Factors leading to slope failure on a sediment-starved margin: the southwestern continental margin of the East Sea, Korea

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Highlights

Four giant submarine landslides identified on southwestern margin of East Sea.

Radiocarbon dates suggest the landslides occurred at ca. 19-11 ka.

Coarse-grained sandy layers likely to have acted as a weak layer.

Earthquakes associated with tectonic activity were most likely the final trigger.

Sediment cores indicate no recent mass-wasting has occurred since ca. 11 ka.

Abstract

Submarine landslides are common geomorphological features of continental margins. Some of the largest submarine landslides occurred on low-angle ($< 4^\circ$), sediment-starved margins, yet their preconditioning and trigger mechanisms are still largely unconstrained. The southwestern continental margin of the East Sea (between 37.5°N and 38.0°N), Korea, occupies a narrow shelf (< 10 km), is characterized by low sedimentation rates ($\sim 3\text{--}7$ cm/ka) with an average gradient of less than 2° . Here, we investigate submarine landslides using newly collected datasets including multibeam echosounder (MBES), chirp sub-bottom profiler, multichannel seismic (MCS) data and ten piston cores. MBES data from the margin reveal at least four major submarine landslides initiated at depths of 400 m to over 600 m. These landslides left clear headwall scarps on the seafloor with reliefs reaching over ~ 130 m and extend for over 40 km. MCS data show that some of the failures have resulted in the complete disintegration of the failed mass, while others have resulted in the deposition of well-defined hummocky debris flows. Sediments recovered downslope of the headwall scarps contain slides and debris flow deposits and turbidites that are overlain by bioturbated hemipelagic layers. Radiocarbon dating from hemipelagic units overlying MTDs within the headwall scarps reveal that major failures occurred at ca. 11 to 19 ka, coinciding with the time of the Last Glacial Maximum (LGM) to early deglaciation. Since then, hemipelagic sedimentation has prevailed throughout the sediment starved slope. Slope stability analyses based on geotechnical properties of sediments indicate that all areas are stable under static, and even stable under loads derived from earthquakes in instrumental records, but there were probably earthquakes in pre-historical records (i.e., with a longer recurrence interval) of potentially significant larger magnitude. We suggest that the preferential occurrence of major failures adjacent to the major faults on the lower slope may ultimately be tectonic-controlled although other factors may have contributed as well. Our work shows that coarse-grained clastic sediments are abundant in the shallow subsurface and that these higher-permeability units, often identified as weak layers, would focus fluid flow and could act as slip planes for slope failure. Our data also indicate that tectonic steepening and gas charging are other key parameters for controlling instability in sediment-starved margins.

49 **Keywords:** submarine landslides, East Sea of Korea, sediment-starved margins, slope stability,
50 headwall scarps, earthquakes

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1. Introduction

Submarine landslides play a major role in the construction of continental margins and involve a series of processes that can result in significant hazards for offshore infrastructure and coastal facilities (Hampton et al., 1996; Locat and Lee, 2002; Lee 2009; Urgeles and Camerlenghi, 2013). Amongst those hazards is the generation of tsunamis (Mosher et al., 2010; Harbitz et al., 2014). The nature and triggering mechanisms of submarine landslides have been discussed intensely during the last few decades, fundamental evidence having come from the repeated seismic and bathymetric mapping of the seafloor. Notable case histories include the particularly well-studied the 1998 Papua New Guinea tsunami event (Tappin et al., 2001), the 1929 Grand Banks event (Bent 1995), and the 1908 Sicily event (Billi et al., 2008). Based on these and previous studies, scientists have pondered possible factors -or a combination of factors- that are often involved in submarine failures. These include steep slope gradients, elevated fluid pressures caused by rapid sedimentation and variations in fluid flow, dissociation of methane hydrates, earthquake shaking, as well as changes in sea level (Hampton et al., 1996; Masson et al., 2006; Chaytor et al., 2009; Lee, 2009; Urgeles and Camerlenghi, 2013).

Remarkably, many large-scale submarine landslides occur on sediment starved margins (< 15 cm/ka) with low gradients ($< 4^\circ$) (Urlaub et al., 2012, 2015). Examples include the northwest African continental margin (Wynn et al., 2000; Antobreh and Krastel, 2007; Krastel et al., 2018), and the eastern Australian margin (Clarke et al., 2016). A number of hypotheses have been put forward to explain the causes of landslides on these nearly flat slopes. Clarke et al. 2016, based on seismic reflection data and sediment cores, favor the earthquake triggering mechanism for the initiation of landslides on the eastern Australian margin. Erosive scouring and over-steepening are inferred to have preconditioned the sediments for failure. Antobreh and Krastel (2007), by analyzing high-resolution seismic reflection profiles, suggest that overpressure along weak layers may have facilitated slope failures on the northwest African margin. These landslides were hypothesized to have been triggered by earthquakes. Urlaub et al. (2012), based on 2D numerical modeling, conclude that overpressure is

not enough for initiating slope failure in locations with low sedimentation rates and, thus, other factors (e.g., weak layers) must be present. Later simulations by Urlaub et al. (2015) show that it is difficult to generate significant excess pore pressure if sediment accumulation is slow on slopes with low gradients. Hence they suggest that other factors (e.g., presence of a layer with high compressibility) need to be considered. Although these numerical models and other studies from geophysical data consider how slope failures may occur, the mechanisms that lead to failure remain a fundamental geological question. Commonly, a combination of internal and external forcing mechanisms as briefly discussed for some case stories above, is involved, especially when slope failures occur on low gradients. In this context, the SW margin of the East Sea provides an excellent location to determine physical processes leading to slope failures in an area where the influence of riverine sediment supply and slope gradients are low.

Here we integrate newly collected data, including multichannel seismic (MCS), chirp sub-bottom profiler, multibeam echosounder (MBES), and piston cores to address the following specific questions. First, what are the characteristics of submarine landslides (i.e., distribution, style, morphology) observed along the SW margin of the East Sea, Korea? Second, what are the age and sedimentological characteristics of the landslides?. Third, we use the Factor of Safety (FOS) analysis to evaluate the present-day stability of the continental slope. Fourth, we aim to investigate what factors (i.e., preconditioning and trigger mechanisms) caused slope failures on this low-angle sediment starved margin. Finally, we develop a depositional model for the preconditioning and triggering of submarine landslides based on insights from this study.

2. Physiographic setting

The Ulleung Basin lies at water depths of 2,000 to 2,500 m and is generally characterized by smooth seafloor except for a few seamounts, such as the Ulleung Seamount (US) and the Ulleung Island (UI) (Fig. 1). The South Korea Plateau (SKP) forms one of the largest submarine continental remnants in the East Sea, occupying the northern part of the Ulleung Basin (Fig. 1). The SKP is suggested to be composed of rifted continental crust (Tamaki, 1988). The eastern part of the SKP

(ESKP) covers an area of about 11,000 km² with a roughly circular shape and lies in water depths of 700 to 2000 meters. It extends about 350 km in the E-W and 260 km in the N-S directions. The WSKP, which is separated from the ESKP by the North Ulleung Trough (NUT, Fig. 1), is characterized by numerous ridges that form an undulating topography of 300 m to 400 m relief. These ridges typically coincide with faulted basement blocks (Cukur et al., 2015).

The western margin of the Ulleung Basin is relatively narrow, with a shelf of about less than 30 km wide and steep (as high as 10°), thinly sedimented (less than 500-m-thick) upper and lower slope (Horozal et al., 2016). Average sedimentation rates on the slope, determined from piston cores, range from 3.2 cm/kyr to 9.2 cm/kyr (Cukur et al., 2016). The low sedimentation rates on the slope are ascribed to the absence of major river systems delivering sediment to the margin (Cukur et al., 2016). The N-S trending Hupo Bank forms a major topographic high on the shelf that rises 80 m above the adjacent seafloor. To the east of the bank, the slope is 3° on average but can be as steep as 10° in places. To the north of 37.10 N, which is the focus of our study, the continental margin trends NNW-SSE (Fig. 1). The margin presents a thinly sedimented upper and middle slope with continental bedrock exposed on its flat, narrow shelf (Cukur et al., 2018). Here the continental shelf is only ~20 km wide with a shelf break at 150 m. The slope is less than 4° and dissected by two major slope-parallel submarine canyons, namely the Gangneung and Donghae (Cukur et al., 2018). The continental slope is bounded on the east by the Onnuri Basin at water depths of 1300-1500 m (Fig. 1).

3. Datasets and methods

The data used in this study include:

(1) Approximately 1800-km-long sub-bottom chirp profiles collected by the Korea Institute of Geosciences and Mineral Resources (KIGAM) in 2016, 2017, and 2018 (Fig. 2). The sub-bottom chirp data were acquired in water depths of less than 50 m to over 1100 m using an SYQWEST-BATHY 2010 sub-bottom profiler system. The chirp system provides a signal penetration of up to 60 m with a vertical resolution of ~8 cm.

(2) Four multichannel seismic reflection profiles (~125-km long) acquired by KIGAM in 2002, 2016, and 2019. The multichannel seismic reflection surveys were carried out onboard the *R/V Tamahe II*. The data acquisition in 2002 was carried out using a six-airgun (2000 psi) array and 80-channel (1000-m-long) streamer with a record length of 6 s. Data processing was carried out using Landmark ProMAX processing software and included geometry setup, CMP sorting, filtering, and normal moveout (NMO) correction. For those acquired in 2016 and 2019, a 48-channel (600-m-long) digital streamer recorded shots with a shot interval of 25 m from a 20.5-1/2,000 psi source array. Systematic data processing included: geometry setup and CMP sorting, gain correction, F-K filtering, deconvolution, normal moveout correction, semblance velocity analysis, and full-offset stack.

(3) Approximately 3000 km² of multibeam echosounder data (MBES) collected by KIGAM in 2016, 2017, and 2018. MBES data were collected using a Kongsberg EM 302 multibeam echosounder with a nominal frequency of 30 kHz and a swath width of 130 m. This data were processed with CARIS HIPS and SIPS hydrographic cleaning system software with a grid resolution of 100 m. The IHS Kingdom Suite and Global Mapper were used for data analysis and interpretation.

(4) Ten piston cores collected by KIGAM in 2017 and 2018 (Fig. 2) from in and outside of submarine landslides to characterize sedimentology and physical properties and to constrain geological ages of surficial sediments (Fig. 2). Six cores (P11, P02, P03, P13, P14, and P16) were collected from within four submarine landslide scars. Four reference cores (P10, P04, P12, and P15) were recovered from adjacent slopes that do not present obvious slide features or morphologies (Fig. 2). All cores were split, photographed, and described macroscopically. X-radiographs were made on 1 cm thick slabs (25 cm long) taken from one-half of the split cores. Digital images of core surfaces were obtained using a line-scan camera, and lightness (L*) profiles were deduced from the other half of the image of the split core. Moisture and density (MAD) measurements, including water content, wet bulk density, and porosity were determined from wet mass, dry mass, and dry volume measurements of ~10 cm³ samples taken immediately after core-splitting following the International Ocean Discovery Program (IODP) onboard laboratory practices (Blum, 1997). Following MAD

measurements, the remaining subsamples were analyzed for grain-size using a laser-scattering particle analyzer (Microtrac S3500) with a measurement spacing of 20 μ m. Undrained shear strength measurements were performed on working half-sections using a miniature vane shear apparatus with a sampling resolution of 20 μ m.

(5) A total number of 33 radiocarbon dates determined by accelerator mass spectrometry (AMS) on samples taken from ten cores (Tables 1 and 2). Dates were obtained from picked planktonic and benthic foraminiferas along with wood and shell fragments extracted from the hemipelagic sediments and were determined at the Arizona AMS Laboratory. Conventional ^{14}C years BP were converted into calibrated calendar ages following Stuiver and Reimer (1993) (Tables 1 and 2). Radiocarbon ages measured from marine carbonates can provide basic constraints of the core chronology, but this sampling method is not ideal for precise age information. Bulk samples can cause uncertainty in interpretation because they can be derived from different sources of different ages (Harris et al., 1996). In particular, bulk sediment samples usually contain old reworked sediments, which result in the calculated sediment age is older than the actual age. According to Urlaub et al. (2013), uncertainties may exist in AMS ^{14}C dating for ± 100 years. Additionally, bioturbation can be another potential error source.

The stability of slope sediments was determined from the calculation of the 2-D factor of safety (FOS), assuming either static loads ($\text{FOS}_{\text{static}}$) or seismic loading (FOS_{eq}) as the driving forces for mobilization on an infinite slope. A slope is considered stable when the FOS exceeds 1, and unstable when it is less than 1 (Morgenstern and Price 1965). In the calculation of $\text{FOS}_{\text{static}}$, the undrained shear strength (S_u) of the sediment column at a given depth z (representing the resisting force) is balanced against the driving forces along a slope-parallel failure plane:

where θ is the slope angle above landslide scarp, and γ' is the buoyant unit weight (density difference between solid and water times g) (Morgenstern 1967). The use of undrained strengths for calculation

of the FOS_{static} implies that FOS values represent minimum conservative values. For the calculation of FOS against seismic loading FOS_{eq} under undrained conditions, the pseudo-static formula of ten Brink et al. (2009) was used:

$$\frac{\gamma \sin \theta}{k_e \gamma' \cos \theta + c + \sigma' \tan \phi}$$

where θ is the slope angle, k_e is the cyclic load of an earthquake represented by a constant horizontal acceleration (expressed as a fraction of gravity) required to cause failure, γ is the unit weight of the bulk sample, and γ' is buoyant unit weight of the material that is likely to fail.

In order to account for the effect of pore pressure on FOS calculations, drained strength parameters such as cohesion and friction angle are needed. Due to the lack of these measurements, we assume that measured undrained shear strength represents the under hydrostatic conditions, and pore pressure would reduce the ratio of shear strength to vertical stress. For normally consolidated sediments, this ratio is about 0.3 ($C_u/\sigma'_v=0.3$); hence we derived strength from the vertical stress. Bulk density measurements were measured only up to ca. 4.8 m below the seafloor for core P10, which is the limit of the piston core. Hence, we adopted density values for depths between 4.8 m to 23.6 m below seafloor from an adjacent well drilled on the slope (Ryang et al., 2014). In the interval from 23.6 to 47 m, densities were defined by using the following equation:

$$(3)$$

4. Results

4.1 Morphology and characteristics of submarine landslides

MBES data collected along the southwestern margin of the East Sea span depths from 120 m down to 1200 m (Fig. 3a). The bathymetric data reveal four major submarine landslides (labeled as SL1 to SL4) that are represented by broad depressions bounded by scarps as much as 130 meters high (Figs. 3a-b).

The southernmost submarine landslide, **SL1**, has an area of 160 km² and a volume of 3.6 km³ estimated by landslide longitudinal and transversal sections from the MBES data assuming a slab-like geometry (Urgeles et al., 2019) (Fig. 3). Its headwall scarp is arcuate-shaped in map view with a slope angle of ~14° and depth of ~35 m below the surrounding seafloor (Fig. 4a). It is located at ~500 m water depth, and sediment evacuation took place to the northeast (Fig. 3). The headwall scarp is ~7 km wide on the southwest (increasing downslope >20 km). Sub-bottom chirp profiles crossing the SL1 show chaotic- to transparent-acoustic facies with hyperbolic surface expression beneath the headwall scarp (Fig. 4b-c). These facies are up to 20 m thick. Chirp profiles further show truncation of flat-lying seismic reflections at the headwall scarps and the presence of stepped headwall scarps. The base of the chaotic seismic facies is characterized by continuous, high amplitude reflections that are found beneath ~10-15 m of subseafloor. Chirp profiles also delineate a sedimentary block (~800 m-wide) at the edge of the southeastern sidewall in about 530 m water depth (Fig. 4c). Here, the sidewall is about 35 m high and has a slope angle of 13°, while the average slope angle of the adjacent slope is about 2.5°. Chaotic- to transparent-seismic reflections with hummocky surface expressions are visible directly beneath the sidewall.

The submarine landslide 2 (**SL2**) covers an area of about 280 km² and the estimated volume of displaced sediments is ~8.4 km³ (Fig. 3). Its headwall scarp is arcuate-shaped, occurs at water depths of 500-600 m and trends northeast (Fig. 3). The height and slope of the headwall scarp range from ~50 to ~130 m and 10° to 25°, respectively (Fig. 5a). The scarp is about ~12 km wide (Figs. 3). Chirp data along SL2 show chaotic- to transparent-seismic facies downslope of the scarp (Fig. 5b). The chirp profile also shows that the area upslope of the headwall scarp is characterized by continuous seismic reflections. One of the chirp profiles, traversing the southwestern corner of the SL2 scarp, also reveals some intact slide blocks (~30 m high and ~200 m wide) with a well-stratified internal character (Fig. 5c). They occur immediately below the headwall scarp, forming closely spaced NW-SE trending slide blocks (Fig. 5d). The Onnuri Basin, located at the lower reaches of the continental slope, is characterized by acoustic facies consisting of stacked, wedge-shaped, chaotic- to transparent-units (with a maximum thickness of 110 m) separated by high-amplitude continuous reflectors (Fig. 6).

These units laterally pinch-out onto the basin flanks.

The submarine landslide 3 (**SL3**) has a surface area of 120 km² and a volume of 5.4 km³ (Fig. 3). The headwall scarp is up to ~60 m high with a slope angle of about 15° (Fig. 7a), occurs at a water depth of ~600 m, is U-shaped, and sediment evacuation happened from SW to NE (Fig. 3). The headscarp is ~9 km wide. The seismic signatures of the SL3 are shown in Fig. 7b-c. Undisturbed parallel seismic reflections characterize the unfailed area to the SW of the headwall scarp. The sediments downslope of the scarp are characterized by chaotic seismic reflections. Also evident is the two stepped headwalls that are up to 30 m high each. The chaotic seismic units extend down to 8 km from the source region. Sedimentary units derived from mass-wasting and those from insitu deformation resulting from the landslide are up to 100 m thick (Fig. 7c). On the most distal areas, compressional features (i.e., folds and thrusts) are developed, giving to the slid sediment a hummocky topography (Fig. 7c). Beyond the toe of the slide, the slope sediments are characterized by high-amplitude continuous seismic reflections.

The northernmost submarine landslide, **SL4**, has affected an area of 90 km² and removed a volume over 2 km³ (Fig. 3). The headscarp has an arcuate-shape in map view and is located in approximately 400 m of water near the shelf edge (Fig. 3). It has walls ranging in height from 50 to 65 m with slopes reaching up to 20° (Fig. 8a). It is ~3 km wide on the headwall region (increasing northeastward) (Fig. 3). The slide scar can be traced ~25 km downslope to 1000 m depth. A chirp profile shows a 55-m-high pronounced morphological step at the seafloor (Fig. 8b). Well-stratified slope sediments upslope from the headwall scarp abruptly terminate at its walls. Laterally continuous high-amplitude reflections, parallel to the adjacent slope stratigraphy, characterize the downslope of the headwall scarp. In contrast to the southern submarine landslide scarps, well-stratified seismic reflections without surface hyperbolae appear downslope from the SL4 scarp (Fig. 8b).

4.2 Sedimentary facies and sediment physical properties

In the ten piston cores from within and adjacent to the slope failures, we distinguish seven

sedimentary facies based on sedimentary structures (Fig. 9a) and sediment physical properties (Figs. 10-11). The surficial sediments recovered from all cores are dominantly composed of bioturbated mud or silt (**BM** facies; Fig. 9). These facies include mottling and pyrite filaments of variously shaped burrows that are mostly horizontal to inclined. Facies BM consists primarily of silt (up to 80%) and clay (20% to 60%) with an average grain size of 7.5 phi (Figs. 10-11). Overall, the physical properties of BM facies show changes with the sediment grain size (Figs. 10-11). Density increases downcore, while porosity and water content decrease. Undrained shear strengths in facies BM is variable but with an overall increase downcore, in line with the increasing densities with depth. A significant increase in undrained shear strength (from ~9 kPa to ~60 kPa) is observed between 385 and 405 cm bsf in core P03 (Fig. 10b). A distinct color change is also evident in the same interval (Fig. 10c).

The faintly laminated sandy mud or mud (**FLM**) comprises continuous or discontinuous horizontal laminae with abundant foraminifera (Fig. 9). The thickness of this facies ranges from less than 5 cm to over 15 cm and alternates with the BM. Bahk et al. (2000) have found the FLM and BM facies in the Ulleung Basin and interpreted them as hemipelagic sediments deposited under well- to poorly oxygenated bottom-water conditions.

Facies **MCM** (mud-clasts mud) comprises mud clasts of varying shape, size, and color (Fig. 9). The clast supported character of facies MCM suggests that these facies represents mass-transport deposits (i.e., debris flow deposits) composed of clasts of multiple lithologies. This facies association, which was penetrated by piston core P02, is mostly characterized by high silt contents (70%) with a mean grain size of ~7 phi (Fig. 10b). Clay content is up to 30%. Facies MCM is characterized by higher densities and undrained shear strengths and low porosity readings compared to the BM facies above (Fig. 10b).

Facies **DM** represents deformed mud layers that appear to be folded or inclined (Fig. 9). It is only found in core P14 (Fig. 11a) and is capped by wavy laminated silty sand/mud (**WLzS/m facies**, see below). The upper boundary of DM forms a sharp contact with the overlying WLzS/m facies (Fig. 9). This facies is interpreted as mass-transport deposits (i.e., slide deposits) associated with submarine

landslide 3 (SL3).

The wavy laminated silty sand/mud facies (**WLzS/m**), interpreted as turbidite deposits, is seen in cores P14, P10, P11, P04, P16, and P14 (Figs. 9, 10 and 11). This facies consists mostly of sand (~50%) with silt contents exceeding 30 %. This facies is rich in clastic minerals (i.e., quartz and feldspar; over 90%) with very rare benthic or planktonic foraminiferas (Fig. 9b). It has higher bulk density and lower porosities than the overlying and underlying facies units (Figs. 10 and 11).

The massive gravelly muddy sand (**MgmS**) facies is found only in core P15 and consists of very poorly-sorted, medium sand (mean grain size: 1.5 phi) with abundant subangular clastic fragments, reaching up to 6-mm-long (Figs. 9a and c). There is also evidence of rounded shell fragments. Facies MgmS contains up to 70% sand with gravel contents up to 7% (Fig. 11b). The physical properties of MgmS are marked by an abrupt increase in density from $\sim 1.7 \text{ g/cm}^3$ to $\sim 2.1 \text{ g/cm}^3$. In contrast, porosity decreases dramatically across this facies interval from $\sim 65\%$ to $\sim 40\%$. Facies MgmS is interpreted as grain flow deposits derived from continental shelf sands that were driven over the continental shelf edge by mass-wasting events (Cukur et al., 2018).

The convoluted-laminated sandy mud (**CLsM**) facies, penetrated by core P12, comprises sand (40%), silt (37%), and clay (23%) with a mean grain size of $\sim 5.9 \text{ phi}$ (Fig. 11a). It is a thin layer (ca. 4-cm-thick) of convoluted, laminated, sandy mud with sharp lower and upper boundaries (Fig. 9). The coarse and fine fraction of CLsM consists chiefly of clastic rock fragments (up to 2 mm thick; Fig. 9d). Although rare, the fine fraction also includes bioclastic materials, including benthic and planktonic foraminifera. Based on sedimentary structures, grain size, and sharp basal and upper contacts, Facies CLsM is interpreted as turbidite deposits derived from the shelf edge.

4.3 Sedimentation rates and age of submarine landslides

Sedimentation rates (Fig. 12a), calculated from ^{14}C age dating, reveal that sediment cores from landslide scars have higher sedimentation rates (ranging from 12 cm/ka to 37 cm/ka) than those upslope from the landslide scars (P12: 3 cm/ka, P10 and P04: 7 cm/ka; Fig. 12a). Also, core P14

shows much higher sedimentation rates during the Holocene. Looking at coring locations, it is also evident that the sedimentation rate increases from the southeast (P11, P02, P14) to northwest (P16). This fact is explained by the location of core P16, which is closer to the present coastline.

Bulk radiocarbon ages were also used to establish minimum and maximum age constraints on the timing of submarine landslides (Fig. 12b). A bulk radiocarbon (^{14}C) age determined for sediment sampled directly above the boundary surface for core P11, places a maximum age of 18.25 ka for SL1 (Fig. 12c). This boundary surface exhibits a distinct gap in the ^{14}C age of at least 20 ka (Fig. 12c). Radiocarbon dates taken from above and below the boundary surface in core P02 of SL2 also present a noticeable time gap in deposition (18,74 ka above and radiocarbon dead below) (Fig. 12d). The 18.74 ka age above the boundary surface indicates the maximum age (oldest age) of SL2. The radiocarbon age from core P02, determined from the hemipelagic interval above the landslide deposits, constrain a minimum age of 13.87 ka for SL2. Core P14 constrains the minimum and maximum ages of the landslide SL3 to 11.35 ka and 19.31, respectively. With regards to SL4, the age of the sediment cover overlying the slide deposits dates back to 12.2 ka, indicating that SL4 is older than this age. When all dates are considered, the submarine landslides appear to have occurred between ca. 11 ka and ca. 19 ka. The age of sand layers (ca. 30.87 ka in core P15 and 40.46 ka in core P12: Fig. 12b) obtained upslope of the landslide scars provides an age for the sediment but is not conclusive when providing a deposition age as these sand layers have been remobilized downslope and redeposited in its current location. ^{14}C dating of sediments below the sand layer in P12 also presents a time reversal.

4.4 Slope stability analysis

Our slope stability model calculations from sediment core P10 above the landslide scarp of SL1 (Fig. 13a-b) show that FOS values generally decrease with depth and become asymptotic at values higher than 5, well above 1, implying that the present-day slope is exceptionally stable under static conditions ($\text{FOS}_{\text{static}}$). The relatively high values in the shallow sediments result from the low driving stresses that make the denominator in eqs. 1 and 2 very low and can be considered apparent.

The earthquake acceleration would provide an additional dynamic driving force that could push a slope that is stable under static conditions towards failure. The earthquake effect is often expressed as peak ground acceleration (PGA). Consequently, the question is: Which earthquakes could induce sufficiently large ground motions (or PGA) to drive the FOS towards unity?. A comparable PGA would be produced either by a large distant earthquake or by a smaller nearby earthquake according to the damping formula of Campbell (2003). Our model calculation shows that for 3° slope, a large earthquake (Mw8; PGA=0.9g; $k_e=0.4725$) within 3 km radius of the slide is not sufficient to destabilize the slope sediments ($FOS_{eq}>1$; Fig. 13a-c). However, assuming a %50 overpressure plus the earthquake shaking (Mw8; FOS_{eq} with overpressure, Fig. 13a) would be enough to drop FOS to less than 1.

At present, an earthquake magnitude of (Mw8) within 30 km of the slope area cannot produce a failure ($FOS_{eq}<1.5$; Fig. 13b-c). However, when including 50% pore pressure (assumed to reduce the undrained shear strength by a factor $1-r_u$ where r_u is $u_e/(\gamma'z)$ and u_e is the excess pore pressure) plus the earthquake (FOS_{eq} with overpressure, Fig. 13b), the FOS falls below 1 within the subsurface at a depth of 35 m. These results indicate that even with a considerable ground acceleration (Mw8; $k_e=0.4725$) we cannot bring the slope to failure as the FOS is still high ($FOS_{eq}>1$). Hence, additional factors (i.e., overpressure) is necessary to set off landslides at present. With regards to past slope failures that occurred at these exact locations, we consider that they were most likely controlled by other factors (i.e., excess pore pressures, weak layers), which are discussed below.

5. Discussion

Our data has identified four large submarine landslides based on i) morphological criteria (i.e., major arcuate-shaped headwall scarps on the bathymetry) in the slope regions, ii) chaotic- to transparent-seismic reflections at the base of the headwall scarps that are interpreted as mass-transport deposits (MTDs) (e.g., slides/slumps, debris flow deposits, and turbidites; Damuth and Olson, 2015; Moore and Sawyer et al., 2016), iii) presence of stacked MTDs in the Onnuri Basin, and iv) MTDs penetrated

by the piston cores. The presence of headwall scarps on the multibeam bathymetry suggests that at least four large-scale slope failures have occurred along the mid- to lower slopes. Major landslides chiefly initiated at water depths of 400 m and appear to have traveled for about 40 km to the east. By far, the most impressive slope failure is the SL2 event resulting in spectacular debris flow deposits. MTDs derived from SL1 and SL2 have accumulated in the Onnuri Basin with a thickness of about 100 m (Fig. 6). The occurrence of stacked, regionally extensive MTDs in the Onnuri Basin indicates that this was an important process during the evolution of the NE margin of the Korean Peninsula.

5.1 Comparative features of the landslides

All four landslides in our study area are characterized by planar glide surfaces that suggest the dominance of translational slides over rotational (curved slip plane) failures. The subbottom chirp profiles reveal that high-amplitude continuous reflectors characterize the glide planes. Similarly, in the literature, the high-amplitude continuous reflectors at the base of slides or chaotic seismic facies imaged on the 3.5 kHz acoustic stratigraphy are interpreted as glide planes (also known as failure planes or slide planes) (e.g., Baeten et al., 2014; Normandeau et al., 2019). The high-amplitude reflection character of the glide plane can be explained by several factors (Cukur et al., 2016). First, the significant acoustic impedance contrast may be caused by abrupt lithological or diagenetic change along this surface. Second, the flow at the top of the surface may have compacted it, thereby increasing its density and, consequently, its impedance. Finally, the slide may have left rubble on the surface that creates substantial sound diffraction resulting in strong reflection. Based primarily on their widespread occurrence, slope-parallel attitude, high-amplitude character, as well as the evidence from cores, we interpret these horizons as coarse-grained sandy layers. Previous studies from the Mediterranean Sea also show that sand to silt layers typically produce strong impedance contrast in high-resolution sub-bottom chirp profiles (Rebesco et al., 2000).

The presence of stepped-headwalls, along with several slope-parallel glide planes (Fig. 4b-c), further indicates retrogressive type failure. Retrogressive failures occur either due to subsequent seismic shaking or high static stress concentrations on the slope after the earthquake, resulting in further slope

failure and movement of sediment away from the scarp (Edwards et al., 1993). Many submarine landslides identified on continental margins are translational and developed retrogressively during multiple episodes of slope failures, e.g., the Hinlopen Slide on the Arctic Ocean Margin (Vannesta et al., 2006), the Mauritania Slide Complex on the northwest African continental margin (Antobreh and Krastel, 2007), and the Storegga Slide in the Norwegian Sea (Haflidason et al., 2004).

In addition to the numerous stepped-headwall scarps, other potential failure blocks have been identified on the slope immediately upslope of the SL1 in water depths of 510 m (Fig. 4c). These blocks may represent material that has previously failed as part of previous mass wasting events and may be reactivated in the future. There is also evidence of crown cracks on the MBES data (Fig. 4d), which typically indicate incipient failure that could be ongoing in the region (Micallef et al., 2007).

Piston cores reveal two distinct boundary surfaces in cores P03 and P11, which may represent failure scar surfaces from which slabs or near-surface sediment were removed during landsliding. One landslide wall core shows a ten- to eleven-fold increase in undrained shear strength (e.g., from 5 kPa to over 55 kPa) at a depth of 4.0 m that is interpreted to represent a failure scar surface. MTDs typically show higher undrained shear strengths and densities than non-failed sediments (see Cores P02 and P14; Figs. 10-11). Sawyer et al. (2009) show that the ultimate deposits (MTDs) have characteristically low porosity (denser) relative to bounding sediment and suggest that the original soil fabric is remolded by the failure process and results in a shear-induced fabric. Similarly, MTDs from the Big'95 slide on the Ebro margin, northwestern Mediterranean, have a lesser amount of water and display high shear strengths and densities than the non-failed sediments (Urgeles et al., 2003).

In all submarine landslides, the resulting MTDs appear to have partly- or totally-disintegrated, a process that is supported by our sediment core and seismic data. For instance, inclined beds and mud clast mud units (MCM facies) recovered in piston cores immediately below the headwall scarps of SL2 and SL3 are diagnostic of slide and debris flow deposits, respectively. The presence of relatively thick (up to 100 m), acoustically transparent sedimentary units in the deep basinal areas such as in the Onnuri Basin (Fig. 6) may suggest that major parts of MTDs associated with SL1 and SL2 evolve

downslope into slides and debris flow deposits. To achieve a disintegrative failure, the sediment must lose significant strength during a transient loading event (i.e., earthquakes). The mass-flows that remain on the slope below the headwall scarps of SL1 and SL2 form rough and irregular patches on the seafloor.

Seismic and MBES data reveal that there is a continuum in disintegrative behavior along a single failure and in between failures. The presence of intact sediment blocks located relatively close to the southern headwall of SL2 (Fig. 5c-d) suggests, for example, that they did not fully disintegrate and that the blocks moved relatively slowly with little internal deformation. These types of sediment blocks are characteristics of failure of sediment into coherent blocks displaced downslope along a planar glide plane (e.g., Micallef et al., 2007; Baeten et al., 2013; Normandeau et al., 2019). The MTDs derived from SL3 also show little deformation, which is reflected in deposits found in sediment cores, i.e., deformed beds in core P14 (Fig. 9). Compressional features (i.e., folds and thrusts; Fig. 7) are also only recognizable in SL3, which was likely the result of loading by the MTDs. On the other hand, SL4 completely disintegrated, producing a very a fast-flowing flow, as seen by the fact that the failed sediment completely evacuated the failure scar. Along SL4, there is no evidence of any large slabs that have survived across the slope or arrested close to the scarps. These differences in the post-failure styles of mass movements in the northern and southern part of the margin may result from differences in the seafloor gradient and sediment composition (Sawyer et al., 2012; Moore and Sawyer, 2016): the southern submarine landslides (SL1 to SL3) derived from fine-grained clay-rich sediments, while SL4 derived from slightly coarser, less cohesive sediments that may have been more readily dispersed to form a turbid flow. Sawyer et al. (2012), based on the experimental mudflows, show that a variety of post-failure behavior can be produced with different combinations of lithology and water content. They demonstrate that coarse-grained or sandy layers rapidly disintegrate and form a turbidity current. The blockier slides, on the other hand, are produced with lower starting water content and/or much more clay content, consistent with our interpretations. Gradients in the northernmost area also exceed 3.5°, whereas gradients are lower (<1.5°) in the southern regions. It might well be that the potential for flow mobilization was probably higher for the more steeply

sloping northern areas than for the southern area. This further implies that the failed section of SL4 must have been unable to withstand the stress of movement.

5.2 Timing of submarine landslides

¹⁴C age dating from cores P11, P02, and P14, which retrieved hemipelagic drape overlying SL1, SL2, and SL3, indicate that all landslides occurred between 11 ka and 19 ka. These findings suggest that these landslides could have happened during either the LGM low stand or early phase of transgression (sea level rise). The hemipelagic samples recovered directly above and below boundary surface in the in SL1 and SL2 present a noticeable hiatus in deposition (ca. 19 ka above and radiocarbon dead below), also suggesting geologically young slides (Figs. 12c-d) The individual age of landslides are close to each other, but with the available data, we cannot determine if they have been triggered at the same time or sequentially. Hemipelagic sediments in the topmost units above the MTDs further indicate that mass- wasting on the slope terminated at ca. 11 ka.

5.2 Some possible causes of landslides

5.2.1 Preconditioning factors

As shown in figure 13, near-surface sediments on the southwestern continental slope of the East Sea are extremely stable under static and seismic loading. This finding points to additional factors such as excess pore water pressures (and resulting low sediment strengths) or weak layers at the times of failure occurrence. Submarine landslides occur where preconditioned (weak) sediments are subjected to triggering events (i.e., earthquakes; Hampton et al., 1996), but the magnitude of such event does not need to be necessarily intense if the slope is sufficiently preconditioned for failure. Factors that can precondition slopes for failure include changes in slope sedimentation rate, sediment texture, tectonic steepening, and sea level (Hampton et al., 1996; Brothers et al., 2014).

5.2.1.1. Slope gradients and tectonic steepening

Slope gradients are obviously one of the factors controlling slope instability, as shown in Eqs. 1 and 2.

Bathymetric data from the area show that slope angles in areas affected by landslides are very gentle, ranging from 1.5° to 4° (Fig. 3b). Nevertheless, some landslide scars (i.e., SL2 and SL3) seem associated with shallow subsurface faults and folds where reflectors and the overlying seafloor abruptly increase their dip (Figs. 5 and 6). Therefore, tectonic steepening might have played a role in landslide initiation. Similarly, submarine landslides on the North Ecuador, South Colombia Convergent Margin, were hypothesized to have been facilitated by tectonic steepening associated with uplift along thrust faults (Ratzov et al., 2007). The absence of a local relief with SL1 and SL4 may, however, suggest that tectonic oversteepening is not responsible for these failures.

5.2.1.2. Glacial/interglacial cycles

Glacial/interglacial cycles may also produce changes in sedimentation rates, generally higher during glacial periods and/or dissociation/dissolution of gas hydrates caused by changes in ocean temperature and/or relative sea-level (Maslin et al. 2004; Sultan et al., 2004a; Sultan, 2007; Brothers et al., 2014). Lee et al. (2010) show that landslide events in the Ulleung Basin are more frequent during the most recent glacial period (ca. 23 to 12 ka), while the frequency appears to have declined substantially during the Holocene (<~12 ka). They suggested that high pore pressures caused by dissociation of gas hydrate during the LGM may have promoted submarine mass movements. Similarly, for the Cape Fear slide on the eastern US Atlantic margin (Hornbach et al., 2007) and Storegga Slide on the mid-Norwegian margin (Sultan et al., 2004b), gas hydrate dissociation driven by glacioeustatic sea-level changes is believed to have played a significant role. This process, however, can be excluded in our study area as multichannel seismic data do not show evidence for reverse-polarity bottom-simulating reflector (BSR), which is an indicator of gas hydrate. Also, a recent analysis by Urlaub et al. (2013) did not find any apparent link to the climate cycle that might implicate large scale triggering of slope instability due to gas hydrate dissociation.

Shallow (biogenic) gas could also increase the pore pressure and impact the stability of the slopes, causing near-surface failures of the slopes (Sultan and Garziglia, 2014). Shallow gas is usually identifiable in high-resolution seismic profiles by enhanced reflections, gas chimneys, and pockmarks

(Judd and Hovland, 1992). Seismic data contain gas chimneys and associated pockmarks near the headwall scarps, indicating local gas charging (Fig. 14). The upward flow of these fluids could horizontally fill in and pressurized some of the shallow layers or glide planes. Lowstand conditions could actually result in a decrease in total stress leading to gas exsolution and increased pore pressure, which could explain the timing of the observed landslides.

During the low stands, sediment delivery to the continental shelf edge and the slope is expected to increase along both glacial and non-glacial margins providing a source of potentially unstable sedimentary deposits (Brothers et al., 2014; Lee, 2009). High sedimentation rates or rapid sedimentation can generate elevated pore fluid pressure and trigger slope failures along the slopes (Dugan and Sheahan, 2012). Along the shelf of the southwestern margin of the East Sea, there are only a few small streams, implying a relatively small amount of sediment input into the eastern shelf during the last glacial period (Chough et al., 2000; Cukur et al., 2016). This is also supported by ^{14}C dating from our piston cores, where sedimentation rates along the unfailed slope regions are exceptionally low (less than 7 cm/kyr) (Fig. 12a). Thus, slope destabilization due to loading by increased sedimentation rates can be ruled out as a preconditioning mechanism for the landslides observed in our study area.

5.2.1.3. Weak layers

Changes in spatial and temporal heterogeneity in sedimentation on the slopes, due to climatic changes in the hinterland, tectonic activity, or sea-level fluctuations, can significantly precondition sediments for failure (e.g., Laberg et al., 2005; Badhani et al., 2020). For example, coarser-grained sediments, interpreted as grain flow deposits (Facies MgmS; Fig. 9) are thought to have been deposited during mass-movements and are found in the sediment core across the gentle slope in our study area (Core P14; See Fig. 2 for location). These coarse-grained sediments are considered as potential weak layers because of their mechanical behavior when subject to earthquake ground motion (Sultan et al., 2004a). These loosely packed, higher-permeability units may collapse when stressed by an earthquake, the effective stress in between sediment grains be lost, resulting in liquefaction (Sultan et al., 2004a). As

the coarse-grained deposits are sandwiched between low permeability clay or silt layers, it is likely that pore fluids would not be able to easily migrate vertically, and thus, they would remain trapped inducing the development of a weak layer, facilitating failure. The laterally extensive weak layer could lead to nearly instantaneous collapse and disintegration of large slabs of strata. Contourites and tephra layers have similarly been implicated as a preconditioning factor as interbedded weak layers of similar nature to those described above may act as glide planes (Laberg and Camerlenghi, 2008; Harders et al., 2010; Baeten et al., 2014). Deposition within the slope was likely influenced by bottom-water currents, and a mounded sediment body to the SE of the SL1 at water depths 400-500 m has been interpreted as bottom-current deposits (Cukur et al., 2018). Hence, current circulation patterns in response to the glacioeustatic sea-level changes are likely to result in the presence of contourite deposits along the slope, particularly in areas adjacent to the local bathymetric highs caused by uplift. However, with the available data, we cannot determine if these deposits are widespread along the slope or the influence of bottom currents on submarine landslide initiation.

The existence of widespread glide planes that follow the slope stratigraphy (Figs. 5-8) in our study area suggests translational failure along lithologically defined weak layers. Similar slope-parallel glide planes have been observed elsewhere, e.g., in the Cook Strait, New Zealand (Micallef et al., 2012) and the eastern Malta Plateau (Micallef et al., 2016). These studies suggest that landslides are stratigraphically-controlled and that weak layers within the bedded marine sediments may be a factor in controlling sediment stability. Newly acquired seismic reflection profile traversing the SL1 and SL2 further show that glide planes for these two landslides correspond to the same seismic reflector (Fig. 14). This reflector must represent a lithological weak layer in the sub-surface sedimentary section. In the chirp profiles, the glide planes are characterized by high-amplitude continuous reflections and most likely represent coarse-grained sediments (Figs. 5-8), which are more easily activated during seismic shaking. We thus predict that many, if not all, prominent reflectors in the seismic profiles along the slope record sandy deposits (Facies MgmS in Fig. 9) resulting from mass flows from the continental shelf edge. However, little is known about the nature of these weak-layers and to properly identify them multi-disciplinary research (i.e., core and insitu testing inside and

outside failure areas, long-term monitoring, etc.) would be needed.

5.2.2. Triggering mechanism for the observed landslides

Earthquakes are an effective trigger mechanism for the initiation of slope failures (Hampton et al., 1996). We consider that because slope failures preferentially occur adjacent to the major faults on the lower slope (Figs. 15a-b), the slope failures may ultimately be controlled by tectonics, although other factors may have contributed as well. Fault displacements exceeding 200 m in vertical throw are recorded along several of the seismic profiles. Recent studies also suggest that the compressional tectonism of the southwestern margin of the East Sea margin developed at ca. 3 to 4 Ma and remained active until today as judged by their deformation of the seafloor (Cukur et al., 2018). Ongoing tectonic activity is also evidenced by the occurrence of numerous, medium-scale earthquakes ($2.5 < M_w < 5.4$) on the slope adjacent to the slope failures (Fig. 15c). However, the magnitude of such earthquakes is rather small to initiate a failure, as shown by our slope stability analysis as well as the presence of hemipelagic sediments in the cores since ca. 11 ka. A crude investigation of seismic loading on the slopes indicates that at least $M_w 8$ would be required to destabilize the slopes of the seafloor in the present study. Such a large earthquake may have occurred in the vicinity of landslides caused by the reverse fault (Fig. 15). The length of the surface trace of this fault is approximately 50 km (Fig. 15), corresponding to a $\sim M_w 7$ earthquake (Wells and Coppersmith, 1994). This shallow, moderately active fault (Fig. 13c) most likely generated bigger ground displacement (equivalent to a distant $M_w 8$ deeper earthquake) immediately beneath the study area that likely triggered the translational landslides. A recent study by Hong et al. (2020) suggests that earthquakes with such magnitudes (e.g., $M_w 7$) may occur in 2550 return period in the eastern continental margin of the Korean Peninsula. These low probability large magnitude earthquakes may have triggered the observed landslides in our study area. Similarly, Kvalstad (2005) concluded that a significantly large earthquake with a low probability of occurrence was the only feasible source of the required loading for the enormous Storegga Slide on a 1° slope.

Sawyer and Devore (2015) have reported that active margins show significantly higher undrained

shear strengths compared to passive margins. Seismic strengthening, which enhances the slope stability, is suggested to explain the observed paucity of submarine landslides on active margins when compared to the short recurrence of strong earthquakes in such a setting. The eastern continental margin of Korea seems passive, with only a few moderate earthquakes ($M_w > 5$; Hong et al., 2020), yet surprisingly there are numerous landslides, as demonstrated by the presented data. Because there seems to be no significant seismic strengthening in the Ulleung Basin, unlike the active margins proposed by Sawyer and Devore (2015), large, infrequent (every ten thousand years or so) earthquakes, in combination with other mechanisms such as pore pressure could likely represent the final trigger mechanism of the observed instabilities.

5.2.3 Tsunami potential of the observed landslides

Submarine landslides with volumes as small as 0.01 km^3 can generate significant tsunamis (e.g., The 1979 Nice event; Assier-Rzadkiewicz et al., 2000). Models of the particularly well known 1998 Papua New Guinea landslide (ca. 4 km^3 ; Synolakis et al., 2002) show that failure of this slide was capable of generating a tsunami with heights of up to 15 m. Another example of a medium scale landslide (ca. 1.7 km^3) comparable to those in our study area that triggered a tsunami (up to 10 m in height) occurred in Santa Barbara Channel. More recently, Urgeles et al. (2019) have done numerical modeling of individual landslides along the eastern coast of Korea and show that landslides with volumes of ca. 2.5 km^3 can generate a tsunami with a height of $< 2 \text{ m}$. The size of submarine landslides (ca. 2 to 8 km^3) in our study area suggests that they have the potential to be tsunamigenic. However, many other parameters, such as initial acceleration, flow dynamics, water depth, and velocity, may also be important (Harbitz et al., 2006). Hence, more comprehensive numerical modeling is necessary to assess the tsunami potential of these landslides.

5.3 A depositional model for preconditioning and triggering submarine landslides on the SW margin of the East Sea

We highlight that a single cause for the slope instability and actual triggering mechanism is often

unlikely for submarine landslides on the low-angle, sediment starved slopes. Instead, as discussed above, a combination of various preconditioning and triggering mechanisms is necessary to initiate landslides. We have now, based primarily on the analysis of piston cores and seismic data, develop a schematic model that shows the interrelationships of various factors that led to the slope failures. (1) Coarse-grained clastic sediments, derived from the shelf edge by mass-movement processes, were deposited on the slope (Fig. 16a). Further sedimentation buried the deposited coarse-grained material beneath fine-grained hemipelagic sediments. (2) Excess pore pressure developed within the coarse-grained sediments may have preconditioned the slope for failure, followed by a final trigger of a significantly large earthquake (Fig. 16b). The resulting MTDs derived from mass failure transformed into debris flows and accumulated downslope of the headwall scarps; some have flowed further downslope and filled the topographic lows (i.e., the Onnuri Basin; Fig. 6). (3) Sediment redistribution by mass-wasting terminated at ca. 11 ka, and since then, hemipelagic sedimentation has played an important role in shaping the sediment-starved slope (Fig. 16c).

6. Conclusions

Seismic, sub-bottom chirp profiler, bathymetric, and piston core data reveal that mass movements are the dominant process modifying the southwestern margin of the East Sea, Korea. At least four submarine landslides (SL1 to SL4) and many more MTDs are documented in multibeam bathymetry, seismic reflection, and sediment core data. These landslides originated on the continental slope in water depths greater than 400 m and left very clear scarps on the seafloor ranging from 40 to 130 m in relief. Downslope from the landslide scarps, well-preserved MTDs characterized by hyperbolic surface and chaotic- to transparent-internal nature exist. Cores taken from downslope of landslide scarps confirm the presence of MTDs and contain a fabric representative of MTDs, such as deformed beds, turbidites, and debrites. One sediment core taken from the slide wall of the SL2 reveals a sharp boundary surface, which we interpret to be a failure scar surface. This boundary present as a sharp color-change across the surface and a marked increase in sediment undrained shear strength. The

arcuate-shape and stepped headwalls and sidewalls of the scarps point to retrogressive-type failures. SL2, covering the seafloor up to $\sim 280 \text{ km}^2$, a volume of $\sim 8.4 \text{ km}^3$, a headwall scar over 130 m high, and a run-out distance of 40 km is the largest of all failures in the study area. In all submarine landslides, MTDs are partly- or totally- disintegrated. Seismic data suggest that significant parts of MTDs associated with SL1 and SL2 deposited in the Onnuri Basin. Radiocarbon dates atop the slide deposits, presumed to be post-failure deposits, indicate that major failures took place at about ca. 11-19 ka. Since ca. 11 ka, the slope became stable, and hemipelagic sedimentation has dominated the area. The largest submarine landslides on the slope occur immediately next to the seafloor-breaching major faults, suggesting an association between these features and triggering of the slope failures. These faults appear to have remained active until today, as shown by the occurrence of earthquakes. Stability analyses based on geotechnical properties of sediments indicate, however, that the slopes should be stable under static and earthquake loading assuming hydrostatic conditions. Therefore, additional factors must act to reduce the shear resistance of these sediments to values low enough to enable slope failures. The existence of mechanically weak layers combined with tectonic steepening and gas charging is thought to be significant factors facilitating the onset of past failures.

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Figure captions

Fig. 1. Bathymetry of the East Sea, Korea, showing the location of the study area, adjacent sedimentary basins, and the major tectonic elements. The eastern continental shelf of the Korean Peninsula is relatively narrow (< 30 km), whereas the shelf to the south is quite broad (30-150 km wide). The continental margin north of 37.50° , which is the focus of our study, trends SE-NW and extends down to water depths of 1600 m. Here, the continental shelf is relatively narrow (< 10 km) with a shelf break at 200 m. Abbreviations: HB: Hupo Bank; NUT: Northern Ulleung Trough; UI: Ulleung Island; UIG: Ulleung Interplain Gap; US: Ulleung Seamount; WSKP: Western South Korea Plateau; ESKP: Eastern South Korea Plateau.

Fig. 2. Map showing the location of multichannel seismic reflection profiles (green, blue, and gray lines), sub-bottom chirp profiles (red, white, and green lines), the outer boundary of multibeam data (thick black line) and piston cores (white circles).

Fig. 3. (a) Bathymetric map collected along the southwestern margin of the East Sea. The bathymetric data reveal four major submarine landslides (Labeled SL1 to SL4) that are characterized by arcuate-shaped head scarps at the upper- to mid-slopes in between ~ 400 and ~ 600 m water depth. The red dashed lines indicate slide scar edges. (b) Slope gradient map of the study area. Slopes adjacent to the landslides have gradients in the range of $2 - 4^{\circ}$, while the slope gradients within the headwall scarps are as steep as 25° .

Fig. 4. (a) Bathymetric profile across submarine landslide 1 (SL1). The upper headwall of the SL1 is ~ 35 m high and shows a slope angle of 14° , while the average slope angle landward of the scarp is 3° . (b) Sub-bottom chirp profile crossing SL1 clearly shows mass-transport deposits (MTDs) beneath the headwall scarp that are characterized by chaotic- to transparent-acoustic facies with hyperbolic surface expression. The base of the chaotic seismic facies is characterized by continuous, high amplitude reflections that represent the glide plane. Also evident is the presence of stepped headwalls. Two cores (P10 and 11) were taken along this profile to characterize the sedimentology, age, and

geotechnical properties of sediments. (c) Sub-bottom chirp profile traversing the sidewall of SL1 showing a potential future failure block. (d) MBES data showing a crown crack, which suggests an incipient failure.

Fig. 5. (a) Bathymetric profile across SL2, displaying headwall height and slope ranging from ~130 m and 25°, respectively. (b) Sub-bottom chirp profile intersecting the SL2 show chaotic- to transparent-seismic facies downslope of the scarp that is typical for MTDs. Three piston cores (P02, P03, and P04) were collected on this profile. (c) Sub-bottom chirp profile from the southwestern corner of the SL2 showing slide blocks with stratified internal seismic character. The unfailed regions adjacent to sidewalls are characterized by well-stratified seismic reflections that are diagnostic features of undisturbed slope sediments. (d) Inset bathymetric image showing the slide blocks that are parallel to the scarp.

Fig. 6. Multichannel seismic reflection profile traversing SL2 shows that most MTDs associated with SL2 were transported over the slope and now reside on the Onnuri Basin. The presence of stacked MTDs further indicates multiple phases of mass wasting events. The small amount of mass-flow material that remains on the slope forms rough and irregular patches on the slopes.

Fig. 7. (a) Bathymetric profile across submarine landslide 3 (SL3), which has a slope of ~15° with a maximum relief of ~60 m. (b) High-resolution subbottom chirp profile showing SL2 on the seafloor, characterized by chaotic seismic facies with hyperbolic seabed that are interpreted as MTDs. (c) Multichannel seismic reflection profile traversing the SL3 shows a prominent headwall scarp and chaotic seismic facies downslope of the scarp. SL2 appears to have traveled about 8 km from the source area.

Fig. 8. (a) Bathymetric profile across submarine landslide 4 (SL4), revealing gradient and height of 20° and 65 m, respectively. (c) High-resolution chirp profile, crossing SL4, showing the 55-m-high morphological step representative of the headwall scarp. Also evident are the truncated sedimentary layers at the scarp walls; the base of the scarp is also characterized by well-stratified seismic

reflections that are parallel to the slope stratigraphy. Also shown is the location of sediment cores collected from in and outside of the SL4.

Fig. 9. (a) Selected X-ray radiographs and photographs of slab samples obtained from piston cores showing nine sedimentary facies in the study area including: (1) BM: bioturbated mud/silt; (2) FLM: faintly laminated mud; (3) MCM: mud-clasts mud; (4) DM: deformed mud; (5) WLzS/m: wavy laminated silty sand/mud; (6) MgmS: massive gravelly muddy sand; and (7) CLsM: convoluted-laminated sandy mud. (b), (c), (d) Microscopic images of selected sedimentary facies from cores P14, P15, and P12. See text for details.

Fig. 10. (a), (b) Vertical profiles of volume percent, mean grain-size, density, porosity, water content, and undrained shear strength of the sediment samples retrieved from inside and outside of the submarine landslides SL1 and SL2. Samples taken outside and inside of submarine landslides are comprised mostly of bioturbated mud and silt (BM). Mud clasts mud facies (MCM), interpreted as debris flow deposits, are found in core P02. Across all the cores, there is a slight increase in density and undrained shear strength with depth, probably associated with sediment consolidation with burial depth. Otherwise, major changes coincide with facies changes, mainly where these correspond to distinct grain size changes. Note also that all of the sediment cores show a fining upward trend. (c) The sediments recovered in core P03 at a depth of ca. 400 cm show a sharp increase in undrained shear strength (up to 60 kPa); this boundary is interpreted as a failure scar surface. Increased values of undrained shear strength below this boundary further suggest stiffer or more compacted material.

Fig. 11. (a), (b) Vertical profiles of volume percent, mean grain-size, density, porosity, water content, and undrained shear strength of the sediment samples retrieved from inside and outside of the submarine landslides SL3 and SL4. The background sediments are dominated by bioturbated mud and silt (facies BM). Deformed mud layers (DM) in P14, capped by facies WLzS, are interpreted as slide deposits. Bulk densities show greater variability in Facies MgmS and WLzS and compared with

Facies BM, and are relatively higher where the sandy sediments are higher. A general increase of densities and undrained shear strengths, as well as a down-core decrease of porosities and water contents, indicate sediment consolidation with burial depth. Note also that marked changes in physical properties correspond to the major facies changes, particularly where they are associated with significant changes in grain size.

Fig. 12. (a) Sedimentation rates determined from ^{14}C ages (see Tables 1 and 2) within 10 piston cores. The upslope areas of headwall scarps are characterized by sedimentation rates of less than 7 cm/ka, while downslope areas of headwall scarps show relatively higher sedimentation rates, ranging from 12 cm/ka to 37 cm/ka. (b) Core logs showing core photos, x-ray images, sedimentary facies, L^* values, and radiocarbon ages (ka: thousand years before present) with depth. Where a distinct boundary surface existed, this is indicated with a thick red line. Red lines across the cores mark the ca. 11 ka age boundary. When all dates are considered from cores, the most plausible timing for the landslides is ca. 19-11 ka. See the text for details. The yellow boxes mark the location of close up images shown in figure 12 (c) and 12 (d). (c) (d) Close up images show the boundary surfaces (dashed red line), inferred to be failure scar surfaces, in cores P11 and P03. Radiocarbon ages are also shown.

Fig. 13. Profiles of $\text{FOS}_{\text{static}}$, $\text{FOS}_{\text{static with overpressure}}$, FOS_{eq} , and $\text{FOS}_{\text{eq with overpressure}}$ of core P10 with a magnitude of Mw8 occurring within 3 km (a) and 30 km (b) of the SL1. From stability analyses under static conditions ($\text{FOS}_{\text{static}}$), static conditions with %50 overpressure ($\text{FOS}_{\text{static with overpressure}}$) and under seismic loading (FOS_{eq}) the resulting factor of safety (FOS) exceeds 1 in all cases, and thus favors a stable slope, given that $\text{FOS} \leq 1$ represents slope failure. However, %50 overpressure plus the earthquake shaking ($\text{FOS}_{\text{eq with overpressure}}$, Fig. 13a-b) is sufficient to drop FOS to less than 1. (c) Earthquake record of the study area since 1978. Earthquake data are from the Korea Meteorological Administration (2020). Dashed black circles indicate the distance to the SL1.

Fig. 14. MCS profile, traversing SL1 and SL2, show the presence of shallow gas indicators such as gas chimneys and pockmarks. The upward flow of these fluids could horizontally fill into the glide planes and induce overpressure. MCS further indicates that SL1 and SL2 have a common glide plane,

suggesting that this seismic reflector must represent a mechanically weak layer on which the sliding occurred. Correlated reflectors 1–4 (number 4 represents the glide plane) are labeled for ease of comparison.

Fig. 15. (a) Slope gradient map showing the location of major faults and submarine landslide scars. Thick red lines represent reverse faults. (b) MCS profile traversing the headwall scarp of SL2. The immediate occurrence of submarine landslide scarp next to the fault may suggest its origin to tectonic activity.

Fig. 16. The depositional model is showing the link between submarine landslides and slope sedimentation. (a) Pre-failure stage (ca. >19 ka) showing the normal hemipelagic deposition interbedded with coarse-grained sandy layers that were derived from the upper shelf by mass-wasting processes. (b) Triggering stage (ca. 19-11 ka) in which earthquakes trigger the landslides. Liquefaction of coarse-grained sandy layers (weak layers) in response to earthquakes could lead to the disintegration of large slabs of strata, resulting in debris-flow deposits immediately below the headwall scars. (c) The slope has been stable since the ca. 11 ka and dominated by hemipelagic sedimentation.

Table Captions

Table 1. Radiocarbon ages of 18MAP core sediments obtained from the southwestern margin of the East Sea by calibration to calendar years using CALB 6.1. Samples were dated the Arizona AMS Laboratory (Lab no.: AA)

Table 2. Radiocarbon ages of 17MAP core sediments obtained from the southwestern margin of the East Sea by calibration to calendar years using CALB 6.1. Samples were dated the Arizona AMS Laboratory (Lab no.: AA)

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Figure 1

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Figure 2

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Figure 3

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Figure 6

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Figure 11

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Table 1. Radiocarbon ages of 18MAP core sediments obtained from the northeastern continental margin of Korea by calibration to calendar years using CALB 6.1. Samples were dated the Arizona AMS Laboratory (Lab no.: AA)

Core	Core depth (cm)	AMS ^{14}C age (years BP)	2σ age ranges (a)	Calendar age, median (a)	Materials	Lab No.
18MAP-P10	94-96	11,148 $\pm$ 41	12,563 to 12,744	12,641	Benthic foraminifera	AA112274
	104-106	11,239 $\pm$ 43	12,597 to 12,871	12,704	Benthic foraminifera	AA112275
	232-234	32,890 $\pm$ 480	36,580 to 38,686	37,572	Planktonic foraminifera	AA112273
	390-392	>44,800	-	-	Benthic foraminifera	AA112272
18MAP-P11	221-222	15,413 $\pm$ 55	18,007 to 18,540	18,248	Planktonic foraminifera	AA112276
	240-242	47,100 $\pm$ 2900	-	-	Planktonic foraminifera	AA112277
	285-287	>49,400	-	-	Benthic foraminifera	AA112278
18MAP-P12	90-92	35,840 $\pm$ 740	38,859 to 41776	40,464	Benthic foraminifera	AA112281
	101-103	29,960 $\pm$ 340	33,313 to 34,799	34,151	Planktonic foraminifera	AA112282
	155-157	>49,000	-	-	Benthic foraminifera	AA112283
	186-188	>46,400	-	-	Benthic foraminifera	AA112284
	266-268	>49,900	-	-	Benthic foraminifera	AA112280
	425-426	>49,900	-	-	Benthic foraminifera	AA112279
18MAP-P13	140-142	4,873 $\pm$ 46	5,028 to 5,301	5,180	Benthic foraminifera	AA112287
	171-173	5,987 $\pm$ 30	6,305 to 6,478	6,397	Benthic foraminifera	AA112288
	266-268	9,872 $\pm$ 37	10,603 to 10,952	10,764	Benthic foraminifera	AA112286
	375-377	>49,500	-	-	Benthic foraminifera	AA112285
18MAP-P14	251-253	10,360 $\pm$ 44	11,223 to 11,629	11,349	Planktonic foraminifera	AA112289
	331-333	16,565 $\pm$ 74	18,935 to 19,455	19,306	Planktonic foraminifera	AA112290
	357-359	33,220 $\pm$ 500	36,519 to 38,689	37,500	Benthic foraminifera	AA112291
	364-366	>49,500	-	-	Benthic foraminifera	AA112292
18MAP-P15	108-110	26,580 $\pm$ 220	30,468 to 31,180	30,873	Shell fragments	AA112293
18MAP-P16	449-451	10,737 $\pm$ 73	11,884 to 12,393	12,147	Planktonic foraminifera	AA112294

Table 2. Radiocarbon ages of 17MAP core sediments obtained from the northeastern continental margin of Korea by calibration to calendar years using CALB 6.1. Samples were dated the Arizona AMS Laboratory (Lab no.: AA)

Core	Core depth (cm)	AMS ^{14}C age (years BP)	2σ age ranges (a)	Calendar age, median (a)	Materials	Lab No.
17MAP-P02	389-391	12,021 $\pm$ 53	13,745 to 14,018	13,874	Wood fragments	AA112295
	501-503	>49,900			Benthic foraminifera	AA112296
17MAP-P03	384-386	15,960 $\pm$ 66	18,599 to 18,740	18,740	Planktonic foraminifera	AA112297
	384-386	15,428 $\pm$ 59	18,821 to 18,659	18,659	Wood fragments	AA112298
	410-412	>49,900	-	-	Planktonic foraminifera	AA112299
	436-438	>49,500	-	-	Planktonic foraminifera	AA112300
17MAP-P04	236-238	29,530 $\pm$ 320	32,967 to 34,600	33,794	Planktonic foraminifera	AA112301
	256-258	33,370 $\pm$ 510	36,583 to 38,765	37,653	Planktonic foraminifera	AA112302
	333-335	>45,400	-	-	Planktonic foraminifera	AA112303
	400-402	>46,700	-	-	Planktonic foraminifera	AA112304