Earthquake-induced soft-sediment deformations and seismically amplified erosion rates recorded in varved sediments of Köyceğiz Lake (SW Turkey)

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Key Points:

- Past earthquakes are recorded in the varved sequence as soft-sediment deformations and amplified erosion rates
- Minimum peak ground acceleration (PGA) required to induce soft-sediment deformations is found to be 20 cm/s^2
- Minimum PGA required to amplify erosion rates is found to be 70 cm/s^2

Abstract

Earthquake-triggered landslides amplify erosion rates in catchments, i.e. catchment response to seismic shocks (CR). In addition to historical eyewitness accounts of muddy rivers

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implying CRs after large earthquakes, several studies have quantitatively reported increased sediment concentrations in rivers after earthquakes. However, only a few paleolimnological studies could detect CRs within lacustrine sedimentary sequences as siliciclastic-enriched intercalations within background sedimentation. Since siliciclastic-enriched intercalations can easily be of non-seismic origin, their temporal correlation with nearby earthquakes is crucial to assign a seismic triggering mechanism. In most cases, either uncertainties in dating methods or the lack of recent seismic activity has prevented reliable temporal correlations, making the seismic origin of observed sedimentary events questionable. Here, we attempt to remove this question mark by presenting sedimentary traces of CRs in the 370-year-long varved sequence of Köyceğiz Lake (SW Turkey) that we compare with estimated peak ground acceleration (PGA) values of several nearby earthquakes. We find that earthquakes exceeding estimated PGA values of ca. 20 cm/s² can induce soft-sediment deformations (SSD), while CRs seem only to be triggered by PGA levels higher than 70 cm/s². In Köyceğiz Lake, CRs produce Cr- and Ni-enriched sedimentation due to the seismically mobilized soils derived from ultramafic rocks in the catchment. Given the varve chronology, the residence time of the seismically mobilized material in the catchment is determined to be 5 to 10 years.

Keywords: Quantitative lacustrine paleoseismology, Fethiye-Burdur Fault Zone, soft-sediment deformation, catchment response, peak ground acceleration, ITRAX μ XRF scanner.

1. Introduction

Paleoseismic information from lake sediments has been commonly revealed through earthquake-triggered soft-sediment deformations [e.g., *Migowski et al.*, 2004; *Monecke et al.*,

2004, 2006; Agnon *et al.*, 2006; Beck, 2009; Kagan *et al.*, 2011] and subaqueous mass-wasting deposits [e.g., Schnellmann *et al.*, 2006; Waldmann *et al.*, 2011; Moernaut *et al.*, 2007, 2014; Van Daele *et al.*, 2015; Avşar *et al.*, 2015]. Recently, the potential of another tectono-sedimentary process has been realized. Earthquake-triggered landslides are expected to increase erosion rates, and hence sediment fluxes to basins. It was empirically shown that earthquakes having magnitudes as low as $M_w = 4.5$ can trigger landslides, and there is a strong power-law correlation between the moment magnitude and the total triggered landslide volume [e.g., Keefer, 1994; Malamud *et al.*, 2004]. Eyewitness accounts of muddy rivers after large historical earthquakes [e.g., Ambraseys, 2009] probably indicate amplified erosion rates due to earthquake-triggered landslides in catchments. It was also found that suspended sediment concentrations in rivers increased 3-7 times after the 1999 M_w 7.6 Chi-Chi earthquake, and returned to preseismic levels within ca. 6 yr [Dadson *et al.*, 2004; Yanites *et al.*, 2010; Hovius *et al.*, 2011]. Similarly, in a more recent study by Wang *et al.* [2015], the suspended sediment concentration also increased 3-7 times after the 2008 M_w 7.9 Wenchuan earthquake in the Longmen Shan, China. However, they found a wide range of time (from 1.2 ± 0.9 yr to 90 ± 65 yr) required for sediment concentration to return to its preseismic levels.

These observations have encouraged subaqueous paleoseismologists to investigate sedimentary records of seismically amplified erosion rates (i.e., catchment responses; CRs). Howarth *et al.* [2012, 2014; 2015], for example, illustrated CRs as siliciclastic enrichments within the background sedimentation of lakes located on the Alpine Fault, New Zealand. Furthermore, Avşar *et al.* [2014] temporally correlated the siliciclastic-enriched intercalations with the large historical earthquakes of the North Anatolian Fault, Turkey. Since these kind of intercalations can also be due to sediment focusing by wave/wind action, flooding, deforestation, etc., their temporal correlation with “several” earthquakes in the region is crucial in attributing them to seismic shocks. This validation can only be done if both; 1) the

sediments are precisely dated, and 2) there are several precisely recorded earthquakes in the region. However, most studies in the literature suffer from either uncertainties in sediment chronologies or lack of recent and reliable earthquake records. Combining a 370-year-long continuous varve chronology with the records of numerous instrumental/historical earthquakes, we present here the evidences of CRs in the sedimentary sequence of Köyceğiz Lake in SW Turkey.

2. Study Site

Köyceğiz Lake (53 km²) is located ca. 30 km to the west of the Fethiye-Burdur Fault Zone (FBFZ, Figure 1a), which is the onland continuation of the subduction transform edge propagator (STEP) fault sinistrally offsetting the Hellenic and Cyprus arcs [Hall *et al.*, 2014]. The bathymetric survey carried out in 2014 revealed that the lake has two basins with maximum depths of 26 and 32 meters, northern and southern basins respectively (Figure 1b). The lake is fed by three main inlets discharging into the northern basin, and connected to the Mediterranean Sea by its outlet at the southern basin. The dominant lithology in the catchment (875 km²) is peridotite, which covers ca. 60% of the catchment. The instrumental period seismicity in the region (Figure 1c) shows a dispersed pattern rather than being localized around individual faults.

The seismic activity during the last century includes eight remarkable events that might have resulted in severe ground motions at the lake site: 1) M_w 5.9 and M_w 5.5 earthquakes in 1959 ca. 3-5 km to the northwest of the coring location; 2) M_w 6.7 and M_w 6.5 earthquakes in 1957 ca. 50 km to the south of the coring location; 3) M_w 6.3 and M_w 6.0 earthquakes in 1941 ca. 56 and 41 km to the northwest of the coring location, respectively; and 4) two M_w 5.3 earthquakes in 1920 and 1921 ca. 15 km to the northeast of the coring location (Figure 1c).

The European Archive of Historical Earthquake Data (AHEAD)

[<http://www.emidius.eu/AHEAD/>; *Stucchi et al.*, 2013] also reports several historical earthquakes in the region during the last 400 years (Figure 1d). Among these earthquakes, the magnitudes and epicentral locations of the 1896, 1874, 1870, 1869 and 1851 earthquakes are relatively well defined, while the others are still debatable.

3. Materials and Methods

A 90 cm-long gravity core was collected from the deepest part (32 m) of the southern basin in April 2014 (28.61° E, 36.89° N, black dot in Figure 1b). A 5 cm-wide and 0.5 cm-thick u-channel was extracted from the core for ITRAX μ XRF scanning, which was done by Mo-tube with exposure time of 10 s for every 200 microns. From the u-channel, 4 cm-wide optical and radiographic images, allowing better evaluation of the sedimentary structures, were obtained by combining the images from three successive scans: one from the middle axis and two more by shifting the ITRAX tray 1 cm left and right. In addition to the radiographic image, the μ XRF data were used to investigate the possible geochemical traces of earthquake-induced changes in sediment flux from the catchment. The varve counting was done using the high-resolution ITRAX μ XRF scanning data and by considering the lateral continuity of the layers in the optical image. Although there are uncertain cases, the majority of the varves in the Köyceğiz sequence are both visually and geochemically discernable. In order to determine the level of uncertainty in the varve chronology, as a first step, the corresponding author determined the maximum possible number of varves (i.e. the older threshold for the chronology) by drawing boundaries on the optical image, where any visual or geochemical implication is seen (Figure S1). Then, eleven colleagues were asked to provide a list of “nondistinct” varves, in order to quantify uncertainties especially at highly deformed levels. In case four or more operators were uncertain about a varve, it was accepted

as “nondistinct”. The younger threshold (minimum number of varves) for the chronology was obtained by excluding the all “nondistinct” varves from the older threshold. The average of the older and younger chronologies was accepted as the final varve chronology. In other words, the nondistinct varves were assigned 0.5 probability of being anomalous. The final varve chronology was validated by ^{137}Cs and excess ^{210}Pb ($^{210}\text{Pb}_{\text{xs}}$) activity measurements on 16 samples (1 cm-thick) by using a low-background, high-efficiency well-type gamma spectrometer, the details of which can be found in *Schmidt et al.* [2014].

The instrumental period earthquake catalogue used in this study (KOERI-NEMC EQ Catalog) does not provide moment magnitudes (M_w) for all of the earthquakes. For this reason, to ensure consistency, we converted all different types of magnitudes in the catalogue (i.e., M_d , M_L , M_s and M_b) to M_w by using the equations proposed by *Akkar et al.* [2010].

Then, to evaluate the level of seismic shaking experienced by the lake site, peak ground acceleration (PGA) estimates were made for both instrumental and historical period earthquakes based on the attenuation relationship by *Akkar and Çağnan* [2010], which was produced by using the Turkish strong motion database containing 1259 records from 573 earthquakes with moment magnitudes ranging between 3.5 and 7.6, and source-to-site distances between 0-200 km.

The degree of deformation on the sediments was quantified by digitizing the layers observed on the radiograph as separate lines and by calculating the mean square error (MSE) of each detrended line. To do so, the radiographic image was first georeferenced according to its true dimensions (i.e., non-earth coordinate system) using a geographic information system software (MapInfo Professional). Next, lines were drawn by following the observable layers on the radiographic image, and then each line was converted to 1-mm-spaced points by using “poly-to-point” feature of the software. In this way, coordinates of the points forming each line were obtained, where x and y coordinates represent the positions of the points along the

width of the radiographic image and depth of the core, respectively. Using these coordinates allow plotting sedimentary layers as quantified curves. In order to avoid biases due to the inclinations of the layers, the curves were detrended and their mean square errors (MSE) were calculated to evaluate the deviations from a horizontal straight line. Hence, the MSE values, being a measure of layer curvature and deformation, were plotted along the core and used as deformation index (DI). Finally, the temporal comparison of the PGA, DI and μ XRF elemental profiles was used to evaluate the effects of earthquakes on sedimentation.

4. Results

Köyceğiz varves are of the carbonaceous-organic type [Zolitschka *et al.*, 2015] in which a yellowish-white carbonate layer precipitated from the water column during late spring and summer is coupled with a runoff-related grayish-brown layer composed of detrital minerogenic and organic matter deposited during the rest of the year. The contrast between the detrital and carbonaceous layers is obvious along the Ca/Ti ratio profile (Figure 2a, and Figure S1 for the complete core). One can assume that Ca roughly represents the carbonate content while Ti represents the detrital minerogenic content of the sediments. Hence, the carbonate-rich layers deposited during spring/summer are represented as significant peaks along the Ca/Ti ratio profile.

Along the ^{137}Cs activity profile (Figure 2b), ^{137}Cs peaks regarding the Chernobyl reactor accident in 1986 and the maximum fallout due to atmospheric nuclear weapon tests in 1963 were detected at depths of 10 and 19.5 cm, respectively. In addition, the onset of ^{137}Cs fallout around 1955 [e.g., Cambray *et al.*, 1989] was detected at 23 cm depth. The slope of the exponential fit along the $^{210}\text{Pb}_{\text{xs}}$ data is calculated as 0.08 with a standard error of 0.011. If the “Constant Flux Constant Sedimentation Rate” model [Goldberg, 1963] is applied, the average sedimentation rate (SR) is calculated as 0.39 ± 0.05 cm/yr. The varve-based age-

depth model plotted together with the ^{137}Cs and $^{210}\text{Pb}_{\text{xs}}$ data (Figure 2c) confirms that grayish-brown/yellowish-white couplets are annually deposited. In the light of twelve independent counting, the maximum error in the varve chronology reaches up to ± 8 years at the bottom of the core, i.e., total of 16 nondistinct varves (See Figure S2 for the complete lists from all operators). Yet, it should be noted that the error in varve chronology for the last 200 years, during which the instrumental and well-defined historical seismicity records are available, does not exceed ± 2 years. The steeper slope towards the upper parts of the varve chronology illustrates compaction effect on sediments.

Numerous studies previously illustrated earthquake-induced soft-sediment deformations that can be used to reveal paleoseismic data. The most commonly observed deformations in lake sediments include micro-faults [e.g., *Seilacher*, 1969; *Monecke et al.*, 2004; 2006; *Beck*, 2009], micro-folds [e.g., *Marco et al.*, 1996; *Monecke et al.*, 2004; 2006], pseudonodules [e.g., *Sims*, 1973; *Rodriguez Pascua et al.*, 2000; *Migowski et al.*, 2004; *Monecke et al.*, 2004], intraclast breccia layers [e.g., *Marco et al.*, 1996; *Migowski et al.*, 2004; *Agnon et al.*, 2006; *Kagan et al.*, 2011], loadcasts [e.g., *Sims*, 1973; *Hibsch et al.*, 1997; *Moretti and Sabato*, 2007], and liquefaction/fluidization/fluid-escape structures [e.g., *Lignier et al.*, 1998; *Chapron et al.*, 2004; *Moernaut et al.*, 2007; 2009; *Beck*, 2009]. In the Köyceğiz record, eight soft-sediment deformation (SSD) zones, which were mainly ductile in nature (i.e., micro-folds), were detected in the sequence (Figure 3; SSD 1-8). In addition to ductile deformations, three micro-faults were observed (F1-F3). F1 leaves the radiographic view at 18.5 cm, making the detection of its upper horizon impossible, whereas F2 and F3 terminate at 21.8 cm and 83.5 cm, respectively. The varve chronology confirms that the instrumental period earthquakes in 1959, 1957, 1941 and 1920-21 induced ductile deformations, where 1957 earthquake also probably triggered the F2 (Figure 3; SSD 2, 3 and 4). The other SSDs imply seismic activities around 2004 (SSD 1), between 1874 and 1852 (SSD 5 and 6), around

1806 (SSD 7) and around 1653 (SSD 8) as well. As provided in Figure 1d, there is a cluster of historical earthquakes between 1874 and 1851, which probably induced SSD 5 and SSD 6. In addition, the deformation around 1653 (i.e., SSD 8) was probably induced by the historical earthquake in 1660. However, no significant seismic activity was reported around 2004 (SSD 1) and 1806 (SSD 7). The sequence allows distinguishing temporally close earthquakes (ca. 20 years apart), which supports the interpretations of previous studies claiming that earthquakes induce soft-sediment deformations in the first centimeters below the water-sediment interface [e.g., *Marco et al.*, 1996; *Rodriguez-Pascua et al.*, 2000; *Monecke et al.*, 2004]. However, as is, visual inspection alone does not allow revealing the source earthquakes for the SSDs in the late 1950s and 1850-1875. Hence, more quantitative evaluation of the sediment deformations and ground motion intensities is required.

To avoid biases of visual inspection and achieve a full overview of soft-sediment deformations along the core, we quantified the layers' degree of curvature by calculating deformation index (DI) for each digitized layer in the radiograph (Figure 4 as an example, Figure S1 for the complete core). In this way, the levels with significant deformation within the lake sequence became more evident due to high deformation indexes, while the undeformed layers remain almost constant and close to zero DI values along the core. For instance, the micro-fault between 22 cm and 24.5 cm (F2 in SSD 2) is represented with a distinct DI peak, reaching a value of 120. The upper limit of the F2 corresponds to 1957 according to the varve chronology. Knowing that M_w 6.7 and M_w 6.5 earthquakes occurred in 1957 at ca. 50 km to the south of Köyceğiz Lake, we interpret that either of these earthquakes can be responsible for F2.

A temporal comparison of the complete DI plot and the PGA estimates of the instrumental and historical period earthquakes provide an extensive sedimentary earthquake record in the Köyceğiz sequence (Figure 5). During the instrumental period, the highest PGA was

estimated as 130 cm/s^2 due to the M_w 5.9 earthquake in 1959 that was ca. 6 km to the NW of the coring location. In addition, the earthquakes in 1941-43 and 1920-21 reached PGAs of 22 and 35 cm/s^2 , respectively. These earthquakes are associated with distinct SSDs on the DI plot as well (i.e., SSD 2, 3 and 4 in Figure 5a). On the other hand, the earthquakes in 2012, 2004, 1981, around 1970, and 1926 resulted in PGA values of around $10\text{-}15 \text{ cm/s}^2$, which did not cause any significant SSDs, except for the event in 2004 corresponding to SSD 1. Being within the highly loose sediments close to the water/sediment interface, the deformation corresponding to SSD 1 could be induced during coring, which would explain observing SSD without a significant earthquake in early 2000s. Considering also the uncertainties in the magnitude/epicentral location calculations and hence the PGA estimations, it seems that 20 cm/s^2 can be confidently accepted as the minimum PGA threshold to create SSDs.

In order to evaluate the changes in the source and composition of detrital sediment flux to the lake, we investigated the inter-element ratios of Cr, Ni, Ti, Fe and K, which can be broadly assumed as of detrital origin. The $[\text{Cr}, \text{Ni}]_{\text{av}} / [\text{Ti}, \text{K}, \text{Fe}]_{\text{av}}$ ratio profile, which was obtained by dividing the average of Cr and Ni profiles by the average of Ti, K and Fe profiles, shows amplified values just after the 1959 (M_w 5.9) earthquake (CR1 in Figure 5a). This observation implies Cr- and Ni-enriched sedimentation after the earthquake. *Avşar et al.* [2014] attributed Cr-Ni enrichments in a lake sequence on the North Anatolian Fault to amplified erosion rates due to earthquake-induced landslides and/or a seismically shattered landscape in the catchment dominated by mafic/ultramafic rocks. Similarly, given that peridotite is the dominant lithology in the Köyceğiz catchment (Figure 1b), we suggest that CR 1 is the result of amplified erosion rate in the catchment due to the strong seismic shock of the 1959 (M_w 5.9) earthquake.

Regarding the historical period, PGAs were calculated by considering the location and magnitude uncertainties (Figure 5c) provided by the European Archive of Historical

Earthquake Data (AHEAD) [Stucchi *et al.*, 2013]. Among them, the PGA values of the 1896 and 1756 earthquakes are relatively well estimated at around 18 cm/s^2 , and they do not correspond to any significant SSDs, supporting the minimum PGA threshold observed for the instrumental period earthquakes. There are also two clusters of earthquakes around 1870 and the early 1850s, which induced distinct SSDs and amplified erosion rates; i.e., SSD 5 – CR 2 and SSD 6 – CR 3, respectively. Around 1870, only the 1869 earthquake ($M_w 6.8 \pm 0.37$), which occurred 20-40 km to the northwest of the lake, could possibly reach PGA levels as high as those for the 1959 earthquake and result in CR 2. Regarding the early 1850s, the 1851 earthquake is relatively well located but it does not seem to reach PGA levels that would trigger CR 3. Although the locations of the 1855 and 1852 earthquakes are ambiguous, the varve chronology implies that the 1852 earthquake was closer to the lake and responsible for the SSD 6 and CR 3 couplet.

The 1741 earthquake, which generated a tsunami damaging the coasts of Rhodes [Ambraseys, 2009], did not affect the sedimentation in Köyceğiz Lake. Thus, its epicenter was probably farther than shown in Figure 5c. Similarly, the 1887 earthquake was unlikely to be as close as shown in Figure 5c. The 1660 earthquake, having the most uncertain location, could be responsible for SSD 8. If so, it must have occurred closer to the lake, possibly somewhere around the 1961 ($M_w 6.2$) earthquake in Figure 5b. Ambraseys [2009] reports a damaging earthquake in Rhodes in 1686, which might have induced the tiny DI peak around 1680. Around the dates of SSD 7, historical resources report no significant earthquakes. This event was probably a small local earthquake around Köyceğiz Lake.

5. Discussion

The temporal comparison between SSDs and earthquakes suggests that only the top few cm below the water-sediment interface is deformed during earthquakes. This implies that the

controlling mechanism for SSDs is seismically induced water oscillations rather than cyclic-loading-induced stresses in the deeper parts of the sequence. Additionally, while PGA values as low as 20 cm/s^2 can induce soft-sediment deformations, erosion rates can be amplified only by stronger and/or closer earthquakes. To evaluate this phenomenon, the magnitude versus epicentral distance data is plotted, and the earthquakes that can be stratigraphically distinguishable are classified based on their effects on sedimentation (Figure 6). We are confident that at least one of the three earthquakes during the period between 1941 and 1943 resulted in a distinct SSD. In addition, at least one of the two earthquakes in 1920 and 1921 also induced a distinct SSD. On the other hand, it is stratigraphically clear that the earthquakes in 1926, 1896, 1874 and 1756 did not induce distinct SSDs. Accordingly, the M_w vs. epicentral distance threshold to induce SSDs should represent the boundary separating these earthquakes that resulted in distinct and indistinct or no SSDs, which indeed corresponds to approximately a PGA value of 20 cm/s^2 (Figure 6). We are also confident that CR 1 was induced by one of the earthquakes during the series between 1957 and 1961. Among these earthquakes, the two in 1959, which generated PGA values around 130 cm/s^2 , were the ones most strongly felt in the area. Hence, at least one of the earthquakes in 1959 must be responsible for CR 1. In addition, the earthquake in 1869 also amplified erosion rates in the catchment resulting in CR 2. Accordingly, the M_w vs. epicentral distance threshold to induce CRs can be defined by these earthquakes, which corresponds to a PGA value of 70 cm/s^2 (Figure 6). Based on this observation, it can be claimed that the 1852 earthquake was stronger and/or closer to the Köyceğiz Lake than provided by the AHEAD (Figure 5c).

The PGA threshold for CRs determined in Köyceğiz Lake (i.e., 70 cm/s^2) is broadly supported by another example from Turkey. In Yeniçağa Lake, located on the North Anatolian Fault, *Avşar et al.* [2014] detected a significant CR induced by the 1944 Gerede (M_w 7.2) earthquake, which ruptured the segment where the lake is located on. According to

the attenuation relationship of *Akkar and Çağnan* [2010], the PGA value at the lake due to this earthquake can be estimated to be around 300 cm/s^2 . On the other hand, the 1999 Izmit (M_w 7.6) and Düzce (M_w 7.1) earthquakes, which had surface ruptures as close as ca. 80 km [*Barka et al.*, 2002] and ca. 50 km [*Akyüz et al.*, 2002] to the west of the Yeniçağa Lake, respectively, generated PGA values around $30\text{-}40 \text{ cm/s}^2$, and did not induce any CR in Yeniçağa Lake [*Avşar*, 2013].

Howarth et al. [2012, 2014 and 2015] investigated lake sequences located on the Alpine Fault (New Zealand), and determined a threshold for shaking intensity to generate CRs as MM IX (New Zealand Modified Mercalli Intensity). Intensities of $MM > IX$ corresponds to PGA values higher than 600 cm/s^2 (according to GeoNet: <https://www.geonet.org.nz>), which is considerably higher than the threshold determined in our study. The CRs detected by *Howarth et al.* [2012, 2014 and 2015] are significant increases in postseismic sediment flux generated by earthquake-induced macro-scale landsliding, so that they increased sedimentation rate by factor of three. On the other hand, the postseismic sediment flux increases in Köyceğiz Lake were at minor amounts probably due to micro-scale landsliding, which did not significantly affect the sedimentation rate, i.e. any distinct increases in varve thickness (Figure 5a). Köyceğiz varves are formed not only by runoff-related detrital material but also by carbonate precipitation from the water column. Although seismically mobilized sediments tend to increase varve thickness during the rainy season, relatively low carbonate precipitation may result in a varve thickness that is the same as the preceding/subsequent years. Moreover, while a drier year lacking extreme runoff events tend to form a thinner varve, the fine fraction of earthquake-mobilized material (Cr and Ni enriched in our case) can still be transported even by ordinary runoff events and can compensate for the loss in varve thickness. In summary, the CRs in Köyceğiz Lake are so minor that their contribution to the varve thickness (i.e. sedimentation rate) could have been easily masked by the climatic

contributions (i.e. carbonate precipitation), and hence they could be detected only by high-resolution geochemical scanning. The difference between the thresholds determined in New Zealand and this study implies that the detection of CRs and determination of a shaking intensity threshold to generate them may depend on the type and resolution of sedimentological methods applied and/or simply on the differences between the local geological/tectonic settings of the studied areas.

Köyceğiz varves reveal another important observation that the geochemical signals of CRs extend along 5 to 10 varves (Figure 5a). This implies that earthquake-mobilized sediments are not instantaneously transported and deposited by extreme runoff events during the rainy season after the earthquake; rather, this takes 5 to 10 years. This continuity also rules out the possibility that the origin of Cr-Ni enrichments is sediment focusing. The 5-to-10-year “residence time”, i.e. the time required to remove earthquake-mobilized sediments from catchments, is consistent with the suspended sediment measurements in rivers before and after the 1999 M_w 7.6 Chi-Chi and the 2008 M_w 7.9 Wenchuan earthquakes [e.g., *Hovius et al.*, 2011; *Wang et al.*, 2015]. Nevertheless, the residence time in the Köyceğiz catchment is much shorter than the one determined in New Zealand (e.g., 58 ± 7 years by *Howarth et al.*, 2014]. Short residence time in Köyceğiz catchment also implies that the CRs detected in this study have minor contribution to sedimentation compared to the ones detected in New Zealand.

In contrast to the limited number of above-mentioned studies that determined thresholds for catchment response, several comprehensive studies from Dead Sea [*Migowski et al.*, 2004; *Agnon et al.*, 2006], Swiss Alps [*Monecke et al.*, 2004], French Alps [*Wilhelm et al.*, 2016], Chile [*Moernaut et al.*, 2014; *Van Daele et al.*, 2015] and New Zealand [*Howarth et al.*, 2014 and 2015] suggested thresholds for SSDs and mass wasting deposits (MWDs) in lake sediments. Figure 7a compares the M_w vs. epicentral distance plots of the thresholds

determined by some of these studies with the ones of the current study. At the first glance, one can recognize that there is significant discrepancy between the SSD thresholds provided by this study, *Agnon et al.* [2006] and *Monecke et al.* [2004]. Similarly, the thresholds suggested for MWDs by *Wilhelm et al.* [2016] and *Monecke et al.* [2004] are quite different from each other. These discrepancies cannot be explained only by the errors related to the determination of magnitude and/or location of earthquakes. It is more likely that the discrepancies are due to different attenuation characteristics of the areas. Therefore, rather than using M_w vs. epicentral distance plots, we estimated PGA values corresponding to the thresholds of different studies by applying local attenuation relationships, in order to make them comparable.

Agnon et al. [2006] investigated the varve-like laminated sediments of the Dead Sea basin and obtained a threshold by comparing the observed SSDs (mainly intraclast breccias) with numerous instrumental and historical period earthquakes. Accordingly, all historical earthquakes with calculated local intensities higher than “V” induced SSDs. The M_w vs. epicentral distance plot of the local intensity “V” is presented in Figure 7a. If the M_w - epicentral distance pairs corresponding to their threshold are used to estimate PGA values by utilizing the attenuation relationship of *Ambraseys et al.* [2005], which is developed for Dead Sea region, we obtain PGA values of 17-19 cm/s^2 (Figure 7b). This result is consistent with the SSD threshold determined in this study (i.e., 20 cm/s^2).

Monecke et al. [2004] investigated the effects of four earthquakes in Switzerland (in 1356, 1601, 1774 and 1964) on the sedimentation of six lake basins located in the Swiss Alps, which also allowed determination of thresholds for SSDs and MWDs. However, after their publication, new magnitudes for the earthquakes they used were suggested by the revised earthquake catalogue of Switzerland [ECOS-09; *Fäh et al.* 2011]. Hence, we revised their data based on the new magnitudes (Figure S3a) and defined M_w - epicentral distance pairs,

which are presented in Figure 7a. For this region, *Cauzzi et al.* [2015] have recently suggested an attenuation relationship, which we used to estimate threshold PGA values for SSDs and MWDs (the details of the PGA estimations can be found in the supplementary file). Accordingly, it seems that 28 cm/s^2 and 55 cm/s^2 can be accepted as the PGA thresholds for SSDs and MWDs, respectively (Figure 7b). The SSD threshold obtained from *Monecke et al.* [2004] is slightly higher than the threshold suggested by the current study. It should be noted that, the SSDs detected by our study, *Agnon et al.* [2006], and *Monecke et al.* [2004] are of different types (i.e. micro-faults, micro-folds, intraclast breccias, and liquefaction structures). Given this fact, and the uncertainties included in attenuation relationships and determination of magnitude/location/depth of earthquakes, the PGA thresholds for SSDs suggested by this study and the other two, which ranges between 17 cm/s^2 and 28 cm/s^2 , seems to be in good agreement (Figure 7b).

In addition to *Monecke et al.* [2004], *Wilhelm et al.* [2016] also suggested thresholds for MWDs by investigating the effects of numerous instrumental and historical earthquakes on the sedimentation of eight lakes located in French Alps. They showed that the thresholds for MWDs can vary depending on sedimentation rate (SR) such that lakes having higher sedimentation rates are more sensitive to earthquake shaking. Among the lakes in their study, based on the occurrences of MWDs, Lake Blanc Aiguilles Rouges (BAR, $\text{SR}=0.9 \text{ mm/yr}$) and Lake Laffrey (LAF, $\text{SR}=2.2 \text{ mm/yr}$) are found to be the least and the most sensitive lakes to earthquake shaking, respectively. Therefore, we used the thresholds provided for these two lakes to determine a PGA range for MWDs and further compare it with our results. Since they used epicentral intensity while determining the thresholds, we first converted epicentral intensities to moment magnitudes by using the *Bakun and Scotti* [2006]'s equation generated for French Alps, and assuming an average depth of earthquakes as 15 km. The corresponding M_w vs. epicentral distance plots for the two lakes are presented in Figure 7a. *Drouet and*

Cotton [2015] have recently provided a local attenuation relationship for French Alps. When the M_w - epicentral distance pairs presented in Figure 7a are used in this attenuation relationship, the PGA values ranges between 40 and 120 cm/s^2 for BAR, and 15-35 cm/s^2 for LAF. Yet, according to common sense, sensitivity of a lake to earthquake shaking is expected to correspond to a constant PGA value, rather than having a wide range. This implies that thresholds determined by epicentral intensity vs. distance data may not follow a perfect logarithmic relation as suggested by *Wilhelm et al.* [2016]. Thus, in attempt to obtain more reasonable PGA threshold values, we followed an inverse procedure such that we calculated M_w - epicentral distance pairs for some constant PGA values, i.e. for 20, 30, 40, 60 cm/s^2 etc. After converting the obtained M_w values to epicentral intensities based on the *Bakun and Scotti* [2006]'s equation, and plotting new epicentral intensity - distance pairs on the original plots provided by *Wilhelm et al.* [2016], we found out that the epicentral intensity - distance pairs obtained for PGA values of 22 cm/s^2 and 60 cm/s^2 well approximates their original thresholds suggested for LAF and BAR, respectively (Figure 7b & Figure S3b). However, these values should not be accepted as strictly absolute values since our epicentral intensity to M_w conversion based on *Bakun and Scotti* [2006]'s equation may reveal slightly different results depending on the depth of earthquakes.

Moernaut et al. [2014] and *Van Daele et al.* [2015] also proposed intensity thresholds for MWDs (i.e., LT1s by their terminology) from Chilean lakes as $MSK\ VI^{1/2}$ and $V^{3/4}$ (Medvedev–Sponheuer–Karnik Intensity Scale), respectively. These intensity levels approximately correspond to 110 and 180 cm/s^2 PGA values according to the USGS ShakeMaps, the intensity vs. PGA scaling of which is based on *Wald et al.* [1999] (<http://earthquake.usgs.gov>). Moreover, *Howarth et al.* [2015] proposed intensity thresholds for MWDs from New Zealand as MM VII (New Zealand Modified Mercalli Intensity), which

corresponds to PGA values around 125 cm/s^2 (according to GeoNet:

<https://www.geonet.org.nz>).

Although it seems that MWD thresholds proposed by the studies from Alps, Chile and New Zealand are somewhat in agreement in terms of intensity, they seem quite different in terms of PGA; ca. $22\text{-}60 \text{ cm/s}^2$ for Alps, $110\text{-}180 \text{ cm/s}^2$ for Chile and New Zealand. Given this difference, it could be the scope of future studies to discuss which parameter is more suitable to improve quantitative lacustrine paleoseismology. Here, our concern is that Köyceğiz Lake experienced PGA values as high as 130 cm/s^2 during 1959 earthquakes, which suggests that we should have observed MWDs according to the above mentioned studies. Although we do not see any MWDs in our core, we cannot precisely say that earthquakes did not trigger MWDs in Köyceğiz Lake as our study is based on a single core from the most central part of the basin. This situation exemplifies the importance of multi-core investigations in order to reveal complete paleoseismic records based on MWDs in lake basins.

6. Concluding Remarks

The varved sequence of Köyceğiz Lake revealed that earthquake-induced soft-sediment deformations (SSDs) can be limited to the top few cm of the sequence, which allows proper dating and disentangling of even temporally close earthquakes (around 20 years apart).

However, detection of micro-scale SSDs requires the application of proper sedimentological methods such as u-channel radiography. Further efforts should be expended to develop methods to detect micro-scale SSDs within non-laminated sequences.

Earthquake-triggered landslides in the Köyceğiz catchment remobilized soils derived from ultramafic rocks (peridotite in this case), which resulted in Cr- and Ni- enriched sedimentation lasting for 5 to 10 years. Seismically amplified sediment fluxes (CRs) may not

be evident along varve thickness profiles. Therefore, geochemical analyses may sometimes be needed along with the physical sediment properties to detect CRs. We also found that while SSDs can be induced by PGA values as low as 20 cm/s^2 , CRs can be triggered by at least 70 cm/s^2 . The PGA threshold determined for SSDs in this study is in good agreement with the ones determined from Dead Sea sediments [Agnon *et al.*, 2006] and Swiss lakes [Monecke *et al.*, 2004]. However, the PGA threshold determined for CRs in our study (70 cm/s^2) is significantly lower than the one found in New Zealand (MM > IX, which corresponds to PGA values higher than 600 cm/s^2) [Howarth *et al.*, 2014 and 2015]. This discrepancy implies that local geological/tectonic/topographic properties of catchments may play an important role on the minimum required acceleration to amplify erosion rates, i.e. CRs.

Although several strong earthquakes affected Köyceğiz Lake during the last 400 years, no subaqueous mass-wasting deposits (MWD) were detected in the core. There may be two explanations for this; i) our study is based on a single core taken from the central part of the lake, and earthquake triggered turbidites might not have reached to the coring location. A multi-core study would probably reveal MWDs if there were any, ii) the southern basin, from where the core was taken, does not have significant tributaries transporting enough sediment to form subaqueous slopes that are susceptible to failure during each earthquake. It should therefore be kept in mind that, MWDs chronologies constructed without basinwide multi-core investigations may not capture the complete seismic history of the region.

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Captions:

Figure 1. (a) Tectonic setting in the Eastern Mediterranean (simplified from *Hall et al.*

[2014]). The Hellenic and Cyprus Arcs are sinistrally offset by a subduction transform edge propagator (STEP) fault composed of the Pliny-Strabo (PSFZ) and the Fethiye-Burdur

(FBFZ) fault zones. (b) Bathymetry, relief-shaded topography and lithology in/around Köyceğiz Lake. White triangles and the associated numbers stand for the elevations of some hills in the catchment. The core was taken from the southern basin (black dot). (c)

Instrumentally-recorded earthquakes ($M_w \geq 4.5$) around the lake during 1900-2013 [KOERI-NEMC EQ Catalog - <http://www.koeri.boun.edu.tr>] are plotted together with the traces of main faults within the PSFZ and FBFZ, which are sinistrally offset by the Gökova-Yeşilüzümlü Fault Zone (GYFZ) (from *Hall et al.* [2014]). (d) Estimates of magnitudes and

epicentral locations of historical earthquakes during the last 400 years [AHEAD:

<http://www.emidius.eu/AHEAD/>; *Stucchi et al.*, 2013]. The grey ellipses and circles stand for the spatial uncertainties of the proposed epicentral locations, and the numbers in parentheses are the proposed moment magnitudes.

Figure 2. (a) Optical image of the top most part of the core overlapped by the Ca/Ti plot.

Note that the yellowish-white layers deposited in summer seasons are characterized by obvious peaks along the Ca/Ti profile. (b) The ^{137}Cs and $^{210}\text{Pb}_{\text{xs}}$ activity profiles along the top 50 cm of the core. (c) Sediment chronologies obtained from ^{137}Cs - $^{210}\text{Pb}_{\text{xs}}$ dating and varve counting, as well as the error in the varve chronology with respect to depth (blue line).

Figure 3. Optical/radiographic images of the core and the soft-sediment deformation zones (i.e., SSD 1-8) in seven close-up views. Interpretative sketches are also provided next to each close-up view. The SSD zones are shown in red within the regular background sedimentation (gray). Note that the fault plane of F3 is shifted due to the twisting of the core sections at 84.2 cm. Based on the varve chronology, the dates of the upper horizons of the SSD zones temporally coincide with the significant earthquakes in the region presented in Figure 1c and 1d.

Figure 4. Illustration for Deformation Index (DI) calculation. The Mean Square Error (MSE) value of each detrended line is in fact a measure of the degree of deformation of the corresponding layer.

Figure 5. (a) Temporal comparison of the Deformation Index (DI), variations in sediment flux from the catchment ($[\text{Cr}, \text{Ni}]_{\text{av}} / [\text{Ti}, \text{K}, \text{Fe}]_{\text{av}}$), varve thickness, and predicted peak ground accelerations (PGA) at Köyceğiz Lake. (b) Epicentral locations and moment magnitudes (in brackets) of significant instrumental period earthquakes. (c) Approximate epicentral locations and magnitudes of historical earthquakes based on the European Archive of Historical Earthquake Data (AHEAD) [<http://www.emidius.eu/AHEAD/>; *Stucchi et al.*, 2013]. PGA predictions for these earthquakes are plotted in (a) (red dots with horizontal lines as error ranges). Note that the error in varve chronology around CR 1 and CR 2 - CR 3 is ± 0.5 and ± 2 years, respectively.

Figure 6. The M_w vs. epicentral distance plot of the earthquakes around Köyceğiz Lake, which are classified according to their effects on the sedimentation; no or indistinct effects (gray dots), only SSDs (red dots), and both SSDs and CRs (blue dots). The lower threshold PGA values to trigger SSDs and CRs are determined as 20 and 70 cm/s^2 , respectively.