Last Deglacial Soft–Sediment Deformation at Shawan on the Eastern Tibetan Plateau and Implications for Deformation Processes and Seismic Magnitudes

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Abstract: The eastern margin of the Tibetan Plateau is characterized by frequent earthquakes; however, research of paleo–earthquakes in the area has been limited, owing to the alpine topography and strong erosion. Detailed investigations of soft–sediment deformation (SSD) structures are valuable for understanding the trigger mechanisms, deformation processes, and the magnitudes of earthquakes that generate such structures, and help us to understand tectonic activity in the region. To assess tectonic activity during the late Quaternary, we studied a well–exposed sequence of Shawan lacustrine sediments, 7.0 m thick, near Lake Diexi in the upper reaches of the Minjiang River. Deformation is recorded by both ductile structures (load casts, flame structures, pseudonodules, ball–and–pillow structures, and liquefied convolute structures) and brittle structures (liquefied breccia, and microfaults). Taking into account the geodynamic setting of the area and its known tectonic activity, these SSD structures can be interpreted in terms of seismic shocks. The types and forms of the structures, the maximum liquefaction distances, and the thicknesses of the horizons with SSD structures in the Shawan section indicate that they record six strong earthquakes of magnitude 6–7 and one with magnitude >7. A recent study showed that the Songpinggou fault is the seismogenic structure of the 1933 Ms7.5 Diexi earthquake. The Shawan section is located close to the junction of the Songpinggou and Minjiang faults, and records seven earthquakes with magnitudes of ~7. We infer, therefore, that the SSD structures in the Shawan section document deglacial activity along the Songpinggou fault.

Key words: lacustrine sequence, soft–sediment deformation (SSD), deformation process, earthquake magnitude, Shawan, eastern Tibetan Plateau


1 Introduction

The eastern margin of the Tibetan Plateau (TP) is characterized by tectonic activity (Deng Qidong et al., 2014; Li et al., 2016), frequent earthquakes (Chen et al., 1994; Zhou Rongjun et al., 2000; Li et al., 2013, 2014, 2016; Ran et al., 2014; Zhong Ning et al., 2017b; Liang et al., 2018), and alpine valleys (Jiang et al., 2014, 2016; Xu et al., 2015). Due to the region’s geomorphology of alpine valleys and lack of Quaternary deposits, it is difficult to conduct paleoearthquake research by excavating trenches. Seismicity can cause liquefaction and/or fluidization of granular solids that result in various types of soft–sediment deformation (SSD) structure. They occur during or after deposition, then the sediment is still unconsolidated, providing valuable information about the syn– or post–depositional physical processes that have taken place in the sedimentary environment (Van Loon et al., 2014, 2016), the hydrodynamic conditions (Li et al., 1996; Ran et al., 2016), and the paleo–seismic characteristics (Sims,1975; Jiang et al., 2014, 2016). In tectonically active regions, SSD structures provide reliable evidence for the occurrence of paleo–earthquakes (e.g., Lowe, 1975; Allen, 1986). In recent years, researchers have found a range of successive dammed palaeolacustrine sediments in the upper reaches of the Minjiang River, and they have ascribed various SSD structures and high–resolution indexes (such as grain sizes and magnetic susceptibility) in the lacustrine sediments to multiple seismic events (e.g., Wang et al., 2011; Jiang et al., 2014, 2016, 2017). Although this work has received much attention, further investigations in the region are necessary for the following reasons.

Firstly, the types of SSD structure described comprise only flame structures, ball–and–pillow structures, microfaults, slump folds, and elastic dykes (Wang et al., 2011; Jiang et al., 2016), whereas elsewhere in the world
liquefied convolute structures (Rodriguez−He Bizhu and Qiao Xiufu, 2015; Van Loon et al., 2016), other structures such as load casts (Owen, 1996; Moretti et al., 1999; Moretti and Sabato, 2007; Van Loon et al., 2014), pseudonodules (Rodriguez−López et al., 2007; He Bizhu and Qiao Xiufu, 2015; Van Loon et al., 2016), liquefied convolute structures (Rodriguez−Pascua et al., 2000; Qiao Xiufu et al., 2017), liquefied breccia (Plaziat et al., 1990; Guiraud and Plaziat, 1993; Qiao Xiuf et al., 2012; Tian Hongshui et al., 2013), thixotropic wedges (Qiao Xiuf et al., 2017), and loop bedding (Rodriguez−Pascua et al., 2000) have also been described. We suggest it is necessary, therefore, to identify more types of SSD structure over a larger area and for a longer time scale than has previously been accomplished. Secondly, in much of the previous work, only the shape, form, and classification of the SSD structures have been considered. What has been lacking is a systematic discussion of the trigger mechanisms, deformation processes, and earthquake magnitudes related to the SSD structures.Initially, some researchers believed that a Ms 2−3 earthquake was large enough to trigger liquefaction (e.g., Seed and Idriss, 1971). Later, Scott and Price (1988) proposed that earthquakes with magnitudes smaller than 5 do not cause significant liquefaction of sediments at distances more than 4 km from the epicenter, and that an event with magnitude 7 does not significantly affect sediments beyond 20 km. These values are consistent with the scientific consensus that the triggering of liquefaction occurs only for earthquakes larger than magnitude 4.5 (Marco and Agnon, 1995). Hibsch et al. (1997) summarized the seismic origin of contorted bedding features, and compared the distributions of the thickness of these horizons with seismic intensities of historical record, then proposed a methodology for the determination of seismic paleointensities. Considering the paucity of the statistical data, and the lack of experimental validation, their methodology is used only as a reference. Rodriguez−Pascua et al. (2003) established an empirical relationship between the mixed−layer thickness of the studied sediments and the earthquake magnitude, and obtained an empirical formula \( M = T/3 + 3.83 \), where \( T \) (unit/cm) is the thickness of the mixed−layers and \( M \) is the magnitude. But there was no clear qualification standard for the thickness of mixed−layer, thus limiting its application. Rodriguez−Pascua et al. (2000) summarized the different SSD structure types and their relationships with earthquake magnitudes, and suggested that mixed layers with fluidization, pseudonodules, ball−and−pillow structures, and liquefied veins correspond to magnitudes of 5.5−6.5, 6.5−8, 6−8, and 5−8, respectively. Taking into account the distances to seismogenic faults, Neuwirth et al. (2006) tentatively postulated that the earthquake magnitude required to generate SSD structures 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 irregular convolute structures or highly distorted stratification. Clearly, therefore, the relationships between the types and forms of SSD structure and earthquake magnitudes remain subjects of debate. Thirdly, almost all the previous dating results in the area were based on the OSL dating method, but errors associated with this method are generally around 10%, sometimes even exceeding 20% (e.g. Liu et al., 2010; Wang et al., 2011; Yang et al., 2012). We suggest, therefore, that it is necessary to improve the precision and accuracy of the dating methods. Fourthly, some researchers suggested that the 1933 Ms 7.5 Diexi earthquake was caused by activity on the N−S−trending Minjiang fault (Chen et al., 1994; Deng Qidong et al., 1994; Qian Hong et al., 1999; Zhou Rongjun et al., 2000; Wang Kang and Shen Zhenkang, 2011), but others have assigned the activity to the Songpinggou fault (Tang Rongchang et al., 1983; Huang Zuzhi et al., 2002; Ren et al., 2018). There is a need, therefore, to clarify the seismogenic tectonics of the Diexi earthquake.

The upper reaches of the Minjiang River are located in the north−central part of the “South−North Seismic Belt of China”, which is controlled by the Minjiang, Huya, Songpinggou, and Longmenshan faults (Fig. 1b), and characterized by frequent earthquakes and widely distributed paleo−dammed lacustrines (Wang et al., 2011). Various SSD structures are readily observed in these lacustrine sediments (e.g., Wang Ping et al., 2009; Wang et al., 2011; Jiang et al., 2014, 2016; Xu et al., 2015). The 1933 Ms 7.5 Diexi earthquake was a catastrophic event with the loss of over 10,000 lives in eastern Tibet (China Earthquake Administration, 1999). The earthquake caused the formation of dammed lakes (eight of which remain today), and two months later a breakout from those lakes killed at least 2,500 people in Maoxian, Wenchuan, and Guanxian (Dujiangyan) counties along the middle and lower reaches of the Minjiang River (Chang Longqing, 1938). However, the 1933 Diexi earthquake occurred during the Chinese Civil War, and society was so chaotic at that time that detailed field investigations were not undertaken. The source of this earthquake remains debated and therefore requires further investigation (Tang Rongchang et al., 1983; Chen et al., 1994; Huang Zuzhi et al., 2000; Wang Kang and Shen Zhenkang, 2011; Zhang Yueqiao et al., 2016; Ren et al., 2018).

In this work, we chose to study the last deglacial lacustrine sediments at Shawan, adjacent to Lake Diexi, in the upper reaches of the Minjiang River. The aims of our research were (1) to study the SSD structures and interpret them in terms of trigger mechanisms, (2) to evaluate the relationships among the various types of SSD structures, the thickness of the deformation layer, and the maximum distance of liquefaction deformation with respect to the earthquake magnitudes in the Shawan section, and (3) to discuss the seismogenic tectonics of the 1933 Diexi earthquake. The results of our work provide a better understanding of the patterns of tectonic activity in the TP, the growth of the TP, and the evolution of depositional features along the eastern margin of the TP during the late Quaternary.

2 Geographic and Geologic Settings

The Shawan section (32.07°N, 103.71°E; 2219.8 m a.s.l.) lies beside the Diexi paleo−dammed lake between Maoxian and Songpan in the upper reaches of the Minjiang River, Sichuan Province, China (Figs. 1, 2). In this area, Quaternary deposits, such as river terraces,
diluvium deposits, and paleo-lacustrine sediments are distributed mainly along the Minjiang River and its tributaries, and the layers of lacustrine sediments generally exhibit different degrees of tectonic deformation (Wang et al., 2011; Jiang et al., 2014, 2015; Li Yanhao et al., 2015; Liang and Jiang, 2017) that are controlled by the active Longmenshan, Minshan, and Songpinggou faults. The 2008 Wenchuan earthquake occurred in the transition zone between the Tibetan Plateau and the Sichuan Basin. Within the Longmenshan region, it produced co-seismic surface ruptures >240, ~80, and 6 km long along the Yingxiu–Beichuan, Anxian–Guanxian, and Xiaoyudong faults, respectively (Li Haibing et al., 2008, 2009; Xu et al., 2009; Liu–Zeng et al., 2010; Zhang et al., 2010). At Shawan, close to Lake Diexi, various types of SSD structures have been identified including liquefied convolute structures, flame structures, pseudonodules, and ball-and-pillow structures, and Wang et al. (2011) suggested these deformation structures were caused by earthquakes. Moreover, in the Xinmocun lacustrine sediments beside Lake Diexi, Jiang et al. (2014) discerned flame and pseudonodule deformation structures, and with the assistance of high-resolution grain-size and magnetic susceptibility...
data they were able to discern repeated events of abrupt coarsening and upwards fining, indicating 26 possible paleo-earthquake events. At Diaolin, between the southern end of the Minjiang fault and the northeastern end of the Maoxian–Wenchuan fault, a paleo-lake formed, probably due to an earthquake in AD 638 that caused a rockfall of poorly–sorted angular phyllite from the opposite mountain slope, as well as SSD structures (folds and microfaults) (Xu et al., 2015). At Lixian, in lacustrine sediments northwest of the southern segment of the Longmenshan fault, we observe SSD structures (e.g. clastic dykes, ball–and–pillow structures, flame structures, and microfaults), as well as variations in the high–resolution grain–size and magnetic susceptibility data that record successions of paleoseismic activity (Jiang et al., 2016, 2017). These prehistoric seismic events, revealed by previous studies, thus provide invaluable information on the long–term behavior of local seismic activity in the eastern TP.

Instrumental data since AD 1900 indicate that the TP has experienced a strong clustering of earthquakes around the Bayan Kala Block from 1995 to the present, known as the Kunlun–Wenchuan earthquake series (Deng Qidong et al., 2014). Our study area lies on the eastern margin of the Bayan Kala fault block and shows a high seismic risk, being frequently struck by strong earthquakes with magnitudes greater than 7 (Fig. 1a). On the basis of historical records that cover the last AD 638 years, there have been six earthquakes with magnitudes greater than 7.0, and more than nineteen with magnitudes between 5.0 and 6.9 in the eastern TP (China Earthquake Administration, 1995, 1999). Those larger than magnitude 6.0 occurred along the Longmenshan and Minshan faults. Furthermore, these events were concentrated in areas with rapid changes in elevation, which also correspond to locations with surprisingly high GPS–measured uplift rates of as much as 2–3 mm/yr (Liang et al., 2013). Tectonic activity along these faults during the late Quaternary has been confirmed by many studies (e.g., Tang Rongchang, 1983; Huang Zuzhi et al., 2002; Zhang et al., 2006; Dengsmore et al., 2007; Royden et al., 2008; Hubbard and Shaw, 2009; Xu et al., 2009; Ren et al., 2010; Ren et al., 2013a, b; Ran et al., 2014; Liu et al., 2015; Ren et al., 2018), and it created the various SSD structures observed in lacustrine sediments because they are located close to the junction of the Minjiang fault and Songpinggou fault.

3 Stratigraphy and Chronology

The well–exposed Shawan section is about 7.0 m thick and mostly composed of light yellow silty clay, light yellow and light gray silt with massive structure, and dark–grayish sand with parallel bedding (Fig. 3). The section is divided into four parts according to lithology. Lithology 1, at a depth of 7.0–6.1 m, consists of dark–grayish sand with a loose texture, see parallel lamination and wavy ripple lamination at the bottom and top of the section, respectively. Lithology 2 (6.1–4.2 m) is defined by light–yellow silt interbedded with dark–grayish fine sand or silt. It can be divided into three sub–layers: 6.1–5.75 m, light–yellow clayey silt with a massive structure; 5.75–4.5 m, dark–grayish silt interbedded with light–yellow clayey silt; 4.5–4.2 m, light–yellow clayey silt with a massive structure. Five types of SSD structures, namely load casts, flame structures, ball–and–pillow structures, microfaults, and liquefied convolute structures, can be observed in the upper part of this unit. Lithology 3 (4.2–2.35 m) consists of dark–grayish sand, and dark grayish fine sand with light–yellow clayey silt at the top part of the section (2.7–2.35 m), with parallel lamination in the bottom. It displays two types of SSD structure, namely pseudonodules and liquefied breccia. Lithology 4 (2.35–0 m) is made up of dark–grayish silt interbedded with light–yellow clayey silt (Fig. 3). It can be divided into three sub–layers: 2.35–2.1 m, a dark–grayish silt with a light–yellow clayey silt; 2.1–1.6 m, dark–grayish silt; 1.6–0 m, dark–grayish silt interbedded with light–yellow silt. Generally speaking, the lithologies and sedimentary structures observed in the Shawan section are typical of the lake–shore sediments in the area, and they therefore provide a good opportunity for studying the SSD structures that formed in this tectonically active region in the upper reaches of the Minjiang River.

The ages of sediments in the Shawan section have been assessed previously. Yu Song (2010) and Wang et al. (2011) measured the optically stimulated luminescence (OSL) of quartz grains with diameters of 4–11 μm, obtaining age groups of 26–22 ka and 25–20 ka, respectively. The OSL ages are apparently consistent; however, they require further examination. We collected three charcoal samples at depths of 2.25 m (SW16–04), 3.05 m (SW16–03), and 3.40 m (SW16–01) in the Shawan section, and then performed 14C age measurements in the Beta Laboratory (Table 1). The samples yielded 14C calibrated ages of 15855–15585 yr before present (BP) (SW16–04), 15585–15285 (SW16–03), and 16870–16465 yr before present (BP) (SW16–01) (Fig.3 and Table.1). Ages from the three samples are well consistent and therefore provide reliable ages for the Shawan section, especially for the layers containing the SSD structures.

We note that the Accelerator Mass Spectrometry 14C ages for charred material, and the OSL ages, provide significantly different results for the Shawan section. The OSL ages seem to be older than 4000–9000 years, which is probably due to the fine–grained quartz (4–11 μm) not having been able to completely reset the OSL signal prior to deposition (P. Wang and J. Chen, pers. Comm.), leaving a higher residual dose (Cunha et al., 2010; Zheng et al., 2010; Yang et al., 2012).

4 Morphological Features of the SSD Structures

The well–exposed Shawan section (Fig. 3) allows detailed observation of lateral variations within individual beds. Thus, deformed beds can be clearly distinguished from undeformed beds. The layers are flat–lying, which rules out the possibility that slope or landslide failures
could have been responsible for the observed SSD structures (Jiang et al., 2016). We identified seven deformed beds within the 7.0–m-thick interval of the Shawan lacustrine sequence (Fig. 3). These deformed beds contain seven types of SSD structures: load casts, flame structures, pseudonodules, ball-and-pillow structures, microfault, liquefied breccia, and liquefied convolute structures (Fig. 3).

4.1 Load casts

Load casts (Alfaro et al., 1997) are observed at depths of 2.7–2.9 and 4.2–4.5 m in the Shawan section (Figs 4, 7, 9), and are overlain and underlain by undeformed beds. It is possible to define several types of load-cast structures, or alternatively drop structures, based on their morphology: simple load-casts, sagging load-casts, and irregular load-casts, which are also known as drop structures. Simple load casts are well developed in the Shawan section (Fig. 4a, c), and they form when the thick overlying sand layer gently sinks into the silty-clay layer below, forming a concave shape. Simple load casts are also seen to occur with flame structures. A typical simple load structure is 0.5–1.0 m wide and 0.2–0.5 m high. Sagging load-cast structures occur in the upper and middle sequence of the Shawan section (Figs. 4b, 7a, b), and these structures are characterized by their perfectly continuous lamination and stratification, and a convex-downwards morphology. Between two adjacent convex lobes, the lamination and/or the stratification form an acute angle upwards that generally exceeds 60 degrees. Sagging load casts are typically 5–10 cm wide and 10–20 cm high. Irregular load-cast structures, originally described by Anketell et al. (1970) and Alfaro et al. (1997), occur in strata ~5–40 cm thick. Some irregular load casts are down-sink drop structures, which develop in fine-sand beds overlying silt or silty-clay, forming small anticlinal and synclinal structures (Figs. 4b, 7c). Other irregular load-cast structures include large, dark sand balls ~20 cm in diameter, which have sunk entirely into the silty-clay layer (Fig. 9b). These deformation structures are typically 5–40 cm wide and 5–25 cm high.

Load structures form in response to unstable density contrasts (i.e., density loading) or lateral variations in loading (i.e., uneven loading) when the sediment becomes liquefied or otherwise loses strength (Owen, 2003). Cyclic or impulsive loading effects caused by events such as storm waves, overloading, and seismic shaking can induce the observed liquefaction features (Owen et al., 1996; Moretti et al., 1999; Owen and Moretti, 2008). However, with respect to the load casts of the Shawan section: (1) the area has frequent earthquakes (Nie Gaozong et al., 2004; Wang et al., 2011; Jiang et al., 2014); (2) the load structures involve fine-grained sediments within deformed layers, and are similar to those observed at other sites nearby (e.g., Anketell et al., 1970; Alfaro et al., 1997, 2010; Owen, 2003; Ghosh et al., 2012); (3) a storm wave origin can be discarded, because the area is inland (Alfaro et al., 2010); and (4) the load structures in the Shawan section can be compared with the results of experimental simulations of load structures (Kuenen, 1958; Owen, 1996; Moretti et al., 1999), produced by seismic shaking in systems with gravitational instabilities (e.g., Owen, 1996;

![Fig. 3. Simplified lithological column of the Shawan lacustrine section showing the positions of SSD layers.](image-url)
Moretti et al., 1999; Moretti and Sabato, 2007).

4.2 Flame structures
The Shawan lacustrine section only contains flame structures at depths of 3.8–4.3 m, and they are overlain and underlain by undeformed beds. The flame structures we observed are typically composed of clay or silty–clay that are overlain by a thick sand layer, and the sizes of the flame structures vary on the meter scale (Fig. 5). At the time of flame structure formation, the underlying layer of silty clay was more viscous than the overlying sand layer, and it was therefore intruded into the sand to form structures shaped like the head of a dragon. The sand layers are massive, whereas clay laminae are fragmentary and undulatory in nature. These structures are typically 1–2 m wide and 0.3–0.5 m high. Flame structures have been reported in the lower Spiti Valley in the Tethys Himalaya (Mohindra and Bagati, 1996), the Nijar Basin in SE Spain.

Fig. 4. Load casts and flame structures in the Shawan lacustrine section. Figures mean the depth of the SSD structures in the profile (similar for Figs. 6–9).
Flame structures form due to reverse density gradients (Visher and Cunningham, 1981; Mills, 1983) that can result from the differential porosities of freshly deposited, poorly-sorted sand and well-sorted silty–clay, and they occur in response to a suitable trigger that perturbs the unstable system. They can also develop by differential dynamic viscosity in heterolithic deposits (Neuwerth et al., 2006). Other non-seismic factors can also produce flame structures, such as channel erosion, gravity slides, sudden changes in groundwater level, and permafrost. Furthermore, under the gravitational stresses of slope sliding and block fall into lacustrine sediments, the underlying sediments are often pushed upwards to form injection structures involving upward movement and drag or fold deformation of the lacustrine strata (Qiao Xiufu et al., 2011). Similar flame structures are induced by current drag, where the deformation was directional, sometimes indicating the direction of the paleocurrent (Suter et al., 2017). Similar flame structures are induced by current drag, where the deformation was directional, sometimes indicating the direction of the paleocurrent (Suter et al., 2011). Deformation features related to cryoturbation are commonly seen as upward–injection structures in vertical sections and as fissure polygons with typical diameters of 40–50 cm in horizontal section. The wings of such structures are symmetrical, and the overburden is mainly coarse gravel (Vandenbergh, 1992; Harris et al., 2000; Petera–Zganiacz and Dzieduszyńska, 2017). The flame structures at Shawan are obviously different from typical structures that were induced by porewater pressure controls on subglacial sediments (Ravier et al., 2014). Therefore, the flame structures at Shawan are typical of those that occur in tectonically active regions, where they are always associated with load structures (Fig. 5) and certain triggering factors such as seismic shaking (e.g., Visher and Cunningham, 1981; Li Shanshan et al., 2008; Jiang et al., 2014, 2016).

4.3 Microfaults

The Shawan section contains a microfault, at depths of 4.4–4.5 m, which is a high-angle planar normal fault that affected beds thicker than ~20 cm, with offsets of ~10 cm (Fig. 5). Microfaults are seen elsewhere, such as at Diaolin (Xu et al., 2015) and in the Lixian lacustrine sections in the eastern TP (Jiang et al., 2016), in the Tamugach Depression in Mongolia (Wang Huai et al., 2008), the Ordos Basin in China (Shao Xiaoyan et al., 2009), the Huimin Sag in China (Zhang Xiaoli et al., 2006), the Bajo Segura Basin in Spain (Alfaro et al., 2001), and at Vens Lake in France (Petersen et al., 2014). We note that (1) the fault in Fig. 5 must be a synsedimentary feature (the fault disappears upwards and downwards, and is overlain by undisturbed levels) that developed during a stage of brittle deformation, (2) the fault is associated with various other SSD structures (flame and ball–and–pillow structures) within the same heterolithic horizon, (3) there is no evidence of gravitational slides, and (4) one can find secondary faults in the Qianguangqiao and Tuanjiecu lacustrine sections close to Lake Diexi (An Weiping et al., 2008; Ren et al., 2018). Although microfaults are seen in a variety of strata with diverse tectonic settings, in all cases they are interpreted to be a consequence of seismic activity.

4.4 Pseudonodules

Pseudonodules are observed in the Shawan section in heterolithic deposits as circular, elliptical, and irregularly shaped balls at depths of 2.3–2.5 m (Fig. 6), and they occur as isolated balls of dark–grayish fine sand that have dropped into the underlying gray silty–clay. The long axes of the pseudonodules are 1–8 cm long, with short axes of 1–3 cm. Pseudonodules occur in the synrift of the Escucha Formation in eastern Spain (Rodríguez–López et al., 2007), the Raniganj Basin of eastern India (Kundu et al., 2011), the Karoo Basin in South Africa (Oliveira et al., 2011), the Santa Fe–Sopetrán Basin of the northern Colombian Andes (Suter et al., 2011), and the study area of Xinxmocun (Jiang et al., 2014). Kuenen (1965) showed experimentally that an external shock might cause pseudonodules to form. Moretti et al. (1999) suggested that pseudonodules can be generated by purely sedimentary processes, and thus are not diagnostic of seismic activity. In the Xinxmocun section beside Diexi Lake, pseudonodules are found at depths of 2.3 and 5.3 m, and although some are still connected with the parent layer, others are completely detached, indicating that the overlying sand layer descended into the underlying silty–clay layer (Jiang et al., 2014). We note that (1) pseudonodules are located in interbedded sand–mud layers, and the overlying and underlying layers are not deformed, (2) detached pseudonodules are consistent with the lithology of the overlying sand layer, (3) and they are similar to those observed at other sites. The association of pseudonodules with various other SSD structures within the same heterolithic horizon suggests their seismic origin (e.g., Rodríguez–López et al., 2007; Kundu and Goswami, 2008), similar to the clastic gravels in the Lixian lacustrine section in the eastern TP (Jiang et al., 2016).

4.5 Ball–and–pillow structures

Ball–and–pillow structures are universal SSD structures that have been observed in the Sao Luis and Grajau basins in northern Brazil (Rossetti and Goes, 2000; Rossetti, 2002), the southeastern Altai (Deev et al., 2009), southern Spain (Alfaro et al., 2010), the Neuquén Basin in northern Patagonia (Moretti and Ronchi, 2011), the Wuqia pull–apart basin in Xinjiang Province (Qiao Xiufu et al., 2012), and the Lixian lacustrine section in the eastern TP (Jiang et al., 2016). At Shawan, the ball–and–pillow structures occur in the middle part of the Shawan section intercalated with undeformed sediments at depths of 4.2–4.45 m, over a thickness of 25 cm (Figs. 3, 7). It is the most common and intense of the deformation structures observed in the lacustrine sediments in the eastern TP (Jiang et al., 2016; Zhong Ning, 2017b), as manifested by (1) being highly obvious, (2) the deformation being intense, and (3) their presence at several stages of deformation. The lithology hosting these structures is dominated by light–yellow clays to silty–clays made up of several depositional beds with thicknesses varying from millimeters to centimeters. The beds are relatively thick in...
the lower parts and become thinner upwards. It is apparent that these deformation structures are asymmetric and show a distinct inclination (Fig. 7a), which means the possibility of heavy loading from the adjacent hill–slope being responsible for the deformation can essentially be ruled out.

The ball–and–pillow structures in the Shawan section consist predominantly of highly distorted laminated beds. Egg–like balls of various sizes are also seen among these deformed beds, with larger examples being over 10 cm in diameter (Fig. 7b). The shells of these balls are moderately consolidated light–yellow clay to silty–clay. The balls themselves are composed of a lithology distinct from that of the matrix, and they consist of dark–grayish medium–to coarse–grained silt that resembles the lithology above the ball–and–pillow structures. This relationship implies that the ball–and–pillow structures were sourced from this overlying layer (Fig. 7b).

The concave–upwards structures in the uppermost deformed beds contain multiple thin beds of clay or silty–clay, which are fainter and closely piled upwards (Fig. 7c). These structures clearly wrap much thinner beds of dark–grayish medium–to coarse–grained silt, and they represent transitional or incomplete ball–and–pillow structures (Fig. 7c). Finer–grained underlying sediment liquefied first, escaped upwards, and then flowed laterally to create these structures. They then intruded and filled low–lying areas to form comparatively thin layers, thus generating ball–and–pillow structures with multiple thin layers of silty–clay to silt–sized material (Fig. 7c).

In tectonically active regions, similar structures to these ball–and–pillow structures have been attributed by most researchers to liquefaction and fluidization processes caused by seismic activity (e.g., Owen, 1996; Molina et al., 1998; Moretti et al., 1999; Qiao Xiufu and Li Haibing, 2008b, 2009; Tian Hongshui et al., 2016a, b; Zhong Ning et al., 2017a). Qiao Xiufu and Li Haibing (2008b) described the deformation process as follows. Fine–sand units display the features of active liquefied deformation while the overlying sand shows mainly the features of passive deformation (hydraplastic deformation). The superposition of coarse–grained sand (heavy) on fine–grained sand (light) represents a driving–force system related to gravitational instability. The heavy sand may sink into the light sand forming load casts and ball–and–pillow structures that retained the primary laminations of the overlying coarse sand. The ball–and–pillow structures are located in different positions within the fine sand, which indicates they dropped through the fine sand unit. The liquefaction and fluidization were triggered by a strong paleo–seismic event.

4.6 Liquefied breccia

Liquefied breccia is a type of deformation structure that
is commonly seen in SSD structures, and they show obvious signs of upwards or downwards penetration (He and Qiao, 2015). They have been observed in the northwestern Red Sea, Egypt (Plaziat et al., 1990), the synsedimentary strike–slip basin of Upper Benue, Nigeria (Guiraud and Plaziat, 1993), the Tarim Basin (Qiao Xiuju and Li Haibing, 2009), and the Longmenshan foreland basin of eastern Tibet in China (Qiao Xiuju et al., 2007, 2012, 2013). The Shawan lacustrine section contains one liquefaction–breccia layer between two undeformed layers, at depths of 2.63–2.69 m (Fig. 8). Liquefied breccia in gray sand, clay, and sand layers shows obvious upwards or downwards penetration of each other, forming edges and corners, and resulting in diverse types of brecciated structure. The scale of deformation is relatively small, with the long axes of breccia clasts typically measuring 1–2 cm, with a minimum length of only a few millimeters. The liquefied–breccia layer at Shawan is about 6 cm thick. Liquefied breccias formed by the intrusion of liquefied sand veins with thixotropic properties that tear soft overlying and underlying sedimentary layers during emplacement. These breccias formed by liquefied flow triggered by seismic activity (Qiao Xiuju et al., 2012, 2013; Gong Zheng et al., 2013; He Bizhu and Qiao Xiuju, 2015), also can see in the Wenchuan earthquake Fault Scientific Drilling project–hole, eastern TP (Li et al., 2013, 2014, 2016).

4.7 Liquefied convolute structures

Liquefied convolute structures in clastic sediments have received growing attention over the past two decades (e.g., Feng Xianyue, 1989; Rodriguez–Pascua et al., 2000; Qiao Xiuju et al., 2007, 2013; Qiao Xiuju and Li Haibing, 2009; Wang et al., 2011; He Bizhu and Qiao Xiuju, 2015).
They are seen in two beds in the Shawan section, at depths of 4.2–4.5 and 5.0–5.2 m, between two undeformed layers (Fig. 9). Their structure implies upwards or downwards emplacement, with the strata exhibiting diverse (i.e., vertical, horizontal, and oblique) orientations. Multiple small, irregular, liquefied veins are seen inside the convoluted bodies, and the silty–clay layer is deformed. These structures are 0.2–0.5 m thick and 0.5–1.5 m wide.

Possible triggering mechanisms for these convolute structures in the lacustrine sediments of the Shawan outcrop are: (1) channel erosion, (2) flash floods, and (3) slope failure. 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 below the sedimentation base (Kundu et al., 2011). Convolute structures caused by flash floods occur in deformed layers that are restricted within parallel and distinct variation in grain sequence, and the magnitude of deformation decreases downwards (Rana et al., 2016). Convolute structures caused by landslides have fold–axes in particular orientations that are controlled by the prevailing physical conditions, whereas convolute structures driven by seismic activity have chaotic orientations of the fold axial planes (Qiao Xiufu and Li Haibing, 2009). With regard to the convolute structures at Shawan, we note that (1) they involve deformed fine–grained sediments within undeformed beds, (2) they are associated with various other SSD structures (load casts, liquefied dykes) within the same heterolithic horizon, and (3) evidence of gravitational slides, channel erosion, and flash floods is absent. We suggest, therefore, that the formation of these convolute structures can be attributed to liquefaction caused by seismic activity in a similar fashion to those described elsewhere (e.g., Feng Xianyue, 1989; Qiao Xiufu and Li Haibing, 2009; Qiao Xiufu et al., 2013).

Interestingly, two contorted laminae observed in proglacial lacustrine sediments at Goting in the Higher Central Himalaya were formed at 18 and 19.5 ka (Juyal et al., 2009), and these times correspond to ephemerally strengthened East Asian summer monsoon events (Wu Jiangying et al., 2009). Taking all available evidence from continental and oceanic sediments into consideration, we suggest that a forcing mechanism behind those events would have been a positive feedback of the tropical Pacific Super–ENSO cycles in response to precessional changes in solar irradiation (Wu Jiangying et al., 2009).

The ages of sediments in the Shawan section range from 17 to 15 cal ka BP, corresponding to the substantial transition during the last deglaciation, which is reflected by changes in vegetation type (Wang Fubao et al., 1996; Ji et al., 2005; Sun et al., 2010; Herzschuh et al., 2014), glacier advance (Wang et al., 2002; Owen et al., 2003; Bae Seong et al., 2009; Nishimura et al., 2014), fluctuating lake levels (Wang et al., 2002; Liu et al., 2008; Mischke et al., 2010; Nishimura et al., 2014) and sea levels (Zhao et al., 2018),
and increases in precipitation (Herzschuh et al., 2014). Moreover, this transformation event has different degrees of expression on the eastern, northeastern, southern, and northwestern margins of the Tibetan Plateau (Table 2). The Shawan section records large and strong SSD structures during this transitional period, possibly caused by strong seismic activity triggered by an increased influence of precipitation on the permeability of fault structures during this transitional period, possibly caused by strong seismic activity (Yu Song, 2010; Wang et al., 2011). Therefore, we collected available data on earthquake-generated load and discussed above, the formation of the Shawan SSD structures were not influenced by slope-loading, overloading, rapid sedimentation, slope-failure, channel erosion, storm waves, permafrost, or overriding ice dynamics effects, which indicates they were a consequence of seismic activity.

It is difficult to determine the relationship between the SSD structures and earthquake magnitudes. Many researchers (e.g., Youd, 1977; Allen, 1986; Scott and Price, 1988; Galli, 2000) 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 is generally insufficient to cause liquefaction. Moreover, these general estimates of possible earthquake magnitudes from SSD structures and the interpretation of earthquake magnitudes also depend on the local geographic position, the tectonic setting, the sedimentary environment, and the sedimentary age, and other things. Earthquakes with shallow hypocentral depths (i.e., <50 km) can cause more severe shaking close to the epicenter than a deeper earthquake, and they can thus result in more liquefaction structures. However, at the epicenter, seismic shaking may be less severe for a shallow earthquake than for a deeper one, and so deeper earthquakes may generate liquefaction at greater distances from the epicenter (Obermeier, 1996).

Load and ball- and pillow structures are the most commonly observed SSD structures in the Shawan section (Yu Song, 2010; Wang et al., 2011). Therefore, we collected available data on earthquake-generated load and

<table>
<thead>
<tr>
<th>Study site</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation (m a.s.l.)</th>
<th>Grain size, C, Fe/Mn, TOC, C/N, pollen</th>
<th>AMS 14C</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pamoyou Co (S)</td>
<td>33.80</td>
<td>90.22</td>
<td>5010</td>
<td>Ca, Fe/Mn, TOC, C/N, pollen</td>
<td>AMS 14C</td>
<td>Wang et al., 2011</td>
</tr>
<tr>
<td>Lake Goting (S)</td>
<td>30.82</td>
<td>90.23</td>
<td>5030</td>
<td>Sand, TOC, δ13C, CO2, pollen</td>
<td>AMS 14C</td>
<td>Nishimura et al., 2014</td>
</tr>
<tr>
<td>Lake Zabuye (S)</td>
<td>31.35</td>
<td>84.07</td>
<td>4421</td>
<td>δ18O, δ13C, TOC, TIC, TS</td>
<td>AMS 14C</td>
<td>Wang et al., 2002</td>
</tr>
<tr>
<td>Lake Naleng (SE)</td>
<td>31.10</td>
<td>99.75</td>
<td>4200</td>
<td>Pollen</td>
<td>AMS 14C</td>
<td>Kramer et al., 2010</td>
</tr>
<tr>
<td>Ruoergai Basin (E)</td>
<td>31.85</td>
<td>101.52</td>
<td>3400-3700</td>
<td>Pollen</td>
<td>AMS 14C</td>
<td>Wang et al., 1996</td>
</tr>
<tr>
<td>Ximencuo (E)</td>
<td>33.38</td>
<td>101.10</td>
<td>4000</td>
<td>Grain-size, Y/Al, pollen</td>
<td>AMS 14C</td>
<td>Herzschuh et al., 2014</td>
</tr>
<tr>
<td>Lake Koucha (NE)</td>
<td>34.00</td>
<td>97.20</td>
<td>4540</td>
<td>Pollen</td>
<td>AMS 14C</td>
<td>Herzschuh et al., 2009</td>
</tr>
<tr>
<td>Lake Kuhai (NE)</td>
<td>35.31</td>
<td>99.18</td>
<td>4150</td>
<td>Pollen</td>
<td>AMS 14C</td>
<td>Mischke et al., 2010</td>
</tr>
<tr>
<td>Lake Qinghai (NE)</td>
<td>13.80</td>
<td>100.52</td>
<td>3200</td>
<td>Pollen</td>
<td>AMS 14C</td>
<td>An et al., 2012</td>
</tr>
<tr>
<td>Lake Chaka Salt (NE)</td>
<td>13.52</td>
<td>100.14</td>
<td>3194</td>
<td>Pollen</td>
<td>AMS 14C</td>
<td>Liu et al., 2003</td>
</tr>
<tr>
<td>Guliya (NW)</td>
<td>35.28</td>
<td>81.48</td>
<td>6710</td>
<td>Dust, δ13C, NH₃, NO₃, δ18O, CH₄</td>
<td>CRN, OSL</td>
<td>Thompson et al., 1997</td>
</tr>
<tr>
<td>Nam Co (S)</td>
<td>30.75</td>
<td>90.78</td>
<td>4720</td>
<td>Ca, Fe/Mn, TOC, C/N, pollen</td>
<td>AMS 14C</td>
<td>Wang et al., 2015</td>
</tr>
<tr>
<td>Lake Ata and Kongur Shan (NW)</td>
<td>38.70</td>
<td>75.02</td>
<td>&gt;3500</td>
<td>Glacier advance</td>
<td>10Be TCN</td>
<td>BaeSeong et al., 2009</td>
</tr>
</tbody>
</table>
ball–and–pillow structures produced in various sedimentary environments around the world (Table 3). Our statistical results show that load and ball–and–pillow structures represent a minimum earthquake magnitude of 5 and a maximum magnitude of more than 8, and typically correspond to magnitudes 6–7. Grimm and Orange (1997) suggested that deep–water lake sediments have stronger cohesive forces and shear strengths than do coarse sediments deposited in shallow water. Therefore, when subjected to the same amount of earthquake shaking, SSD structures should be expected to occur first in shallow water. For a given intensity of load and ball–and–pillow structure deformation, it then follows that a larger earthquake magnitude is required to generate these structures in lacustrine sediments than in fluvio–lacustrine or marine sediments. Too few examples of SSD structures are available for pluvial and fluvial facies to be compared with other environments. We applied the relationship between seismic intensity and thickness of the contorted bedding horizon (Hibsch et al., 1997) to the 25–50 cm thick horizon of contorted bedding in the Shawan section with load and ball–and–pillow structures, and the corresponding seismic intensity is IX–X (Table 4). We also applied the relationship between earthquake

Table 3 Load and ball-and-pillow structures resulting from seismic activity in various sedimentary environments

<table>
<thead>
<tr>
<th>Study location</th>
<th>Sedimentary environment</th>
<th>Lithology</th>
<th>Age</th>
<th>Distance to fault (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sickierki</td>
<td>Lacustrine</td>
<td>Clay, silty clay</td>
<td>Late Pleistocene</td>
<td>&gt; 10</td>
</tr>
<tr>
<td>Xinnocun</td>
<td>Lacustrine</td>
<td>Clay, silty clay, fine sand</td>
<td>Late Pleistocene</td>
<td>&lt; 10</td>
</tr>
<tr>
<td>Hazar lake</td>
<td>Lacustrine</td>
<td>Clay, silty clay, fine sand</td>
<td>Holocene</td>
<td>&lt; 10</td>
</tr>
<tr>
<td>Shawan</td>
<td>Lacustrine</td>
<td>Clay, silty clay, fine sand</td>
<td>Late Pleistocene</td>
<td>&lt; 10</td>
</tr>
<tr>
<td>Shawan</td>
<td>Lacustrine</td>
<td>Clay, silty clay, fine sand</td>
<td>Late Pleistocene</td>
<td>&lt; 5</td>
</tr>
<tr>
<td>Lixian</td>
<td>Lacustrine</td>
<td>Clay, silty clay, fine sand</td>
<td>Late Pleistocene</td>
<td>20</td>
</tr>
<tr>
<td>Zaguan River</td>
<td>Lacustrine</td>
<td>Clay, silty clay, fine sand</td>
<td>Late Pleistocene</td>
<td>20</td>
</tr>
<tr>
<td>Issyk-Kul lake</td>
<td>Lacustrine</td>
<td>Clay, silty clay, fine sand</td>
<td>Late Pleistocene</td>
<td>30</td>
</tr>
<tr>
<td>Wujia</td>
<td>Lacustrine</td>
<td>Siltstone</td>
<td>Middle Jurassic</td>
<td>50</td>
</tr>
<tr>
<td>Northern Patagonia</td>
<td>Fluvio-lacustrine</td>
<td>Silt, fine sand, coarse sand</td>
<td>Late Pleistocene</td>
<td>30–40</td>
</tr>
<tr>
<td>Southern Alps</td>
<td>Fluvio-lacustrine</td>
<td>Silt, fine sand, medium sand</td>
<td>Early Cambrian</td>
<td>&lt; 20</td>
</tr>
<tr>
<td>Kathmandubasin</td>
<td>Fluvio-lacustrine</td>
<td>Clay, silt, sand, gravel</td>
<td>Holocene</td>
<td>&lt; 100</td>
</tr>
<tr>
<td>Laiyuan</td>
<td>Marine</td>
<td>Limestone</td>
<td>Mesoproterozoic</td>
<td>QiaoXiufu and Gaolinzhi, 2007</td>
</tr>
<tr>
<td>Baltic Sea</td>
<td>Marine</td>
<td>Silt, sand</td>
<td>Late Pleistocene</td>
<td>&lt; 20</td>
</tr>
<tr>
<td>Sa AoLao/As Basin</td>
<td>Marine</td>
<td>Fine sand, sandstone</td>
<td>Cretaceous</td>
<td>&lt; 30</td>
</tr>
<tr>
<td>Northwestern Estonia</td>
<td>Marine</td>
<td>Siliciclastic sandstone</td>
<td>Middle Ordovician</td>
<td>Põldsaar and Ainsaar, 2014</td>
</tr>
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<td>Longmenshan</td>
<td>Marine</td>
<td>Sandstone</td>
<td>Upper Triassic</td>
<td>50</td>
</tr>
<tr>
<td>Zhucheng</td>
<td>Fluvial</td>
<td>Sandstone</td>
<td>Cretaceous</td>
<td>&lt; 20</td>
</tr>
<tr>
<td>East Kunlun</td>
<td>Fluvial</td>
<td>Sand, gravel</td>
<td>Modern</td>
<td>&lt; 100</td>
</tr>
</tbody>
</table>

Table 4 Earthquake magnitudes inferred from Shawan soft-sediment deformation structures

<table>
<thead>
<tr>
<th>Layer</th>
<th>Type</th>
<th>Estimated magnitude</th>
<th>Deformation type</th>
<th>Contorted bedding horizon</th>
<th>Accumulation sand layer</th>
<th>Maximum liquefied distance</th>
<th>Empirical formula</th>
</tr>
</thead>
<tbody>
<tr>
<td>SSD1</td>
<td>Pseudonodules</td>
<td>&gt; 5</td>
<td>6–7</td>
<td>10</td>
<td>VIII–IX</td>
<td>20</td>
<td>&gt; 6</td>
</tr>
<tr>
<td>SSD2</td>
<td>Pseudonodules</td>
<td>&gt; 5</td>
<td>6–7</td>
<td>15</td>
<td>VII–IX</td>
<td>&gt; 5</td>
<td>5.6</td>
</tr>
<tr>
<td>SSD3</td>
<td>Liquefied breccia</td>
<td>&gt; 5</td>
<td>6–7</td>
<td>6</td>
<td>VII–VII</td>
<td>2</td>
<td>&gt; 4</td>
</tr>
<tr>
<td>SSD4</td>
<td>Load casts</td>
<td>&gt; 5</td>
<td>5–6</td>
<td>25</td>
<td>VIII–X</td>
<td>20</td>
<td>&gt; 6</td>
</tr>
<tr>
<td>SSD5</td>
<td>Load casts</td>
<td>&gt; 5</td>
<td>6–7</td>
<td>50</td>
<td>X–XI</td>
<td>&gt; 100</td>
<td>&gt; 8</td>
</tr>
<tr>
<td>SSD6</td>
<td>Flame structure</td>
<td>&gt; 5</td>
<td>6–7</td>
<td>50</td>
<td>X–XI</td>
<td>&gt; 100</td>
<td>&gt; 8</td>
</tr>
<tr>
<td>SSD7</td>
<td>Ball-and-pillow</td>
<td>&gt; 5</td>
<td>6–7</td>
<td>25</td>
<td>IX–X</td>
<td>&gt; 100</td>
<td>&gt; 8</td>
</tr>
<tr>
<td>SSD8</td>
<td>Liquefied convolutes</td>
<td>&gt; 5</td>
<td>6–7</td>
<td>50</td>
<td>X–XI</td>
<td>&gt; 100</td>
<td>&gt; 8</td>
</tr>
<tr>
<td>SSD9</td>
<td>Microfault</td>
<td>&gt; 5</td>
<td>6–7</td>
<td>20</td>
<td>VIII–X</td>
<td>12</td>
<td>&gt; 6</td>
</tr>
<tr>
<td>SSD10</td>
<td>Liquefied convolutes</td>
<td>&gt; 5</td>
<td>6–7</td>
<td>20</td>
<td>VIII–X</td>
<td>8</td>
<td>&gt; 5</td>
</tr>
</tbody>
</table>
magnitude and the accumulated thickness of the sand layer (Zhong Ning, 2017) to the 20–100 cm thick sand layer that contains load and ball–and–pillow structures in the Shawan section, and the corresponding earthquake magnitude is >6.6 and >8.0 (Table 4). The load and ball–and–pillow structures in the Shawan section may therefore represent earthquake activity with magnitudes 6–7.

Besides seismic intensity, the distance from the earthquake focus is also an important factor when estimating earthquake magnitudes from SSD structures. Historical data suggest that 79% of liquefaction occurs within 30 km of the earthquake epicenter (Galli, 2000), and liquefaction at distances greater than 15 ± 20 km from the epicenter requires strong earthquakes with magnitudes larger than 6 (e.g., Sims, 1975; Mohindra and Bagati, 1996; Galli, 2000). In Italy, Galli and Ferreli (1995) analyzed 12,880 examples of liquefaction structures generated by historical and instrumented earthquakes over the past 158 years, and they found that 95% of these structures were generated within a 25 km radius of the epicenter of the causative shallow earthquakes.

Paleoearthquake magnitudes can be calculated on the basis of instrumentally recorded occurrences of liquefaction of sand layers induced by previous and recent earthquakes. Kuribayashi and Tatsuoka (1975) were commended for their most useful compilation of the occurrences and distribution of liquefaction during

Fig. 9. Liquefied convolute structures in the Shawan lacustrine section.
Japanese earthquakes of the past century (1872–1968), and they produced the first map to show 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. Youd (1977) compiled data for 14 earthquakes (M >5.3) from 1897–1976 in the USA, India, New Zealand, and Chile, and 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 water-jet and sand-emit data for the 900 years before 1995, and they established the relationship between magnitude and maximum epicentral distance to liquefied deformation in China (Fig. 10a). Subsequent revision and provision of further data by many researchers has now resulted in a reasonable, widely-used diagram of SSD structures and their relationships to earthquake magnitude (Ambraseys, 1988; Obermeier, 1996, 1998; Obermeier et al., 2002; Qiao Xiufu et al., 2017).

The two relationships between earthquake magnitude and distance from the epicenter at which liquefaction occurs, as given by Liu Ying and Xie Junfei (1984) and Qiao Xiufu et al. (2017), are shown in Fig. 10a and 10b. We used first the magnitude data listed in Table 3 to establish the relationship between magnitude and epicentral distance or maximum liquefaction distance. Our results show that the load and ball-and-pillow structures at Shawan correspond to earthquakes with a magnitude of 6–7 (61% of the total, N = 18) at epicentral distances of 20–70 km; the correlation coefficient is 0.896 (Fig. 11a). We also used the epicentral distance data listed in Table 3 to establish the relationship between epicentral distance or maximum liquefaction distance and magnitude. These results show that 67% (N = 15) of the total events in the catalog correspond to magnitude 6–7 earthquakes with epicentral distances of 20–50 km; the correlation coefficient is 0.892 (Fig. 11b). It is possible to assess the epicentral location or the maximum liquefied distance by using the earthquake magnitude determined from the intensity of SSD structures. The vertical distance of SSD structures from the seismogenic fault can also be used to estimate the earthquake magnitude. When load and ball-and-pillow structures are induced by seismic activity, they imply earthquake magnitudes of 6–7 and epicentral distances of 20–70 km. The distance of the Shawan section to the Minjiang fault and/or the Songpinggou fault is <5 km (Fig. 10), corresponding to possible magnitudes of 5.8–6.8. Nevertheless, as described above, the lacustrine sediments examined in our study have SSD features and sedimentary environments that are similar to those in other regions. This similarity supports the hypothesis that the size and type of a seismite may be a proxy for the relative intensity of seismic shaking, so that a larger or thicker structure probably indicates a larger-magnitude earthquake.

Our study area is located on the eastern margin of the TP. Crustal shortening is an important process that is responsible for the high topography of the Longmenshan region on the eastern flank of the plateau (Hubbard and Shaw, 2009; Xu et al., 2009). The shortening is not only important for the accumulation of energy along the eastern margin of the plateau, but also responsible for generating strong and frequent earthquakes with magnitudes larger than 7 along the active margin of the Bayan Kala fault–block in the eastern TP (Deng Qidong et al., 2014). Since 1725, at least eight earthquakes larger than magnitude 6 have occurred on the 350–km-long Xianshuifu fault (Allen et al., 1991). Similarly, at least eight earthquakes larger than magnitude 6 have occurred along the Minshan fault since 1933 (Fig. 1). Among the many types of SSD structures in the region, the Shawan section probably records the most intense examples of liquefaction, and so probably corresponds to the location of...
that experienced the most severe ground motions during deposition of the sediments (Wang et al., 2011). Similarly, a variety of SSD structure types in one deformation layer ought to indicate the most intense liquefied deformation. Considering that the liquefaction strength of flame structures may be greater than that of load structures, and that pseudonodules, microfaults, liquefied breccia, and liquefied convolute structures may be stronger than ball–and–pillow structures, we estimate that the Shawan lacustrine section documents six strong earthquakes of magnitude 6–7 and one >7 (Table 4).

6 Discussion of the Tectonic Activity of Faults

The 1933 Ms 7.5 and 1713 Ms 7.0 earthquakes occurred in the Diexi area (Fig. 12). The generation of the 1933 Ms 7.5 Diexi earthquake have been ascribed to four different faults: (1) the E–W–trending Canlingshan fault (Chang Longqing, 1938), (2) the NW–SE–trending Songpinggou fault (Tang Rongchang et al.,1983; Huang Zuzhi et al.,2002; Ren et al.,2018), (3) the N–S–trending Minjiang fault (Chen et al., 1994; Deng Qidong et al.,1994; Qian Hong et al.,1999; Wang Kang and Shen Zhenkang, 2013), and (4) a buried ramp–type thrusting fault (Zhang Yueqiao et al., 2016). The Canlingshan fault trends E–W, extends for almost 1 km, has a breaking distance of 25–36 m, and forms a 60–70 m wide rift. The Canlingshan fault is a bedrock fault that formed when the earthquake struck, and it cannot be the seismogenic structure responsible for the Diexi earthquakes.

Previous work on the Minjiang fault has been focused either on the strata involved, landscapes, terraces, and hot springs along the fault with their associated calcium accumulations (Kang Wenqing et al., 2004), or GPS observations, from which it is speculated that the horizontal slip rate is 1–2 mm/a (Zhao Xiaolin et al., 1994; Kang Wenqing et al., 2004) and the vertical slip rate 0.37–0.53 mm/a (Zhou Rongjun et al., 2000; Zhang Junlong et al., 2013). Recently, one large trench was excavated along the northern segment of the Minjiang fault and a record of three paleoearthquake events was revealed (Zhang Junlong...
et al., 2013). However, these studies have been focused mostly on the northern segment of the fault. The southern segment of the Minjiang fault crosses the Diexi area, but hitherto there has been no direct seismicity or geomorphic evidence to establish the nature of activity on this part of the fault. In addition, the inferred surface rupture of the 1933 Diexi earthquake indicates an extensional setting, which is inconsistent with the proposed sense of motion along the Minjiang fault (reverse slip with a component of right-lateral strike-slip) (Chen et al., 1994).

Based on field investigations of morpho-structures around the dammed lake in the Diexi area, and faults affecting the latest Pleistocene lacustrine sediments, and by taking into account historical earthquakes and paleoearthquake studies in this zone, Zhang Yueqiao et al. (2016) proposed an alternative view of the Diexi earthquake seismogenic structure, which was that it was analogous to the 2013 Lushan Ms 7.0 earthquake that ruptured the southern segment of the Longmenshan fault zone. Their model of a buried ramp-type thrust proposes a W-dipping ramp at depths of 10–15 km beneath the Minjiang River that thrusts eastwards and produces repeated earthquakes in this deeply incised valley. However, this view awaits further testing by geophysical and geodetic investigations.

Recently, based on an investigation of surface ruptures, an analysis of the distribution of disasters, and trench research, Ren et al. (2018) proposed that the Songpinggou fault trace represents the surface rupture of the 1933 Diexi earthquake for the following reasons. (1) The Songpinggou fault is located in the mezoisoseismal area, and the isoseismal contours trend northwest. (2) Casualties along the Songpinggou valley were roughly twice those along the Minjiang valley. (3) Radiocarbon dating indicates that the last surface-rupture event occurred at c. 1760–1960 AD, overlapping with the 1933 Diexi earthquake.

The Shawan section is located at the intersection between the Songpinggou and Min River faults. We infer that the 1933 Diexi earthquake resulted from slip along the northwest-trending Songpinggou fault. Given that the Shawan section documents seven M ~7 earthquakes, we conclude that the observed SSD structures record deglacial activity along the Songpinggou fault.

7 Conclusions

Based on the soft-sediment deformation (SSD) structures and chronological results in a well-exposed 7.0 m-thick lacustrine sequence at Shawan, integrated with the regional geology, we briefly summarized the last deglacial tectonic active of the Diexi eastern Tibetan Plateau and reached the following conclusions:

(1) The structures are distributed over seven stratigraphic levels and comprise load casts, flame structures, pseudonodules, ball- and pillow structures, liquefied convolute structures, liquefied breccia, and microfaults. Most of the SSD structures indicate deformation mechanisms related to liquefaction and/or fluidization processes, probably triggered by paleoseismic events.

(2) Integrating the data on the type, form, maximum liquefaction distance, and thickness distributions of these SSD structures in the Shawan section, we determined that liquefaction deformation was extremely intense, and likely records six strong earthquakes of magnitude 6–7 and one over 7.

(3) The Songpinggou fault is thought to represent the seismogenic structure that resulted in the 1933 Ms 7.5 Diexi earthquake. As the Shawan section is located at the intersection between the Songpinggou and Minjiang faults and records seven M ~7 earthquakes, we infer that the investigated SSD structures may record deglacial activity along the Songpinggou fault.

Acknowledgements

We thank Drs. QIAO Xiufu, HAN Fei, WANG Hu, SHAO Zhigang, WANG Ping and ZHANG Huiping for helpful discussions that improved the manuscript. We gratefully acknowledge the joint support by the National Natural Science Foundation of China (41807298, 41672211, 41572346), and the Special Project of Fundamental Scientific Research of the Institute of Geology, China Earthquake Administration (IGCEA1713).

Manuscript received July 10, 2018
accepted Oct. 27, 2018
associate EIC HAO Ziguo
edited by LIU Lian

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