The Use of Soft-Sediment Deformation Structures As Proxies For Paleoseismic Activity And Shaking: A Review

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Abstract

Quantifying the magnitude of an earthquake is very important for long-term and medium-term earthquake prediction, post-earthquake emergency rescue and seismic hazard assessment. Paleoseismology is the investigation of past earthquakes in the geological record, in particular their location, timing and size. Uncertainties remain in the paleoearthquake magnitudes determined by traditional surface rupture parameters, especially because most seismic events do not result in surface ruptures. In order to address the problem of magnitude evaluation of earthquakes that did not reveal major dislocations, this paper deals with the methods used to determine the seismic shaking intensity based on the types and forms of soft-sediment deformation structures, including maximum liquefaction distance, thickness of disturbed layer, empirical formulae, and thickness of rapidly deposited sand layer. Then we discuss and analyze these methods in terms of their theoretical basis, advantages and disadvantages, accuracy, applicability and problems. We chose two case studies: first, a typical seismics-related deposit (liquefied layer and disrupted layer) represented by a seismite in the late-Pleistocene Lake Lisan section near Masada in the Dead Sea Basin; and second, the liquefied diapir triggered by an earthquake in the late-Quaternary lacustrine sediments at Luobozhai in the upper reaches of the Minjiang River, east Tibet. The six methods listed above are employed to determine earthquake magnitudes associated with the seismics-related deposit and liquefied diapir, yielding magnitudes of 5.5-6.5 and 6-7, respectively. The combination of the six methods, provided a new and relatively convenient method for determining seismic shaking, especially in lacustrine sediments. This study can serve as a valid reference for comparing methods of calculating the magnitude of a paleoearthquake based on surface rupture parameters, and provides a better understanding of the long-term seismic activity and risk in tectonically active regions.

Introduction

Strong earthquakes cause primary and secondary effects on the natural environments. These have been codified in an Environmental Seismic Intensity (ESI) Scale 2007, which can be used together with macroseismic scales or alone, in case of inhabited regions. This is particularly valuable for strong earthquakes ($M \geq 7.5$), such as the 1999 M_S 7.6 Chi-chi (Taiwan) earthquake (Chang et al., 2012), 2004 M_S 9.0 Sumatra-Andaman (Indonesia) earthquake (Lay et al., 2005), 2008 M_S 8.0 Wenchuan (China) earthquake (Liu-Zeng et al., 2012), 2011 M_S 9.0 Tokyo (Japan) earthquake (Sasaki et al., 2012), and 2015 M_S 8.1 Nepal earthquake (Yang et al., 2016) (Fig. 1). Determining the magnitude for ancient earthquakes is very important for long-term and medium-term seismic hazard assessment (Galli and Ferrelì, 1995).

Paleoseismology is the investigation of past earthquakes in the geological record, especially their locations, timing and magnitudes, and focuses mainly on surface ruptures (Sieh, 1978). Nevertheless, it also benefits from the investigation of liquefaction and instantaneous deformation of landform studies. The paleoliquefaction approach in the paleoseismology focuses on soft-sediment deformation structures (SSDS) and ground failures resulting from earthquake-related liquefaction (e.g. McCalpin and Nelson, 1996; Tuttle et al., 2019). Usually, the magnitudes of paleoearthquakes are determined by empirical relationships based on the surface rupture lengths, and maximum and average displacements per event (e.g. Wells and Coppersmith, 1994; Stirling et al., 2013). Although these methods have received much attention and made great progresses, there are three key deficiencies still heatedly debated. First is the inability to infer the earthquake existence when fault displacements do not reach the surface (Ran and Deng, 1999). For example, emergency surveys after the 2013 M_S 7.0 Lushan (Chen et al., 2014), 2015 M_S 6.4 Pishan (Li et al., 2016), 2015 M_S 8.1 Nepal (Liu et al., 2015) and 2017 M_S 7.0 Jiuzhaigou (Xu et al., 2017) earthquakes yielded no obvious co-seismic surface ruptures, yet, these earthquakes produced a significant number of tensional ground fissures, landslides, bedrock collapses and liquefaction, in addition to the damage to buildings. What's more, only few paleoearthquake events can be identified though the trench method, and the time scale is relatively short and discontinuous (Zhong, 2017). Thirdly, much controversy remains regarding which is the magnitude threshold for ruptures reaching the surface ground. Tang et al. (2015) analyzed 56 moderate-size earthquakes worldwide occurred in the period from 1950 to 2014 AD, which caused clearly documented surface ruptures. Their analysis showed that the lowest limit of earthquake magnitude associated with surface rupture was below 6.5, most likely about 5, and even as low as 3.6 under extreme conditions (e.g. the 1966 Lompoc earthquake, with a focal depth of only 1.1 km). Therefore, establishing a complementary method for evaluating the magnitude of paleoearthquakes remains an issue in paleoseismology.

In addition to ruptures, earthquakes can cause liquefaction and/or fluidization of granular solids, if their magnitude is sufficiently high ($M \geq 4.2$; Galli, 2000); this is reflected in layers characterized by abundant SSDS (seismites, *sensu* Seilacher, 1969) also over long distance (van Loon et al., 2016). Lowe (1975) supposed the liquefaction is marked by the sudden breakdown of a metastable, loosely packed grain framework, the grains becoming temporarily suspended in the pore fluid and settling rapidly through the fluid until a grain-supported structure is re-established, and fluidization occurs when the drag exerted by moving pore fluids exceeds the effective weight of the grains; the particles are lifted, the grain framework destroyed, and the sediment strength reduced to nearly zero. In tectonically active regions, SSDS provide evidence for the occurrence of earthquakes (e.g., Sims, 1973; Hempton and Dewey, 1983; Allen, 1986; Galadini et al., 1995; Marco and Agnon, 1995; Moretti et al., 1999; Neuwerth et al., 2006; Moretti and Sabato, 2007; Suter et al., 2011; van Loon and Pisarska-Jamroz, 2014; Jiang et al., 2016; van Loon et al., 2016; Liang et al., 2018; Salomon et al., 2018; Zhong et al., 2019). As early as forty years ago, researchers had started to analyze and discuss the relationship between SSDS and earthquake magnitudes. Commonly used methods and parameters derived from the SSDS include the thickness of disturbed beds (Hibsch et al., 1997; Rodríguez-Pascua et al., 2003), the maximum liquefaction distance (Kuribayashi and Tatsuoka, 1975; Liu and Xie, 1984; Ambraseys, 1988; Galli and Ferrelì, 1995; Papadopoulos and Lefkopoulos, 1993; Obermeier, 1996; Galli, 2000; Obermeier et al., 2002; Maurer et al., 2015; Qiao et al., 2017), the types of SSDS (Rodríguez-Pascua et al., 2000; Neuwerth et al., 2006; Berra and Felletti, 2011), and the thickness of rapidly deposited sand layers (Moernaut et al., 2014). However, these methods have some disadvantages, as follows. (1) Besides earthquakes, many processes can produce SSDS, such as gravity slides (Owen and Moretti, 2008; van Loon, 2009); overloading (Moretti and Sabato, 2007; Owen and Moretti, 2008), rapid sedimentation (Lowe, 1975; Postma, 1983), channel erosion (Dasgupta, 1998), storm waves (Molina et al., 1998; Alfaro et al., 2002), and ice over-riding permafrost (Anketell and Dzulynski, 1970; Vandenberghe, 1992; Zhong et al., 2020a). Consequently, it is necessary to determine the cause of each deformation feature. However, liquefaction can be induced by 21 triggers, earthquake only represents the one of

these triggers (Shanmugam, 2016). What's more, considering that the deformation mechanism of SSDS is complicated, there are still uncertainties in the interpretation of its genesis and trigger mechanism, especially the fluvial and lacustrine deposits with complex sedimentary processes (Zhong et al., 2020b). Therefore, it is very important to determine the origin and trigger mechanism of SSDS. (2) The nature of the driving force, the sediment rheology, deformation mechanism and timing of deformation relative to sedimentation, can affect the final morphology and deformation style of SSDS (Owen, 1987). (3) The epicenter distance, depth of hypocenter, and intensity and duration of seismic shaking can also affect the final morphology and types of SSDS (Obermeier, 1996). (4) Determination of the seismic shaking using SSDS also depends on the regional geology, tectonic setting, sedimentary environment, sedimentary age, and in particular the grain-size distribution, water content and consolidation degree of the sediments (Owen and Moretti, 2011; Moretti et al., 2016; Zhong et al., 2017). Therefore, the relationships between the types and forms of SSDS and earthquake magnitudes need further researches. Here, we summarize and analyze the research methods employed in the SSDS investigations, including their theoretical basis, advantages and disadvantages, accuracy, applicability and problems.

This paper aims to: (1) summarize the methods for determining seismic shaking intensities using SSDS; (2) analyze the applicability, advantages and disadvantages of each method; and (3) develop a new method for determining shaking intensities. The results of our work can provide a better understanding of the regional seismic risk.

Previous Studies And Proposed Methods

Determining earthquake magnitude or seismic shaking based on SSDS includes observations of the types and forms of SSDS, maximum liquefaction distance, thickness of a disturbed layer, empirical formulae, and thickness of a rapidly-deposited sand layer (Table.1). Among these, the most commonly used methods are the maximum liquefaction distance (Qiao et al., 2017) and the deformation type (Rodríguez-Pascua et al., 2000). Historical liquefaction method (Sims, 1975) and the thickness of the disturbed layer (Hibsch et al., 1997) are based on the liquefaction deformation intensity and its relationship with the deformation amplitude of the SSDS layer and the seismic shaking, respectively. These methods only provide the lower limit of seismic shaking, and are used less frequently. The empirical formula (Rodríguez-Pascua et al., 2003) and the thickness of rapidly-deposited sand layer (Moernaut et al., 2014) methods are based on the thickness of the SSDS layer and its relationship with the thickness of the overlying sand layer and the earthquake magnitude, respectively. Due to the complexity of SSDS deformation mechanisms and lacustrine sedimentation processes, this method is still under investigation. The method also has strict requirements on the thickness of the deformed layer (≤ 15 cm) and thickness of the rapidly-deposited sand layer (≤ 50 cm) (Zhong et al., 2020c), which limits the scope of its use. In the following paragraphs, we will analyze the applicability, and advantages and disadvantages of each method.

Historical liquefaction cases for determining lower boundary curve for liquefaction occurrence

Controversy remains regarding the lowest limit of earthquake magnitude that can cause liquefaction. Some researchers initially believed that a M_s 2–3 earthquake was large enough to trigger liquefaction (e.g., Seed and Idriss, 1971). Later, Scott and Price (1988) proposed that earthquakes with magnitude smaller than 5 do not cause significant liquefaction of sediments. Valera et al. (1994) suggested that the threshold magnitude to induce liquefaction effects in the most susceptible gravel deposits is about 7 (e.g. in the 2008 Wenchuan M_s 8.0 earthquake, the liquefaction of gravelly soils accounted for 80-85% of all liquefaction points, see Cao et al., 2010), whereas the threshold magnitude for sands is about 5.5. Marco and Agnon (1995) proposed that the triggering of liquefaction occurs only for earthquake magnitudes larger than 4.5; these values are consistent with the historical case studies (Kuribayashi and Tatsuoka, 1975; Youd, 1977; Galli, 2000). Many researchers (e.g. Allen, 1986; Audemard and De Santis, 1991; van Loon et al., 2016) have suggested that Richter magnitudes larger than 5 are required to produce significant liquefaction effects in near-surface, water-saturated, semi-consolidated to unconsolidated sediments. The duration of ground shaking during seismic events of smaller magnitude ($M < 5$) is often insufficient to cause liquefaction.

An MMI (Modified Mercalli Intensity Scale) of about VI is the threshold for widespread development of small-scale SSDS features such as folds, pseudonodules, contorted laminations and recumbent folds (Sims, 1975). Although liquefaction effects have occurred at MMI values as low as V and VI (Galli, 2000 reports I to 5-6 MCS, Mercalli Cancani Sieberg scale), Keefer (1984) suggested that the lowest shaking at which liquefaction-induced features can become common is VII. Monecke et al. (2004) found that lake sediments are only affected if they are situated within an area that underwent ground shaking of at least shaking VI to VII. Small-scale SSDS, such as disturbed and contorted laminations as well as liquefaction structures (e.g. dish structures), are generated by earthquakes when $M = 5-5.5$ (Monecke et al., 2004, 2006). When an earthquake causes liquefaction and/or fluidization of water-saturated sediments, the magnitude could be $M \geq 5$ or the MMI greater than VI. The advantage of the historical liquefaction method for assessing seismic shaking is its convenience, speed, and wide range of applicability. The disadvantages are that the accuracy is limited and it only provides the lower limit of seismic shaking. Therefore, historical liquefaction method can be used only as a preliminary seismic shaking estimation method.

Maximum liquefaction distance for determining earthquake magnitude

The sand or soil liquefaction and water ejection features caused by earthquakes show a clear correlation with the magnitude and epicenter distance (Kuribayashi and Tatsuoka, 1975; Qiao et al., 2017). The larger the earthquake magnitude, the wider the distribution of liquefied sites (Papadopoulos and Lefkopoulos, 1993). Paleoequake magnitudes have been calculated based on sand-layer liquefaction data for modern and historical earthquakes (Qiao and Guo, 2013). The key is first to establish the location of the epicenter, and then to determine the location of the farthest liquefaction from the epicenter (Qiao et al., 2017). This requires identification of the location of liquefaction deformation that is the furthest from the seismogenic fault that induced the

earthquake, when undertaking geological field surveys and paleoearthquake research. Kuribayashi and Tatsuoka (1975) have been commended for their comprehensive compilation of the occurrence and distribution of liquefaction during Japanese earthquakes of the past century (1872–1968), and they produced the first map revealing the relationship between the maximum epicentral distance of liquefied sites (R) and magnitude (M) for at least 44 earthquakes ($M > 5.3$) and liquefied sites (Fig. 2). Youd (1977) compiled data for 14 earthquakes ($M > 5.3$) from 1897 to 1976 in the USA, India, New Zealand and Chile; their data support the conclusion of Kuribayashi and Tatsuoka (1975) that there is a maximum distance beyond which liquefaction is not likely to occur for an earthquake of a given magnitude. Liu and Xie (1984) compiled sand-bowls and sand-emission data for the 900 years before 1995, and established a relationship between magnitude and maximum epicentral distance to liquefied deformation in China. Ambraseys (1988) re-examined the 44 events analyzed by Kuribayashi and Tatsuoka (1975) and eliminated 5 cases, then added the further 14 cases reported by Youd (1977) from other parts of the world, 6 by Davis and Berrill (1983) and 7 by Fairless and Berrill (1984), as well as 70 new cases; in total 137 earthquakes known to have caused ground failures due to liquefaction were analyzed. Ambraseys (1988) then established a relationship between maximum epicentral distance of liquefied sites R (km) and moment magnitude M . Galli (2000) updated the Italian catalog of liquefaction (Galli and Meloni, 1993; Galli and Ferrel, 1995) by means of new systematic historical researches. This database, permitted the construction of empirical relationships between the epicentral parameters of the earthquake (I_0 , M_s , and M_w) and the distance of the observed liquefaction for 317 cases related to 61 different earthquakes in Italy since 1117. The triggering events range from intensity (MCS) 5.5 to 11, from magnitude (M_s) 4.2 to 7.5, and magnitude (M_w) 4.83 to 7.46. Papadopoulos and Lefkopoulou (1993) propose a slight modification of the M/R and M/R relations suggested by Ambraseys (1988). Subsequent revision and addition of further data by Obermeier (1996) and Obermeier et al. (2002) has yielded other widely-used diagram showing the relationship between SSDS and earthquake magnitude. Qiao et al. (2017) produced a detailed plot of maximum liquefaction distance and earthquake magnitude (Fig. 3) which shows an obvious relationship between the earthquake epicenter and the farthest surface liquefaction. We can use the R - M map to estimate earthquake magnitude based on the largest distance between the observed liquefaction points and the earthquake epicenter (or responsible fault) in the field.

Conversely, in the field, it is often difficult to determine whether the observed liquefaction deformation may be the farthest from the epicenter. Nevertheless, the R - M map (Fig. 3) can still be used to estimate the magnitude of the earthquake, because it gives a reasonable range of paleoearthquakes magnitudes triggered by active faults that may have formed SSDS, and once the responsible fault has been identified, the R - M map (Fig. 3) can be applied. The epicenters of paleoearthquake should be distributed along the fault, similarly to epicenters of modern earthquakes. When the epicenter of a paleoearthquake is unknown, we can also project the vertical distance from the furthest liquefaction deformation record (or the location of the liquefied deformation section) to the seismogenic fault onto the R - M map; this provides an estimate of the range of the paleoearthquake magnitude. We note that the R - M map only can provide a quasi-quantitative paleoearthquake magnitude, and cannot be as accurate as modern earthquake magnitudes obtained by seismic wave measurements.

Thickness of disturbed layers for determining seismic shaking

Some attempts (e.g. Sims, 1975) have been made to establish a relationship between seismic shaking and thickness levels, based on SSDS. Guiraud and Plaziat (1993) measured the degree of deformation in seismites observed within the Bima Sandstone Formation of the Lower Cretaceous Beune Basin, Northeastern Nigeria. Hibschi et al. (1997) compared the distribution of coseismic deformation thicknesses (disturbed or contorted layer) and the distribution of historical seismic shaking in the Holocene fluvial-lacustrine sediments of the northern Quito Basin, leading to the proposal of a new method of paleoseismic analysis which achieved relatively good for seismic shaking over VI. Fig. 4 shows the historical seismicity of Quito and attribution of seismic shaking (MMI) to the deformation levels. Considering that the sedimentological characteristics have a greater control than the layers thickness, then for layers with a similar lithology (grain-size, sorting, petrology, compaction, etc.), a stronger seismic shaking should result in a thicker disturbed layers. Disturbed layers thicknesses of 0–8 cm and 27–53 cm could tentatively correspond to seismic intensities of VI and XI, respectively. It is important to note that the disturbed layers occurred at the water/sediment interface, and are identified by the following features: constant thickness of contorted and disrupted layers not related to slumps; horizontal post-earthquake deposits showing the same lithology as that of the deformed horizon (Hibschi et al., 1997). Considering the paucity of the statistical data, and the lack of experimental validation, the accuracy of the disturbed layers thickness methodology is limited. In addition, this method does not consider the effect of the sedimentary layer on seismic waves (amplitude and travel time) (Li et al., 1991). In a specific succession, earthquake-induced seismic waves will deform the same layer (or set of layers), possibly independent of their magnitude. A strong earthquake may deform only a thin layer, and a weak earthquake may deform a thick unit, which need further research.

SSDS can evolve in one liquefiable layer (silt), or concurrently in two (silt and sand), or even in three layers (silt, sand and gravel), as a function of the energy involved (Rodríguez-Pascua et al., 2010). Rodríguez-Pascua et al. (2010) proposed a new methodology (DDI, degrees of intensity deformation) to estimate the intensity of SSDS. Through comparative analysis of multiple SSDS sections in the Tierras Blancas Basin, Rodríguez-Pascua et al. (2010) established a regional paleoseismic isoline by the DDI method, and inferred the seismogenic source of this paleoearthquake. This method can therefore benefit regional seismic hazard assessment. Based on the Kelvin-Helmholtz instability mechanism, Lu et al. (2020) modeling the relationship between soft deposition deformation structure and peak ground acceleration (PGA), and constrain the magnitudes of the six paleoearthquakes (seismites) with intensities of VII (0.18g), VI (0.13g), VI (0.09g), VII (0.18g), VI (0.09g), and VII (0.18g) as M_w 5.6, M_w 6.1, M_w 6.2, M_w 7.1, M_w 6.0, and M_w 6.9, respectively, which are in line with recorded historic magnitudes.

Empirical formulae for determining earthquake magnitude

Rodríguez-Pascua et al. (2003) found that the frequency-thickness distribution of 73 seismites obeys an exponential law, and that the “ b ” value of this distribution is the same as the “ b ” value for the seismicity of the area (b is defined by the Gutenberg–Richter law, and reflects the proportional relationship

between large and small earthquakes). In addition, they compared the deformation caused by modern and historical earthquakes in Spain with the deformations caused by paleoearthquakes, and found very similar characteristics. Then they established an empirical relationship between the seismics-related deposit (liquefied layer and disrupted layer) thickness of the studied sediments and the earthquake magnitude, yielding the empirical formula $M = T/3 + 3.83$, where T (in cm) is the thickness of the seismics-related deposit and M is the magnitude. Three zones characterized by distinctive styles of deformation can be distinguished in a fully-developed seismics-related deposit: an uppermost fluidization zone, an intermediate zone of ductile–fragile deformation (breakage and fragmentation of lamina), and a lower zone showing features typical of ductile deformation (folding of lamina). This lower zone overlies undeformed laminae (Rodríguez-Pascua et al., 2003).

This characteristic seismics-related deposit structure was first defined by Marco et al. (1994) for laminated sediments of the Pleistocene Lisan Formation in the Dead Sea region (Middle East). Marco and Agnon (1995) suggest that the seismics-related deposit are composed of fine-grained matrix-supported texture and tabular laminated fragments, underlain by asymmetrical and recumbent folds, and overlain by undeformed sand layers. Marco and Agnon (2005) changed the term “seismics-related deposit” to “breccia-layers”, and interpreted them as seismites. Undisturbed laminated layers between these breccia layers represent inter-seismic intervals. To separate observations from interpretations, “intraclast” refers to the clasts that were reworked from within the sedimentary section; Agnon et al. (2006) replaced the term “breccia-layers” with “intraclast breccia layers” and stressed that the term “breccia” refers to the texture of the deposit. In addition, they considered the formation of intraclast breccias, proposing that earthquakes are the main mechanism. The Dead Sea intraclast breccias have many similarities to seismites described in a lacustrine environment (Zhong, 2017), as well as glacial deposits (Davenport and Ringrose 1987). Based on field observations and research experience, we propose that the seismics-related deposit (named breccia layer by Agnon et al., 2006) should comprise a liquefied or fluidized layer and overlying detrital layer, but should not include the overlying undeformed sand layer with a grain size change in vertical section.

No clear standard has been defined for the thickness of the seismics-related deposit (Rodríguez-Pascua et al., 2003), thus limiting its application. Many empirical data show that the magnitude is not strictly linear with the logarithm of earthquake frequency (Fu, 1997), and a robust error estimation and confidence interval calculation is needed for “ b ” values. Mugnier et al. (2011) found that the lithology, grain size, sedimentary environment and acceleration can all influence the thickness of seismites, and also the number of individual shocks and number of cycles (earthquake recurrence cycle) can affect the final thickness of the deformed layer. Meanwhile, numerical simulations indicate that, firstly, deformation intensity increases with increasing ground acceleration and decreasing layer thickness; and secondly, the duration of the earthquake affects the deformation geometry, rather than the deformation amplitude (Wetzler et al., 2010). Therefore, the relationship between coseismic shaking and thickness of the seismics-related deposit is very complex and is not single-valued (Mugnier et al., 2011); as such, one should be very careful in applying the empirical formula.

Types of SSDS for determining earthquake magnitude

In water-saturated and unconsolidated sediments, different seismic shaking and deformation mechanisms (liquefaction, fluidization, thixotropy etc.) can result in different types of deformation structure. For example, gravity and liquefaction can gradually form load, flame, ball-and-pillow and pillow-bed structures. Firstly, the superposition of coarse-grained sand (heavy) on fine-grained sand (light) is analogous to gravitational instability (Anketell et al., 1970), allowing coarse-grained sediment to sink into fine-grained sediment when the fine-grained (underlying) material liquefies, resulting in load casts. Secondly, as a load cast develops, the underlying, finer sediment compensates its loss of space with upward intrusions forming flame structures. Thirdly, at the base of the source layer, horizontal motion generates horizontal shear stress in the over-pressurized fine-grained layer, and causes the laminae to convolute within the clay to form contorted layer (Suter et al., 2011). Fourthly, with stronger and/or longer seismic shaking, the heavy sand completely sinks into the light sand, forming pseudo-nodules (Anketell et al., 1970) and ball-and-pillow structures (Owen, 1987), and eventually forming pillow-bed structures (Qiao and Li., 2008). Sims (1973) suggested that the evolution of the load cast phenomenon towards pseudonodule formation may be an indication of stronger and/or longer seismic shaking. Under identical conditions (i.e. in the same sedimentary environments), the seismic shaking required to create ball-and-pillow structures should be greater than that required to create load cast and flame structures, while the shaking needed to create pillow beds should be larger than that needed to generate the ball-and-pillow structures (Owen, 1987).

Rodríguez-Pascua et al. (2000) summarized the various types of SSDS and their relationships with earthquake magnitudes, suggesting that seismics-related deposit, pseudonodules and ball-and-pillow structures correspond to magnitudes of 5.5–6.5, 6.5–8 and 6–8, respectively (Fig.5). Taking into account the distances to seismogenic faults, Neuwerth et al. (2006) tentatively postulated that the earthquake magnitude required to generate SSDS is 5–7. Berra and Felletti (2011) concluded that an earthquake should have a Richter magnitude of between 6 and 8 if it is to be capable of generating ball-and-pillow structures or irregular, contorted stratification. Qiao and Guo (2013) used the relationship between earthquake magnitude and the maximum distance from epicenter established by Liu and Xie (1984) and Obermeier (1996), and suggested that liquefaction or fluidization deformation (droplets and a homogenized layer) in early Jurassic lacustrine sandstones in the Wujia region of the SW Tianshan Mountains may have recorded paleoearthquake magnitudes ranging from M 6.5 to 7. In addition, a large number of load and ball-and-pillow structures have been reported from a liquefied deformed layer with the same age (Qiao and Guo, 2013). Zhong et al. (2017) proposed that the load and ball-and-pillow structures were triggered by earthquakes with magnitudes of 6–7. It is possible to obtain the seismic shaking through comparative analysis of similar types of SSDS triggered by earthquakes in the same sedimentary environments with similar lithology.

Thickness of rapidly deposited sand layer for determining earthquake magnitude

The M_s 8.0 Wenchuan earthquake struck on 12 May 2008, triggering more than 190,000 landslides and collapses along ridges and hillslopes of the Longmen Shan (Sichuan, China) (Xu et al., 2014). These detrital materials still remain in the hillslopes and valleys of the earthquake area. The transfer of a

large amount of landslide material from the hillslopes to the river network resulted in a large number of dams and associated lakes that formed immediately after the Wenchuan earthquake (Fan et al., 2012). Wang et al. (2015) used suspended sediment (<0.25 mm) concentration measurements from several rivers (Minjiang River, Fujiang River and Tuojiang River) in the epicentral area to analyze the evolution of suspended sediment flux; their results showed suspended sediment discharge increased by factors of 3-7 after the Wenchuan earthquake relative to 2006-2007 levels. Furthermore, ^{10}Be mixing budgets indicate that the sediment flux of the 0.25–1 mm size fraction increased by up to six-times following the Wenchuan earthquake (Wang et al., 2017). Across the Longmen Shan, at the present rate of post-earthquake fluvial export, Wang et al. (2015) estimate that it will take 33 ± 24 yr to remove all material <0.25 mm delivered by coseismic landslides, and coarse sediment (> 0.25 mm) will take about 1000 yr. The entrance of the Zipingpu reservoir at the highest water level is ~ 2 km downstream from the epicenter of the 2008 Wenchuan earthquake, which can capture a significant portion of the landslide material associated with the 2008 Wenchuan earthquake. The deposition record in the Zipingpu reservoir core shows the grain size and magnetic susceptibility have an abrupt coarsening (increasing) and upward fining (decreasing) after the 2008 Wenchuan earthquake (Zhang et al., 2019). However, these changes (grain size and magnetic susceptibility) were delayed until 2 years after the event. Based on this delay, Zhang et al. (2019) emphasized the importance of the interplay between a major earthquake and the prevailing monsoonal climate, highlighting the central role of runoff as an erosional agent in removing earthquake-triggered landslide material and in creating a corresponding depositional signal of the earthquake. After the 2008 Wenchuan earthquake, the large quantities of dust in many of the landslides and collapses also caused frequent dust storms. This dust accumulated on local roads, reaching several centimeters in thickness (Liang and Jiang, 2017). Based on the geochemistry, quartz particle scanning electron microscopy and grain size, Jiang et al. (2014) proposed that lacustrine sediments in east Tibet were transported by wind and trapped in lakes; most of this material originated from local wind-blown sediments. From these results it can be hypothesized that large earthquakes in east Tibet might have caused widespread rockfalls and landslides, leading to exposure of fine-grained sediments that have accumulated on mountain slopes before undergoing aeolian transport into the lake. Interestingly, post-seismic vegetation recovery at landslide sites indicates that the Wenchuan earthquake's impact on regional post-seismic landslide frequency may disappear within two decades of a major event (Yang et al., 2018). This implies that the amount of detrital material released in the area gradually decreases after an earthquake.

The above discussion shows that, strong seismicity ($M \geq 5$) can cause not only liquefaction and/or fluidization of water-saturated sediments, resulting in various types of brittle and/or ductile deformation, but can also generate a large amount of debris (landslides, dust, etc.), which is rapidly transported into lakes by wind and/or rivers, and deposited as a series of turbidites (Howarth et al., 2012; Moernaut et al., 2014; Archer et al., 2019; Fan et al., 2020; Molenaar et al., 2021). Seismic shaking may also trigger sedimentary instabilities on the steep slopes of lake basins (Katz et al., 2009) or lake seiches (Beck, 2009; Avşar et al., 2014), resulting in mobilized masses at proximal sites (i.e., slides, slumps, debris flows) (Zhang et al., 2014) and deposition from induced turbidity currents at distal sites (i.e., seismoturbidites). When an earthquake occurs, pre-existing unconsolidated lake deposits (below the event horizon) may deform and generate various SSDS (Fig. 6), followed by the deposition of medium- to coarse-grained particles released by earthquake-induced processes to form a so-called event layer (rapidly-deposited sand layer formed by material supplied due to effects of the seismic shock) above the event horizon (Jiang et al., 2014; Zhong et al., 2020c) (Fig. 6). Event layers were found to have varying thicknesses in the upper part of the SSDS in the Lixian lacustrine sediments, east Tibet (Jiang et al., 2017). In these SSDS, the grain-size and magnetic susceptibility repeatedly show abrupt coarsening and upward fining, probably because frequent earthquakes have generated a large amount of debris in the lacustrine sediments of the present study area, resulting in an abrupt increase in the terrigenous flux of coarse-grained (magnetic material) in the Lixian dammed paleo-lake (Jiang et al., 2016, 2017). Such earthquakes ($M > 5.0/5.5$) have not always resulted in the formation of SSDS in lacustrine sediments, perhaps due to their greater distance from the epicenter (Owen and Moretti, 2011) or the higher sand/mud ratio of lacustrine sediments (Jiang et al., 2017); nevertheless, they can still generate a large amount of debris within the lake basin. A positive correlation between the thickness of the seismites and the seismic shaking has been identified (Hibsch et al., 1997; Rodríguez Pascua et al., 2003), but it remains unclear whether there is a correlation between the thickness of the event layer (post-seismics deposit) and the seismic shaking.

Keefer (1984) produced the first map to show the area affected by landslides during earthquakes of different magnitudes for 40 historical world-wide earthquakes covering the years 1811 to 1980. The correlation between magnitude (M) and landslide distribution shows that the minimum magnitude likely to trigger landslides is $M \approx 4.0$. This limit was subsequently revised and constrained with further data by Rodríguez et al. (1999) and Xu et al. (2014). Bommer and Rodríguez (2002) presented data for earthquake-triggered landslides in Central America but noted that their characteristics can be compared with global relationships between the area of landslides and earthquake magnitudes (Fig. 7a). Malamud et al. (2004) confirmed the results of Keefer (1984) and proposed that the minimum earthquake magnitude required to trigger landslides is $M_w = 4.3 \pm 0.4$. Under the same geological conditions, the stronger the earthquake, the larger the area and volume of landslides. A strong earthquake can induce tens of thousands of landslides, especially in alpine valleys (Xu et al., 2014; Zhang et al., 2017).

By establishing a high-resolution Holocene seismic history of the Dead Sea Transform, Migowski et al. (2004) found the thickness of the disturbed layer has reasonable correlation with magnitude. This disturbed-layer sedimentary pattern is punctuated by successions of disturbed sedimentary structures that typically consist of aragonite fragments "floating" in a silty detrital matrix without any indication of transport. Furthermore, Moernaut et al. (2014) found that a turbidite's spatial extent and thickness are a function of the local seismic shaking and can be used for reconstructing paleo-intensities. They also established a linear correlation between cumulative turbidite thickness and seismic shaking of the causative earthquake. Theoretically, a larger magnitude earthquake produces more debris (landslide, dust etc.) and a thicker rapidly deposited sand layer (event layer) in local water bodies. Initial conditions should also be considered: for example, the degree of gravitational instability of the sedimentary body; textural parameters; depth of sliding surface; etc.

To assess this theory, we assimilated data on the thicknesses of rapidly deposited sand layers from 32 historical world-wide earthquakes (France, Turkey, New Zealand, and Chile) covering the years 181 to 2010. For more details on determining the genesis and thickness of sand layers, based on abnormal changes in sedimentary indexes, see Avşar et al. (2014) and Jiang et al. (2014). Next, we established a linear correlation between the cumulative sand layer

thickness and magnitude of the causative earthquake (Fig. 7b), yielding a residual mean square of 0.40. Correlation between magnitude (M) and cumulative sand layer thickness shows that the minimum thickness likely to be affected by debris material in a sand layer increases from approximately zero at M = 5.0 to 50 cm at M = 8.0. A cumulative sand layer with a thickness of 1 cm corresponds to earthquake magnitudes of ~4.0-6.0. A cumulative sand layer with a thickness of 10 cm corresponds to earthquake magnitudes of ~5.8-8.4. If the debris (landslide, dust etc.) in the lake drainage basin is induced by earthquakes, and is transported into the lake, then the correlation between the cumulative sand layer thickness and the magnitude can be used to determine the magnitude or shaking of the seismic event under study.

Considering that the denudation mechanism of debris in a lake drainage basin is complicated, and that other factors such as provenance, topography, meteorological conditions and hydrodynamic conditions can also affect the sediment transport, we note that most lake sediments comprise mixtures of sediment populations derived from different sources and transported to the site of deposition by different mechanisms. Therefore, it is difficult to assess the true significance of variations in sediment properties, especially when considering that the provenance of lake sediments may also be influenced by extreme weather events (e.g. rain storms, sand storms, etc.). Consequently, we propose that components of the lake sediments should be separated (see the end-member modeling method: Weltje, 1997), by distinguishing tectonic and climatic components, when studying a seismic event layer and its thickness, for the purpose of determining the seismic shaking. Admittedly, this method is only based on statistical results of the data, it requires more data support and further examination. Therefore, the method of rapidly deposited sand layer thickness can only be used to gain a rough magnitude estimate with limited accuracy, and needs to be compared with alternative methods of verifying seismic shaking.

Other methods

Two independent methods for determining earthquake magnitude are numerical simulations and shaking table experiments. Wetzler et al. (2010) carried out field observations in the Lisan Formation sediments (Dead Sea) and numerical simulations based on peak ground acceleration and its relationship with deformed layer thickness and deformation morphology. Their results reveal that deformation morphology is controlled by ground acceleration and deformed layer thickness. Shaking table experiments are not only limited by the available instruments and equipment, but are also affected by the sediment materials, grain size, porosity and permeability during the experiment (Shen, 2014). Considering the complex rheologies involved, and because the developing structures are impossible to observe in sediments (Harrison and Maltman, 2003), shaking table experiments are used less frequently than other methods mentioned above.

The other method is based on geotechnical engineering techniques at sites of marginal liquefaction (a critical state of liquefaction), to constrain the peak accelerations as a function of epicentral distance: these accelerations can then be compared with predictions from seismological models, and used for back-calculating prehistoric magnitudes (Obermeier, 1996, 1998).

Case Studies

Lake Lisan section

The Dead Sea Basin is a pull-apart graben enclosed between two overlapping segments of the active and seismogenic Dead Sea Fault (Fig. 8a, b), at the boundary between the Arabian plate and the Sinai plate (Galli, 1999 and reference therein). Tens of earthquakes with $M > 6$ have occurred along the Dead Sea Fault during the past ~3 ka and have been the object of studies by several authors (Ambraseys and Finkel, 1991; Ambraseys, 2009; Lu et al., 2017). The two strongest seismic shocks recorded in modern times occurred at the Gulf of Aqaba (southernmost Dead Sea) in 1995, with $M_w = 7.3$ (Klinger et al., 1999; Galli, 1999), and in the Jericho Valley in 1927, with $M = 6.2$ (Shapira et al., 1993). Most of the Lisan Formation is made by laminated sediments of ancient Lake Lisan, composed of alternating laminae of authigenic aragonite and detritus; the latter consists of fine-grained calcite, dolomite, quartz and clay. Undisturbed laminated layers between deformed layers have been interpreted as interseismic intervals (Amotz and Marco, 1995). The deformed layers in the lacustrine Late Pleistocene Lisan Formation and Holocene Dead Sea sediments comprise an almost continuous 70 ka paleoearthquake record in the Dead Sea Basin (Marco et al., 1996). Based on the historical earthquakes, current seismic activity near the Dead Sea, a syndepositional normal fault in the Lisan Formation (near Masada) unconformably overlain by undisturbed layers, sedimentary markers (e.g. fragment-supported texture, etc.) and the SSDS (folded layer), Marco and Agnon (1995) interpreted such layers as seismites formed by fluidization of the upper sediment layers at the bottom of the lake during strong earthquake shaking, representing strong seismic events ($M \geq 5.5$).

To better constrain the magnitudes associated with the mixed layers in the Lake Lisan section (Fig. 8c), we use the six methods reported above to determine the magnitude or shaking of these seismic events. These yield the following results: (1) the minimum earthquake magnitude required to induce liquefaction should be higher than 5; (2) the distance of the Lake Lisan section to the syndepositional faults at Masada is < 5 km (Marco and Agnon, 1995), which according to the R-M map (Fig. 3) yields earthquake magnitudes of 5.2–6.8; (3) applying the relationship between seismic paleointensities and thicknesses of the paleoseismic horizons (Fig. 4) to the 13 cm-thick disturbed layer in the Lake Lisan section (Fig. 8c) yields a corresponding seismic shaking of 0–0; (4) applying the empirical formula ($M = T/3 + 3.83$) to the 10 cm-thick seismics-related deposits in the Lake Lisan section yields a corresponding earthquake magnitude of 6.83; (5) applying the relationship between mixed (breccia) layers and earthquake magnitudes (Fig. 5) yields corresponding earthquake magnitude of 5.5–6.5 (Table. 2); and (6) applying the relationship between earthquake magnitude and rapidly deposited sand layer thickness (Fig. 7b) to the 2 cm-thick sand layer above the seismics-related deposit in the Lake Lisan section yields a corresponding earthquake magnitude of 4.2–6.6. Therefore, we suggest that the seismics-related deposit presented in Fig. 7 of the Lake Lisan section near Masada in the Dead Sea graben correspond to earthquake magnitudes of 5.5–6.5 (Table 2).

Table 1
Comparative analysis of six methods for determining the magnitude of paleoearthquake.

Method	Advantage	Shortage
Proposed methods	Wide applicability; convenient and efficient	Only the lower limit of magnitude can be given, and have a large error.
Maximum liquefaction distance	Wide applicability; higher accuracy	The location of seismogenic faults is uncertain.
Thickness of disturbed layers	Wide applicability; convenient and efficient	The definition of disturbance layer thickness is indeterminacy, and have a large error.
Empirical formulae	Convenient and efficient	The definition of mixed layer thickness is indeterminacy, and only suitable for deformation structures with thickness less than 15cm.
soft-sediment deformation structures	Wide applicability; higher accuracy	It is necessary to determine the origin and types of soft sediment deformation structures
Thickness of rapidly deposited sand layer	It is suitable for seismic events that do not induce the soft sediment deformation structures in the profile and only can verify the earthquake magnitude estimated by soft sediment deformation structure	It is necessary to determine whether the rapidly deposited sand layer is caused by earthquake, and have a large error.

Table 2
Earthquake magnitudes inferred from Lake Lisan and Luobozhai soft-sediment deformation structures.

Study location	Types	Historical liquefaction method	Maximum liquefied distance	Thickness of disturbed layer		Empirical formula		Deformation types		Thickness of rapidly-deposited sand layer		M
		M	Distance (km)	M	Thickness (cm)	Shaking	Thickness (cm)	M	M	Thickness (cm)	M	
Lake Lisan	Seismics-related deposit (liquefied layer and dsrupted layer)	> 5	5–10	5.2–6.8	13	VII–Ⅷ	9	6.83	5.5–6.5	2	4.2–6.6	5.5–6.5
Luobozhai	liquefied diapir	> 5	5–10	5.2–6.8	10	Ⅷ–Ⅸ			6.5–8.0	5	5.2–7.6	6–7

Luobozhai lacustrine section

The upper reaches of the Minjiang River are located in the north–central part of the “South–North Seismic Belt of China”, which is affected by the Minjiang, Huya, Songpinggou, and Longmenshan faults, and characterized by frequent earthquakes (Fig. 9b). In recent years, four earthquakes with $M_s \geq 7$ have occurred in east Tibet (Wenchuan 2008, Yushu 2013, Lushan 2013 and Jiuzhaigou 2017), evidencing that the area is ongoing a strong active tectonics. Due to the region’s alpine valley geomorphology and lack of Quaternary deposits, it is difficult to conduct paleoearthquake research by excavating trenches. The region contains many lakes dammed by paleoearthquake-induced landslides (Wang et al., 2005). Researchers have ascribed various SSD structures and environmental indexes (such as grain size and magnetic susceptibility) in the lacustrine sediments to multiple seismic events (e.g., Wang et al., 2011; Jiang et al., 2014, 2016; Xu et al., 2015; Zhong et al., 2019).

We recognized liquefied diapir in the Luobozhai Late-Quaternary lacustrine section, at Maoxian in the upper reaches of the Minjiang River, which is overlain and underlain by undeformed beds (Fig. 9b). Criteria for correlating SSDS with seismic events have been widely discussed (e.g. Sims, 1975; Bowman et al., 2004; Owen and Moretti, 2011; Jiang et al., 2016), with the following considerations. (1) Most researchers emphasize the location of the SSDS in a seismically active region. Since 1933, at least five earthquakes with $M > 7$ have occurred along the nearby Minjiang River fault, at Diexi (1933), Songpan (1976), northern Pingwu (1976), Wenchuan (2008) and Jiuzhaigou (2017). (2) The Luobozhai section is located close to the Minjiang Fault and is only 10 km from the Songpinggou Fault. Tectonic activity along these faults during the late Quaternary has been confirmed by many studies (e.g. Chen et al., 1994; An et al., 2008; Zhang et al., 2016; Zhong, 2017, Ren et al., 2018). (3) The widely distributed lacustrine sediments mostly comprise liquefaction-prone deposits, such as silt and fine sand, with grain sizes concentrated in the range of 20–200 μm : for example, the nearby (about 10 km) Xinmocun (Jiang et al., 2014) and Shawan sections (Zhong, 2017), which have great potential for liquefaction and fluidization (Lee et al., 2003). (4) The SSDS in the study area can be traced laterally for 30 km along the Minjiang River, and are repeated vertically (Wang et al., 2011; Jiang et al., 2014; Zhong et al., 2019). (5) The liquefied diapir are separated by undeformed beds, and only contains centimeter-scale deformation structures (Fig. 9b). (6) The liquefied diapir in the Luobozhai section is comparable with those produced experimentally (Anketell et al., 1970; Alfaro et al., 1997) and with those described in studies of paleoearthquakes (Jiang et al., 2014; Zhong et al., 2019). (7) The deformation structures in flood retention deposits have parallel sandy layers, a distinct change in grain size, and a deformation magnitude that decreases downwards in the vertical section (Rana et al., 2016); typical morphologies are simple

folds to complex contortions. However, if the disturbance were due to fluvial processes, then it is likely that the deformation caused by liquefaction and fluidization would have taken place just only near the sedimentary surface (Kundu et al., 2011). Structures typical of rapid pore-water movements (Li et al., 1996), slope failure (Alsop and Marco, 2011), channel erosion (Dasgupta, 1998) and permafrost (Vandenberghe, 1992) were not observed. The liquefied diapir were likely triggered by seismic activity, although overloading (Moretti and Sabato, 2007) and rapid sedimentation (Lowe, 1975) cannot be discounted. Therefore, the liquefied diapir in the Luobozhai lacustrine section are attributed to seismic activity, and were not influenced by flash floods, fluvial processes or other processes. Considering that the liquefied diapir in Luobozhai section are strongly deformed, and that the deformed layer contains a notable clay content that will result in a relatively high viscosity, a relatively large magnitude would be required to cause its liquefaction related deformation (Tian et al., 2016; Su and Qiao, 2018). By using the six methods described above to determine the earthquake magnitudes, we suggest that the liquefied diapir in the Luobozhai section may represent earthquake activity with magnitudes 6–7 (Table 2).

Liquefaction hazard

Sand or soil liquefaction and liquefaction-induced lateral spreading are the most pervasive types of seismic damage to structures and critical facilities infrastructures (e.g. buildings, roads, bridges etc.) (Obermeier, 2005; Chen et al., 2013; Liu et al., 2016; Das et al., 2018; Naik et al., 2019). Such damages have been observed in almost all recent major earthquakes, including the 1999 Chi-chi (Taiwan) earthquake (Chang et al., 2012), the 2008 Wenchuan (China) earthquake (Liu-Zeng et al., 2012), the 2010 Maule (Chile) earthquake (Yasuda et al., 2010), the 2011 Christchurch (New Zealand) earthquake (Robinson et al., 2014), the 2011 Tohoku (Japan) earthquake (Sasaki et al., 2012), the 2012 Emilia (Italy) earthquake (Papathanassiou et al., 2015), and the 2018 Palu (Indonesia) earthquake (Song et al., 2019). Current methods of predicting the liquefaction potential were largely developed using data from relatively young deposits or deposits that have been frequently disturbed (i.e., in areas of high seismicity) (Saftner et al., 2015). However, it is difficult to assess the liquefaction potential in older deposits. Liquefaction and fluidization, as well as SSDS, are common phenomena associated with seismic ground shaking, and have been reported from modern to Paleozoic times (Mazumder et al., 2006; van Loon, 2009; Qiao et al., 2017) in all sedimentary environments (Moretti et al., 1999). In many places, the SSDS record can be interpreted far enough back in time to yield sufficient data (e.g., paleoseismic activity and earthquake magnitude) for hazard assessment, thereby providing a sound basis for the engineering and design of critical structures such as hospitals, bridges, and nuclear facilities (Obermeier et al., 2005). Therefore, the use of SSDS constitutes an effective and inexpensive approach to recognize zones of different liquefaction hazard, making it readily applicable in addressing (1) whether there has been strong Holocene or late Pleistocene shaking in the region (Obermeier et al., 1985); (2) where the tectonic sources (e.g. paleo-earthquake epicentre) were located (Rodríguez-Pascua et al., 2010); and (3) what was the strength of the shaking (Zhong et al., 2017). Therefore, the SSDS have a lower formation threshold, higher preservational potential and longer time than geomorphic markers formed by surface ruptures. Thus, future combined studies of SSDS and Quaternary surface faults are required to provide reliable paleoseismological information in tectonically active regions.

Discussion

The types of SSDS and the maximum liquefaction distance together yield a more reliable method to determine the seismic shaking than the six methods described above, and their usefulness is demonstrated by the two case studies. However, the following five points should be considered when we determine the magnitude of an earthquake using SSDS. (1) Due to the complex deformation mechanism and triggering factors of the SSDS, we must firstly determine the origin and trigger mechanism of the SSDS (Sims, 1975; Mills, 1983; Jones and Omoto, 2000; Owen and Moretti, 2011; Zhong et al., 2020b). (2) The morphology of the SSDS and the thickness of seismites are controlled by the properties (e.g. sedimentary environment, lithology, physical characteristics of the sediment, liquefaction and/or fluidization factors) of the SSDS (Hurst et al., 2011). These factors could lead to incorrect determination of the type of SSDS (e.g. flame vs diapir); therefore, careful consideration needs to be paid when analysing the SSDS characteristics (type, deformation strength, disturbed thickness etc.). (3) Uncertainty of the paleoearthquake location and the spatial distribution of the paleoliquefaction point must be considered. To minimize uncertainty, we use the horizontal distance between the liquefaction deformation record (or the location of the liquefied deformation section) and the seismogenic fault or epicenter. (4) In order to verify the method's reliability, the reconstructed seismic shaking should be comparable with the magnitudes obtained by the modern earthquakes, paleoearthquakes in the study area. (5) Earthquakes with $M > 5$ may not cause SSDS or may not be recorded in the sedimentary succession (Moretti et al., 1999), and a single seismic shock can induce superimposed deformed beds (Gibert et al., 2011). This can lead to the omission or enhancement of seismic events.

Hence, when studying paleo-earthquakes on the basis of SSDS, the sedimentology and micro-indices (e.g. grain size, magnetic susceptibility etc.) are necessary to identify those seismic events that did not cause SSDS but may have induced large amounts of debris (e.g. landslides, dust etc.). This can help us to obtain long sequences of seismic events. However, the source and chemical properties of the lacustrine sediments are affected by the strong-earthquake-induced debris, and may also be affected by extreme weather (e.g. sand storms, Chen et al., 2013), rapid weathering (Liang and Jiang, 2017) and human activities (Karlin and Abella, 1996). Therefore, quantifying seismic events in lacustrine sediments requires a combination of macroscopic SSDS and micro-environmental indicators, as well as sedimentary evidence. Similarly, for lacustrine sediments, it is best to use a combination of an underlying SSDS layer and an overlying sand layer, in addition to the six methods described here, to obtain accurate magnitudes of paleo-earthquakes.

Conclusions

Six methods developed to infer the magnitude of past earthquakes were applied to a succession of the Lisan Formation, and yielded earthquake magnitudes of > 5.0 , 5.2–6.8, 6.83, 5.5–6.5, and 4.2–6.6, respectively. We therefore propose that the seismics-related deposit in the Lake Lisan section most likely represents earthquakes with magnitudes of 5.5–6.5, thus confirming previous estimates of magnitudes ($M \geq 5.5$). Application of these same methods to the liquefied diapir in late-Quaternary lacustrine sediments in the upper reaches of the Minjiang River at Luobozhai indicates that the liquefied

diapir are associated with earthquake magnitudes of 6–7. This estimate agrees well with the results of previous studies. Our work provides a means of gaining a better understanding of the regional seismic hazard at sites where it is difficult to conduct paleoearthquake research by excavating trenches, for example due to rough topography, strong erosion or the absence of co-seismic ruptures, especially the paleoearthquakes recorded by lacustrine sediments. The combination of the six methods can provide a relatively convenient and novel method for determining seismic shaking. This study also can serve as a reference for comparing methods of calculating the magnitude of a paleoearthquake based on the surface rupture parameters. This new, integrated method provides reliable and comparable information on the seismic shaking, suggesting that a similar approach could be successfully applied to quantify the magnitude of an earthquake on the Tibetan Plateau or elsewhere. It also provides a novel method for us to research the long-term seismic activity and risk in tectonically active regions.

Declarations

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Figures

Figure 1

Spatial distributions of relationship between soft sediment deformation structure or rapidly deposited sand layer thicknesses and earthquake magnitude, and earthquakes with magnitude over 6.0 from 1900 to 2021 all over the world (data sources from <http://earthquake.usgs.gov>).

Figure 2

Relationship between the maximum epicentral distance of liquefied sites R and magnitude M (modified from Kuribayashi and Tatsuoka, 1975). Numbers represent earthquakes which induced liquefaction in Japan during the years 1872-1968. The mean line is for all dates, and the lower bound is for $M \geq 6$.

Figure 3

Relationship between earthquake magnitude and the epicentral distance to farthest liquefaction feature (modified from Qiao et al., 2017). The gray rectangle indicates the recorded liquefaction in the Lake Lisan and Luobozhai section ($M = 5.2-6.8$).

Figure 4

Historical seismicity of Quito (Ecuador) and the attribution of seismic shaking to the paleoseismic levels (modified from Hibschi et al., 1997). The X scale is based on by historical seismicity data. The Y and Z scales both consider a multiplicative factor of X based on observations of paleo-seismic horizons in lacustrine deposits in California (Sims, 1975) and on mathematical relations between horizontal ground acceleration and seismic shaking (Trifunac and Brady, 1975).

Figure 5

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Schematic showing different types of seismically-induced SSDS in lacustrine deposits, with an indication of the earthquake magnitudes at which the different SSDS are interpreted to form (modified from Rodriguez-Pascua et al., 2000; Berra and Felletti, 2011; Zhong et al., 2017).

Figure 6

Correlation of soft sediment deformation structures with changes in sedimentary index values (grain size and magnetic susceptibility) from Xinmocun, east Tibet (Jiang et al., 2014).

Figure 7

(a) Area affected by landslides during earthquakes of different magnitudes; (b) suggested linear correlation between rapidly deposited sand layer thicknesses and earthquake magnitudes of the responsible earthquakes.

Figure 8

(a-b) Tectonic plates in the Middle East. The Dead Sea Transform (DST) transfers the opening of the Red Sea to the East Anatolian Fault and the Taurus–Zagros collision zone (modified from Marco and Agnon, 2005). (c) Elements of a typical intraclast breccia layer overlying a disrupted sequence. A typical seismics-related deposit in late-Pleistocene fan delta sediments which overlie laminated layers shows a gradual upwards transition from the folded strata of Masada near the lacustrine Lisan Formation in Israel, through a fragment-supported texture, to a matrix-supported texture at the top (modified from Marco and Agnon, 1995).

Figure 9

(a) DEM of the study region illustrating the geomorphology and drainage system, (b) and the location of the Luobozhai section (purple star). The red solid circles show the seismicity between 1900 and 2012 (modified from Jiang et al., 2014). Surface ruptures and active faults are after Xu et al. (2009). (c) A typical liquefied diapir in the Luobozhai lacustrine sediment was induced by an earthquake during the late-Quaternary, in the upper reaches of the Minjiang River, east Tibet.