (e.g., Toppozada et al., 2002; Doser, 2004). Even more astonishingly, large trunks, especially of sequoia trees, have been split into two by a crosscutting earthquake surface rupture (G in Fig. 1A; e.g., Carver et al., 2004). This illustrates that growth anomalies of tree rings can be considered as direct or indirect evidence of surface faulting, depending on the degree of association with the ground rupture. The limitation of this kind of evidence, as seen from a paleoseismic viewpoint, resides in the low probability of its preservation and the even lower probability of actually finding the preserved evidence. It has certainly proved to be useful as long as the tree lives. In that sense, this technique in very recent times is being used widely for mass movement monitoring and detection (e.g., Pop et al., 2008; Surdeanu et al., 2008), although its applicability has been known for quite a while (e.g., Terasmae, 1975; Hupp et al., 1987). In coastal areas along subduction zones, it has been observed that after the occurrence of large subduction earthquakes, large coastal areas have drowned after the seismic energy was released, as evidence of coseismic subsidence, due to elastic rebound. This has led to the death of extensive stands of trees as a result of the flooding by marine-water incursions of the low-lying areas and is commonly referred to as "drowned or ghost forests" (e.g., Atwater, 1987; Hamilton et al., 2005). This type of event is preserved in the sedimentary record as a couplet consisting of a horizon of organic-rich terrestrial sediments overlain by a marine sediment layer (A in Fig. 1A; e.g., Nelson et al., 1996; Cundy et al., 2000).

OFF-FAULT PALEOSEISMIC OBJECTS

The term "off-fault evidence" as defined by McCalpin (1996), or "indirect perturbations" as originally defined by Audemard (2005) and enlarged herein, covers all the earthquake-related or earthquake-triggered evidence different from ground-surface rupturing coinciding with the surface trace of the causative fault, although they can occasionally concur with the fault plane as near-field features. To illustrate this eventual coalescence of processes, a sand dike, which results from earthquake shaking that triggers liquefaction, takes advantage of the recently displaced causative fault plane, as has been observed in trench walls by Rockwell et al. (2001) and Audemard et al. (2008), because it probably has a transient higher permeability that facilitates the venting process several seconds to a few minutes after the earthquake occurrence. Since the off-fault evidence is commonly not associated with the causative fault plane, its paleoseismic interpretation needs more elaboration than the on-fault objects. For instance, the interpretation of these features in terms of earthquake magnitude is rarely unequivocal, unless their evaluation is carried out spatially (e.g., Ricci Lucchi, 1995).

Each object of the same nature or type (e.g., slide, sand blow, or broken speleothem) should be envisaged as a single "seismometer," and only the integration of them in a "network" relation can provide a reliable estimate of earthquake magnitude. This tacitly implies that all objects of the same type need to be ascribed to the same event. They must be temporally related; otherwise they cannot belong to a network. Significant progress has been achieved by this approach for mass movements and seismically induced liquefaction. Using modern earthquakes as analogues, relationships between earthquake magnitude and size or spatial distribution of slides have been derived (e.g., Youd and Hoose, 1978; Keefer, 1984; Wilson and Keefer, 1985; Keefer and Manson, 1989; Rodríguez et al., 1999; Crozier, 1992; Keefer, 2000) that can be applied, as an inverse method, to fossil landslides in paleoseismic studies to estimate the magnitude and localization of the epicenter of the triggering event, and by association, the causative fault as well. The same procedure has been carried out for liquefaction features (e.g., Kuribayashi and Tatsuoka, 1975; Youd and Perkins, 1987; Ambraseys, 1988; Barlett and Youd, 1992; Papadopoulos and Lefkopoulos, 1993; Munson et al., 1995; Allen, 1986; Galli, 2000; Castilla and Audemard, 2002; Rodríguez et al., 2002; Rodríguez Pascua et al., 2003; Papathanassiou et al., 2005; Rodríguez et al., 2006; Castilla and Audemard, 2007). On the other hand, and this is a very sensitive issue, the size of these objects may have no relation with the earthquake magnitude or the seismic attenuation laws. For instance, the lateral extent and thickness of homogenites, and tsunamites, tend to reflect pre-earthquake local conditions of the area where they occur. This is also the case for sand dikes formed in association with lateral spreading (Fig. 1A). Their final width, contrary to claims made by Obermeier (1996), is strongly conditioned by the size of the nearby available void (both in width and height) that sets in motion the horizontally sliding mass, creating the fractures through which venting will occur. In many cases, these off-fault features provide a good time constraint for the occurrence of the inducing earthquake. So, they can be considered as reliable chronometers. The features implying redeposition appear to be best suited in this respect. These include tsunamites, homogenites, and erratic (tsunami-transported) boulders and blocks, but also broken speleothems, affected bioherms, mass movements, particularly those with ground-surface disruption, and even sand blows. In fact, their occurrence gives a good indication of the timing of the earthquake. On the other hand, other objects can merely reveal that an area is seismo-tectonically active, only due to their occurrence within the sedimentary record. Many soft-sediment deformations, even after reliably confirming their seismic origin, fall into this category. All those soft-sediment deformations that maintain parallelism to the stratification after remobilization, including convolute bedding, contorted layers, basal surface of slumped beds, flame structures, ball and pillars, sand-vented sills, etc., are useless in terms of earthquake chronometers. Also, penetrative liquefaction features, such as sand dikes, like the ones depicted by Rodríguez-Pascua et al. (2000), need careful handling in terms of age determination of the triggering earthquake. If there is no convincing proof that such a dike reached the pre-earthquake ground surface by being connected to a sand blow at the top, the age of the youngest disrupted bed should be considered as the oldest age of occurrence of the associated earthquake. In other words, the age of the event in such a case is younger than the last affected stratigraphic marker. This is also valid in the cases when the sand-venting dike is later truncated by erosion.

Based on their generating mechanism, in combination with the paleoseismic approach that could be applied to their study, we have subdivided the indirect or off-fault objects into three subcategories: Type A includes seismically induced effects; type B consists of remobilization and redeposition of sediments and/or rocks; and type C entails regional markers of vertical deformation (surface level changes). These are the features that have been the subject of significant development in the past years; in some cases, huge progress has been made. Paradoxically, the inaccessibility of subduction fault zones can be primarily held responsible for the recent progress in the study of all these indirect types of earthquake evidence, as anticipated by Adams (1996b). The reality that subduction zones lie underwater and are responsible for the largest known and most destructive contemporary earthquakes worldwide (e.g., 22 May 1960, Valdivia; 27 March 1964, Anchorage; 26 December 2004, Sumatra; and 27 February 2010, Concepción, earthquakes), in combination with their very large extent around the Pacific rim, as well as their position in relation to large population centers, with their costly infrastructure along the Pacific shores, make the assessment of their seismic potential an urgent necessity. This all has led to the realization that it is necessary to study ground deformations and on-land sedimentary disturbances as evidence of earthquakes in the coastal zones (e.g., Atwater et al., 1995), as well as along the submarine trench during earthquakes. Adams (1996b) had already conducted a rather thorough overview of the ways in which the seismic hazard assessment of such a complex active tectonic and geodynamic setting could be tackled, by presenting the case of the West Canadian coast. A multiproxy investigation, using several of the potential indirect objects (Fig. 1), actually appears to be the most appropriate approach in order to derive the seismic history (earthquake chronology) of a given subduction segment or a given region. It includes any or some of the phenomena, such as earthquake-triggered liquefaction, coseismic land-level changes, and associated sedimentary response (interbedding of marine and continental deposits, growth increase or truncation of living [brain] corals, emerged or submerged coseismic marine terraces), and onshore tsunamites, as well as underwater mass wasting, pockmarks from degassing/dewatering, mud and sand volcanoes, offshore turbidites, and homogenites. This concept of a multiproxy approach has been applied to continental regions as well, for instance, in the interior of Canada (Adams, 1996b), in the alpine lakes of Switzerland and Italy (Becker et al., 2005; Fanetti et al., 2008), and in Venezuela (Audemard, 2005). Eventually, by applying some of the existing relationships treated in the literature for mass wasting and/or liquefaction, mentioned earlier, we will be able to make a minimum estimate of the triggering paleo-earthquake size.

Seismically Induced Effects

Mass movements and ground liquefaction are the two most common and widespread natural phenomena associated with earthquakes worldwide. These effects, induced by earthquake shaking, have been commonly recognized and described for onshore environments in historical documents over the centuries. For instance, description of liquefaction features by Sarconi (1784) and illustration of mass movements by Simón (1627) are classical examples of historical reports for seismically induced geological effects. Mass movements, because of their requirement of an energy gradient, typically occur in mountainous regions, but they can also affect large flat areas in the form of lateral spreading (Fig. 1A). Conversely, liquefaction tends to affect geologically young, low-lying, flat alluvial fill areas of fluvial or coastal environments. However, these two phenomena are not exclusive of onshore environments. On the contrary, they appear to develop underwater, in sea-bottom and lake-bottom sediments, and are more frequent and more widespread, as observed during contemporary earthquakes (e.g., Sims, 1973, 1975; Mosher et al., 1994; Nakajima and Kanai, 2000), as well as known from the past geologic record (e.g., Hempton and Dewey, 1983; El-Isa and Mustafa, 1986; Ringrose, 1989; Roep and Everts, 1992; Maltman, 1994; Hampton et al., 1996; Marco et al., 1996; Mörner, 1996; Rodríguez-Pascua et al., 2000; Mörner et al., 2000, 2003). These induced effects have proved, during many destructive earthquakes, to be more harmful than the earthquake shaking itself. As an illustration of this, there is the large destruction produced by the Prince William Sound, Anchorage-Alaska, 1964, earthquake at Turnagain Heights due to lateral spreading (Hansen, 1965). In the same way, either large mass-wasting bodies (e.g., Ancash, Perú, 1970, earthquake, with 18,000 casualties; Cluff, 1971; Plafker et al., 1971) or widespread mass movements have caused generalized destruction and numerous fatalities in urban areas, such as during the Deixi, Sichuan-China, 1933, earthquake, with 6800 casualties (Li et al., 1986). Associated with slope instabilities, an even more harmful factor has been the breaching of slide-dammed lakes days or months after the earthquake (Fig. 1B; e.g., the Mocotíes 1610 event in the Mérida Andes—Singer, 1998; Ferrer, 1999; the Deixi, Sichuan-China, 1933, earthquake, with 2500 casualties—Li et al., 1986).

As for outcropping active fault planes, trenching has been practiced onshore for both earthquake-induced gravitational scarps (B in Fig. 1A; e.g., Wallace, 1984; Crozier, 1992; Nolan and Weber, 1992, 1998; McCalpin, 1999; Onida et al., 2001; McCalpin and Hart, 2002; Tibaldi et al., 2004; Gutiérrez et al., 2005, 2008) and liquefaction features (e.g., Tuttle et al., 1990; Audemard and De Santis, 1991; Tuttle and Seeber, 1991; Clague et al., 1992; Walsh et al., 1995; Tuttle, 2001; Mörner et al., 2003; Cox et al., 2004; Guccione, 2005) in many different tectonic settings. Not only have these features proved to be the only means of assessing the seismic history of subduction zones (A in Fig. 1A), but they have been used elsewhere where faults have not ruptured the ground surface. For instance, trenching for liquefaction features has been carried out in the New Madrid area to recognize not only the effects of the 1811–1812 earthquake sequence and their extent, but also those of its precursors (Tuttle, 2001; Cox et al., 2004; Guccione, 2005). In cratonic areas subject to glaciation and deglaciation processes during the Quaternary, such as northern North America and Scandinavia, where Holocene sediments act as young recorders of recent deformation, and recognized onshore (intraplate) faults with Holocene surface rupture are scarce (e.g., Lagerbäck, 1979; Adams, 1989; Grant, 1990; Adams et al., 1991; Fenton, 1994), the study of indirect paleoseismic objects, as well as other features or approaches (e.g., tsunami deposits, age determination of fault-scarp exposure and scarp-profile degradation studies, dendrochronology, lichenometry; Fig. 1B), has clearly established that these so-called stable regions are seismically active. Adams (1996b) and Mörner (2005) have found that unexpected contrasting seismic patterns at the time of deglaciation, characterized by large and very frequent earthquakes, also took place. In addition, the study of earthquake-induced effects (mass wasting and soft-sediment deformation in a more general way) is becoming a complementary tool to on-fault trench studies. In Venezuela, several periglacial (paleo-)lakes directly offset by the active Boconó fault have been cored and sampled in order to demonstrate the applicability of this approach, with the intention of extending its use to seismically active regions with no outcropping seismogenic faults (Carrillo, 2006; Carrillo et al., 2006a, 2006b, 2009). In Israel, very long earthquake histories have been reconstructed through the study of soft-sediment deformations in late Pleistocene to Holocene laminated lacustrine sequences of the Dead Sea region (Marco et al., 1996; Agnon et al., 2006). In addition to directly exposing the seismic evidence through trenching, these past seismically induced effects in the geologic record have been assessed through indirect methods, either invasive (coring and geotechnical methods; E in Fig. 1A and H in Fig. 1B) or noninvasive (geophysical methods). Fossil liquefaction features (particularly sand blows) in the subsurface have been uncovered by GPR (e.g., Liu and Li, 2001). Since these earthquake-induced effects may also occur underwater, their investigation under such conditions has mostly relied on the combination of geophysical methods and sediment core recovery (Figs. 1A and 1B). The overall geometry of the sedimentary bodies and their stratigraphy are obtained from the geophysical investigation, which can call upon methods such as side-scan sonar to provide highly detailed sea-bottom topography, high-resolution shallow seismic, including multi-beam, sparker, boomer, and pinger (subbottom profiler), among others, while coring reveals the physical characteristics of the sediments and deformations, and also brackets the time of occurrence of such effects (if dated). These underwater objects have been studied both at sea (Schafer and Smith, 1987; Piper et al., 1999) and in lakes (Beck et al., 1992; Shilts and Clague, 1992; Thomas et al., 1993; Beck et al., 1996; Chapron et al., 1996; Becker et al., 2005; Schnellmann et al., 2005; Carrillo, 2006; Carrillo et al., 2006a, 2006b, 2009; Beck, this volume), employing some of the aforementioned methods.

The timing of activation (or reactivation) of earthquake-related mass movements has been derived from trenching across the scarp or scar at the slide crown of such mass wasting (e.g., Wallace, 1984; Onida et al., 2001; McCalpin and Hart, 2002; Tibaldi et al., 2004; Gutiérrez et al., 2010). In this sense, large sackungen have received particular attention (B in Fig. 1A; e.g., Gutiérrez et al., 2005, 2008). From these studies, it is noteworthy that a significant uncertainty remains as to the completeness of the earthquake chronology derived by these methods. It still remains a difficult task to ascertain that a particular slide of a given size and certain characteristics will always be prone to slide during an earthquake of a particular magnitude or intensity above a certain threshold. It is equally difficult to assign a seismic origin to many sackungen in high mountain regions, especially if their occurrence is concurrent with deglaciation periods (Hippolyte et al., 2006). If the displacement of the mass movement is large enough and the mobilized mass does not completely fall down, the consequent depression generated at the foot of the scarp may create new space for the emplacement of small lakes. In such a case, the earthquake chronology can be assessed by the study of the perturbations recorded in the bottom sediments of these lakes (B in Fig. 1A; soft-sediment deformation, turbidites, [seiche-induced] homogenites, slumps, among others). If these lakes or ponds happen to be small in size, only coring can be performed. If a lake is large and deep enough, geophysical prospecting can be undertaken.

Some large mass movements, particularly rock avalanches, but also huge debris and mud flows, may have important run-off and may block the valleys by their downstream deposits. This may result in the damming of water bodies of considerable volume upstream. This process has been observed in several historical and contemporary earthquakes (e.g., Simón, 1627; Li et al., 1986; Clague and Evans, 2000). Consequently, the recognition of the aforementioned markers (soft-sediment deformation, turbidites, seiche-induced homogenites, slumps, among others) in lake-bottom deposits in slide-dammed lakes has also been used as a natural indicator of past earthquakes (Costa and Schuster, 1988; Weidinger, 1998; Hermanns et al., this volume). This same approach can be applied to water bodies that are pounded against a coseismically growing active fold (H in Fig. 1B). An example of this happened at Oued Foda during the 1980 El Asnam–Algeria earthquake, above an active blind thrust fault (Philip and Meghraoui, 1983). The analysis of such lake fills may help to reconstruct the seismic history of these faults that show no evidence of brittle expression at the surface as direct evidence. Geophysical methods, such as vertical electrical soundings (VES) and ERT, have been used jointly to calibrate multielectrode profiles and to estimate the depths of landslide deposits in the town of Celano (Fucino Basin, central Italy; Rinaldini et al., 2008).

Assuming that the quantum leaps forward in multichannel seismic acquisition technology are maintained in the years to come, we could easily envisage a situation in which much higher-resolution three-dimensional (3-D) seismics (seismic cube) would allow the regional-scale mapping of most of these seismically induced effects, particularly those paleoseismic objects that rest on a perturbed ground surface or seabed (e.g., sand blows, slides, but also mud volcanoes and pockmarks; Fig. 1). This modern seismic method currently allows the construction of time slices, although not yet of sufficiently high resolution. These constitute a reconstruction of the environment or topography at a given time and thus would fulfill the same objectives as current macroseismic surveys, from which magnitude and epicenter of the causative earthquake could be derived by applying known relationships. In other words, at each identified event horizon, corresponding to a time slice in the seismic cube, the paleoseismic objects would be mapped in as much detail as the resolution of the method would permit.

Those phenomena described here are known in the literature as seismically induced effects, but other, less well-known, natural features also form as a result of seismic shaking, among which we shall discuss the following: mud volcanoes, oil seeps and pockmarks, broken speleothems, and toppled precariously balanced rocks. Mud volcanoes/diapirs (F in Fig. 1A), and even salt domes, can be used as analogs for earthquake-triggered sand venting in the form of dikes connected to sand blows. Although it is known that mud diapirs may form and flow to the surface during earthquakes (e.g., Arnold et al., 1960), their use as paleoseismic indicators, to our knowledge, has not yet been reported in the literature. This can be attributed to the fact that they are not exclusively associated with earthquakes. Furthermore, many of these features do not have an episodic motion, but rather they have a steady-state ascending motion, which would make it more difficult to ascribe a seismic origin, although their upward motion and also final eruption are known to accelerate during earthquake shaking. The process of seabed degassing or gas seepage, which produces pockmarks as surface evidence (Hovland and Judd, 1988; Hovland et al., 2002; Pinet et al., 2008) is also known to accelerate during earthquakes (Fig. 1A). Pockmarks are shallow depressions (5–300 m in diameter and 2–20 m deep) formed by the explosive release of gas (Gluyas and Swarbrick, 2004). Seismic shaking can provoke unusually large flows from gas or oil seeps (Levorsen, 1967; Field and Jennings, 1987), similar to what happens with water springs that may resume or stop flowing as a consequence of earthquake action (e.g., González et al., 2004). For instance, following the 1971 San Fernando Valley (California) earthquake, several of the previously inactive oil seeps in the area resumed activity (Mandel, 2001). Unfortunately, there is no way that degassing evidence will be recorded in the geologic record, but pockmarks are perfectly well preserved on the seabed, as revealed by side-scan sonar. These features potentially provide the same paleoseismic information as earthquake-triggered sand blows. Underwater, they can be detected by high-resolution single-channel seismics (sparker or subbottom profiler) or multibeam surveys, whereas age bracketing can be obtained on coring. So far, it has not been applied for specific paleoseismic purposes.

Speleothems are also excellent recorders, as well as highly precise chronometers of seismic shaking (C in Fig. 1A). For over 20 yr, speleothems have been used as indicators of seismic activity (Postpischl et al., 1991; Bini et al., 1992; Gilli, 1996, 1999; Delaby, 2001; Pérez López et al., 2009). This has been stimulated by the fact that they are easily datable (by ^{14}C, for instance), they record growth perturbations very neatly, and they also grow by annual precipitation, similar to tree rings. Their principal limitation is that their occurrence is restricted to soluble carbonatic rocks. Anyhow, their steady way of growing permits detection of growth pattern disturbance as a result of catastrophic events with a temporal resolution of only 1 yr, similar to varve chronology and dendrochronology. Because of these characteristics, speleothems have been an object of many paleoseismic studies, as confirmed by the large number of studies in which historic (Gilli, 1999; Lemeille et al., 1999) and prehistoric (Postpischl et al., 1991; Delaby, 2001; Lignier and Desmet, 2002; Lacave et al., 2004; Kagan et al., 2005) earthquakes have been dated or constrained. Good overviews on the use of speleothems as paleoseismic indicators have been published by Forti (2001), Gilli and Delange (2001), and Becker et al. (2006). However, their interpretation is not straightforward, as demonstrated by Cadorin et al. (2001) and Gilli (2004). The latter author indicated that the breakage or failure of speleothems may also be caused by other processes different from seismic shaking, such as glacial advances or retreats. The possibility that sliding of the limestone masses containing the caves occurred as a result of gravitational processes also needs to be ruled out before ascribing a seismic origin, unless sliding can be assumed to have been triggered by a given earthquake.

Precarious rocks or precariously balanced rocks indicate that strong earthquake motions have not occurred at a site since their particular precarious situation developed. In that particular negative sense, these rocks or groups of rocks can be effectively used as earthquake strong ground-motion seismoscopes that have been operating on solid rock outcrops for thousands of years (Brune and Whitney, 1992; Brune, 1994, 1996, 1999; Shi et al., 1996; Brune et al., 2003). Thus, estimates of the threshold ground acceleration for the toppling of these rocks can provide constraints on the peak ground accelerations experienced during previous earthquakes. Consequently, the presence of precarious rocks is an indicator for the absence of earthquakes above a certain magnitude. Preservation of these rocks in their precarious position is thus a function of the seismicity of the region under study with a propensity for the larger events, and this reduces its applicability to high-seismicity regions of the world. On the other hand, toppled precarious rocks can be used as chronometers of the latest earthquake, the magnitude of which can be estimated as having exceeded the threshold value needed for toppling. The age of the place on which the precariously balanced rocks used to stand (commonly made of hard rock) could be determined by cosmogenic dating, thus providing time of occurrence of the latest event. Likewise, the age of the precarious rock toppling could be derived from dating the age of the ground surface on which the block fell (similar to J in Fig. 1B). If it fell on soil, the ^{14}C method of dating could be applied, whereas if the block has fallen on hard rock or on sediments, even luminescence (thermoluminescence or optically stimulated luminescence) techniques could be applied.

Remobilized and Redeposited Sediments

We propose to include in this type of indirect paleoseismic object all those deposits of any grain size that have been transiently mobilized by the earthquake or associated phenomena (e.g., underwater or subaerial mass wasting, tsunami waves) and redeposited shortly after. Usually, this type of deposit, which includes turbidites, homogenites, and tsunamites, is generated in aqueous environments but results from different processes or intervening mechanisms. Seismically induced turbidites and homogenites require that underwater (eventually, subaerial) sediments were set in motion by the earthquake shaking (e.g., Francis, 1971; Siegenthaler et al., 1987; Shiki et al., 2000). Turbidites are generally the result of submarine mass-wasting processes causing hyperpycnal inflow that generates turbidity currents, which transport sediments downslope along the seabed (e.g., Adams, 1990, 1996a; Doig, 1991; Inouchi et al., 1996; Gorsline et al., 2000) in the days or months following an earthquake, as evidenced in the case of several twentieth-century earthquakes (e.g., Heezen and Ewing, 1952; Thunell et al., 1999). Homogenites, instead, would rather require the oscillatory motion (seiche) of a confined water body, although this motion is also fed from underwater mass wasting (e.g., slumps; e.g., Beck, this volume) or nearshore subaerial mass movements (e.g., collapse of the edifice of the Santorini Volcano as proposed by Cita et al., 1996). Conversely, most tsunamites are fed from nearshore sediments that are carried inland by tsunami waves (e.g., Atwater, 1987; Clague and Bobrowsky, 1994a, 1994b; Clague et al., 1994; Chagué-Goff and Goff, 1999; Bourgeois et al., 1999; Sawai et al., 2008; among many others). This does not exclude the situation where tsunamites can also appear as homogenites in deep and/or confined water bodies (e.g., Kastens and Cita, 1981; Beck et al., 2007). Typically, tsunamites are thought to be fine- to medium-grained deposits (e.g., Dawson et al., 1988; Shi et al., 1995; Paris et al., 2007), but in fact, the size of the transported material is a function of the tsunami wave energy and availability of materials in the nearshore environment in the case of onshore tsunamites (for a review, refer to Scheffers and Kelletat, 2003), or in the sea bottom for deep-water tsunamites (homogenites). Tsunami waves have been capable of removing and transporting large boulders inland (e.g., Bourrouilh-Le Jan and Talandier, 1985; Jones and Hunter, 1992) and huge (several m^3 in size) blocks of coral reefs, eolianites, or other nearshore deposits (Scheffers, 2004; Scheffers and Kelletat, 2005; Scheffers and Scheffers, 2007).

Onshore tsunamites can be assessed in outcrop or by shallow coring (A in Fig. 1A). The coast of the Cascadia subduction zone of the northwestern United States has been a natural laboratory for over the past 40 yr, during which their study has much evolved (Atwater et al., 1995). In the early days, recognition of the tsunamites in that area relied mostly on sedimentary features recognized with the naked eye. The identification of these marine incursions nowadays relies on the combination of several disciplines: sedimentology, geochemistry, paleontology, including micropaleontology, malacology, palynology, and geochronology, among others, which has enhanced and facilitated their recognition in the geologic record. Since all other paleoseismic objects of this sort are mostly preserved in still-active natural environments, particularly in lake- or sea-bottom sequences, they require an integrated paleoseismic assessment relying on joint subaqueous geophysical prospecting methods and piston coring (e.g., Beck et al., 1996; Cita and Rimoldi, 1997; Doig, 1998; Carrillo et al., 2006b; Beck et al., 2007; Beck, this volume). However, paleoseismic indicators of this subset have also been retrieved from outcropping lake sequences (e.g., Doig, 1991; Carrillo et al., 2006a), which unfortunately have been truncated and as a consequence provide only incomplete paleoseismic records for the recent time window.

Paleoseismic Indicators of Vertical Motion

Most of the indirect paleoseismic objects discussed here constitute an invaluable help when direct evidence of coseismic faulting is unavailable. In other words, their recognition has been greatly enhanced by their study in subduction-related seismogenic source regions and occasionally also in association with blind thrusts. Earlier, we explained how elastic rebound acts coseismically in subduction zones and can be recognized by the presence of the so-called ghost or drowned forests after the occurrence of large subduction earthquakes in the coastal zone, where it is clearly recorded in the onshore sediments by a couplet of organic soils capped by a horizon of marine sediments (A in Fig. 1A). In addition, it can be perfectly recorded by marine biogenic constructions such as coral reefs and (micro-)atolls on the foreshore side of the coastal bulge (Fig. 1B). In inverse relation to the onshore sediment couplet, which registers emergence during the interseismic period followed by subsidence during the earthquake, the coral growth offshore first indicates steady subsidence during the interseismic period followed by abrupt uplift during the seismic energy release (Taylor et al., 1987; Sieh et al., 1999; Zachariasen et al., 1999, 2000). This sudden uplift, when of sufficient magnitude, can result in aerial exposure and subsequent coral head truncation by wave erosion or natural death of the coral species. Similar to tree rings and varves in temperate regions, growth of corals indicates seasonality in tropical regions, and coral growth is registered in annual bands (e.g., Knutson et al., 1972; Woodroffe and McLean, 1990), but their occurrence is of course limited to the warm regions of the world. Furthermore, certain species of the coral genera *Porites* and *Goniastrea* are sensitive natural recorders of lowest tide levels, which make them ideal natural instruments for measuring emergence or subsidence relative to a tidal datum (Scoffin and Stoddart, 1978; Taylor et al., 1987). These *Porites* and *Goniastrea* coral heads grow radially upward and outward until they reach an elevation that allows their highest corallites to be exposed to the atmosphere during lowest tides. Consequently, rather small changes in elevation of the coral substratum or sea level induce very clear changes in the growth patterns of these coral colonies or micro-atolls (Zachariasen et al., 2000; Natawidjaja et al., 2007). Thanks to the excellent temporal and vertical resolution of these micro-atolls, it has been possible to decipher the seismic history of particular segments of subduction zones in tropical regions (e.g., Sieh et al., 2008).

In the past 30 yr or so, the development of new, modern, and more precise dating techniques such as uranium series, thermoluminescence (TL), optically stimulated luminescence (OSL), and electron spin resonance (ESR), among others, in combination with a more accurate reconstruction of the history of Quaternary sea levels based on variations in the oxygen isotopic composition of seawater (Chappell, 1974; Chappell et al., 1996a), has stimulated and accelerated the study of both emerged and submerged coastal marine features in tectonically active regions worldwide, as well as improved their temporal resolution (Fig. 1B). This has led to an improved determination of vertical slip rates based on marine terraces, either constructional (biogenic or detrital) or erosional. Many appropriate areas have been studied in that respect, and some regions of the world have actually become natural laboratories, such as Huon Peninsula in Papua New Guinea, where the matching of these marine terraces to sea-level curves (Chappell, 1974; Chappell et al., 1996a) has allowed the recognition of other marine terraces, the origin of which has been ascribed to coseismic vertical tectonic motion (Ota et al., 1993; Chappell et al., 1996b; Ota and Chappell, 1996). Other regions of the world have been assessed likewise, such as Chile by Vita-Finzi (1996), New Zealand by Pillans and Huber (1995), and the Gibraltar Strait zone by Zazo et al. (1999). The Ventura Avenue anticline in California deserves particular mention in this respect, where Stein and Yeats (1989) established that this active fold is repeatedly raised by earthquakes recurring roughly every 600 yr, as registered in a series of coseismically uplifted marine terraces.

The equivalent of marine terraces in coastal areas, alluvial terraces of the inland environment have been occasionally assessed for the purpose of determining coseismic vertical or horizontal deformations (e.g., Wellman, 1955; Suggate, 1960; Lensen, 1968; Lensen and Vella, 1971; Van Dissen et al., 1992) in continental interiors. However, the most notable contribution of these flights of staircase-style marine, lake, or alluvial terraces is in permitting us to arrive at reliable estimates of the vertical slip rate of tectonic blocks or folds that have been uplifted or subsided.

CONCLUDING REMARKS

Over the past 40 yr, there has been a steady increase in the number of natural objects (generic name herein given to geologic and geomorphologic evidence, indices, markers, effects, indicators, etc.), both onshore and offshore on the seabed, that have been identified as being of distinct relevance to paleoseismology. This increase in the number and variability of natural objects has been in response to the need for an improved and wider-ranging characterization of the seismic potential of inaccessible seismogenic sources, such as subduction zones and intracontinental seismic sources (e.g., New Madrid seismic zone). This has led to the identification of numerous indirect indicators and to the understanding of their functioning. Consequently, the appropriate paleoseismic approach is intimately related to the seismic landscape under assessment and is defined as the cumulative geomorphological and stratigraphic effect of the signs left on the environment of an area by its past earthquakes over a geologically recent time interval (Vittori et al., 1991; Michetti et al., 2005b). Likewise, the use of other natural paleoseismic markers has partly decreased, been abandoned, or fallen into disuse because of their limited applicability or limited results in terms of quantifying the seismic potential, for any or several of the following reasons: (1) The occurrence of the object is restricted to a specific region, a climate, or process, like lake varves, tree rings, or coral bands. (2) There is high specificity of the object to local conditions, e.g., scarp degradation to highly variable diffusion equations, lichenometry. (3) There is limited or unlikely preservation of the object in the geologic record, such as dead trees for dendrochronology. (4) There is difficulty in determining the magnitude and location of past earthquakes based on the regional distribution of the natural objects, which results in great uncertainty in assigning the observed seismically induced effects to a single common earthquake. (5) There is a degree of uncertainty in the cause-effect association between the natural indicator and earthquakes, for instance, liquefaction or mass movements. In the ultimate instance, the final selection of the natural indicators available for use is the responsibility of the paleoseismologist in accordance with those objects that each environment offers in terms of natural features. Finally, if available, direct objects are certainly everyone's preference because they provide straightforward answers to the seismic history and potential for future rupture of a given fault or fault segment.

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