The influence of overpressure and focused fluid flow on subaquatic slope stability in a formerly glaciated basin: Lake Villarrica (South-Central Chile)

Jasper Moernaut^{*1,6}, Gauvain Wiemer², Anna Reusch^{3,9}, Nina Stark⁴, Marc De Batist⁵, Roberto Urrutia⁷, Bruno Ladrón de Guevara⁸, Achim Kopf², Michael Strasser¹

¹ Institute of Geology, University of Innsbruck, Innrain 52, Innsbruck, Austria

² MARUM, Centre for Marine Environmental Sciences, Bremen University, Leobener Strasse, Bremen, Germany

³ Geological Institute, ETH Zürich, Sonneggstrasse 5, Zürich, Switzerland

⁴ Department of Civil and Environmental Engineering, Virginia Tech, 750 Drillfield Drive, Blacksburg, USA

⁵ Renard Centre of Marine Geology, Ghent University, Krijgslaan 281 (S8), Belgium

⁶ Instituto de Ciencias de la Tierra, Universidad Austral de Chile, Campus Isla Teja, Valdivia, Chile

⁷ Centro EULA, Universidad de Concepción, Casilla 160-C, Concepción, Chile

⁸ BENTOS, Av. Suecia 3005, Santiago, Chile

⁹ Faculty of Geosciences, University of Bremen, Klagenfurter Straße, Bremen, Germany

*Corresponding author: Jasper.Moernaut@uibk.ac.at

Abstract

Pore fluid pressure is a key parameter for subaquatic slope stability and has been put forward to explain the development of many submarine landslides, especially when occurring on very gentle slopes or in areas with high sedimentation rate. Due to the sparse availability of in-situ pore pressure data, a profound understanding and quantification of excess pore pressure development and its influence on subaquatic landslide processes is still missing. In this study, we use glacigenic Lake Villarrica (Chile) as a model basin, in which we document in-situ excess pore pressure, focused fluid escape features and subaquatic slope failures at relatively shallow subsurface depth (< 10 m). A subaquatic slope was characterized in great detail by a dense network of high-resolution seismic-reflection profiles, multibeam bathymetry and sediment cores. In-situ undrained shear strength and formation pore fluid pressures were documented using free-fall piezocone penetrometer (CPTU) and subsequent pore pressure dissipation tests. We show four independent lines of evidence for

overpressure of variable magnitude within the sedimentary slopes: i) pockmarks and other fluid escape features on seismic profiles and bathymetric maps, ii) estimation of ambient in-situ pore pressure via CPTU dissipation tests, iii) underconsolidation and downward decrease in in-situ undrained shear strength within units of uniform lithology, and iv) hydrofractured glacio-lacustrine sediments. Locally, overpressure reaches about 80-90% of the hydrostatic vertical effective stress. We hypothesize that glacier-proximal sediments are strongly underconsolidated and overpressured due to rapid deposition and/or subglacial meltwater pumping. Postglacial consolidation of a thick glacier-proximal unit leads to the upward expulsion of excess pore water. A permeability barrier is formed by the overlying glacio-lacustrine clays and fine silts, focusing these migrating fluids to topographic highs. There, very high overpressure ratios facilitate the development of hydro-fractures, repeated fluid escape and pockmarks on the lake bottom. Overpressure may decrease slope stability in Lake Villarrica in different manners: by focused fluid escape weakening sediments, by generating underconsolidated sequences of low undrained shear strength, and by facilitating static failure. Based on high-quality in-situ data, this study confirms that high overpressure ratios in shallow sediments may be a typical feature at formerly-glaciated marine and lacustrine environments and highlights the major role of pore fluid overpressure for preconditioning subaquatic slopes to failure.

Keywords

Subaquatic landslide, excess pore pressure, slope stability, deglaciation depositional history, subsurface fluid flow

Highlights

- In-situ geotechnical data reveal severe shallow overpressure in a subaquatic slope
- Spatial variation in overpressure ratio determines subaquatic slope failure location
- Deglaciation depositional history leads to underconsolidated units and permeability barriers
- Subsequent consolidation and flow focusing determine overpressure distribution

1. Introduction

One of the key challenges in submarine landslide research is to understand and quantify the role of different preconditioning factors that affect subaquatic slope stability. Pore water overpressure is regarded as one of the most

important preconditioning factors, as it decreases the effective stresses and, thus, the resisting forces of a sedimentary sequence against downslope driving forces. Overpressure has often been invoked to explain the initiation and retrogressive behavior of extensive submarine landslides on low ($<3^\circ$) slope gradients (e.g. Field et al., 1982; Bryn et al., 2005), the process of which has been experimentally confirmed (Vendeville and Gaullier, 2003; Lacoste et al., 2012; You et al., 2012). However, in-situ pore pressure data is rarely measured in submarine landslide studies (Vanneste et al., 2014). Therefore, it is difficult to assess the impact of pore pressure on subaquatic slope stability and the initiation of landslides in natural settings.

It was thought for a long time that overpressure in fine-grained material could only exist at significant subsurface depths (>1.5 km; Mello et al., 1994). During the last decade, laboratory consolidation experiments (e.g. Dugan and Germaine, 2008) and the installation of in-situ piezometers and pore pressure penetrometer tests (e.g. Flemings et al., 2008; Lafuerza et al., 2009; Stegmann et al., 2011) reveal more and more evidence of “shallow” overpressure, i.e. within the uppermost hundreds of meters. In young sedimentary basins, the main process for shallow overpressure generation is considered to be disequilibrium compaction due to rapid sedimentation of compressible low-permeability material, which restricts the drainage of excess pore water. Consequently, the pore water supports some of the overburden weight and pore pressure rises above hydrostatic values, called “overpressure” (Dugan and Sheahan, 2012 and references therein). Spatial differences in burial rate can result in overpressure gradients, causing lateral flow of pore water through high-permeability layers from areas with a thick overburden to those with thinner overburden (Dugan and Flemings, 2000). Hence, shallow overpressures can also prevail in areas of low sedimentation rate as a result from lateral (and vertical) pore water transfer.

When such pressure-gradient-driven fluid flow breaches the seafloor surface, it may create characteristic morphological features such as pockmarks and mud volcanoes. Consequently, these features are often used as indicators of overpressured fluids (water and/or gas) in the underlying sedimentary sequence (Hovland et al., 2002). A link between focused fluid escape and landslide initiation was suggested in several studies by i) the presence of pockmark fields and mud volcanoes in and near the failed slope (Piper et al., 1999; McAdoo et al., 2000; Naudts et al., 2006), ii) headwall scarps that spatially link with presumed pockmarks (Lastras et al., 2004; Brothers et al., 2014) or iii) reflection-seismic facies anomalies in landslide areas, such as enhanced reflections and chimneys of acoustic blanking, respectively indicating gas accumulation and vertical transfer of fluids (Bünz et al., 2005; Gee et al., 2007; Sultan et al., 2010; Berndt et

al., 2012; Riboulot et al., 2013; Brothers et al., 2014). If fluid flow is a persistent or recurrent phenomenon, it may allow for repeated slope failure at a single location, as is hypothesized for example at the “Ana Slide complex” in the Eivissa Channel, western Mediterranean Sea (Berndt et al., 2012). There, the presence of free gas at the time of failure (61.5 ka BP) was attested by $\delta^{13}\text{C}$ measurements on benthic foraminifera (Panieri et al., 2012), and is assumed to be a major factor leading to slope instability. However, in the present-day sequence sediment cores or penetrometers failed to provide evidence of gas-charged sediments (Lafuerza et al., 2012). This contrast illustrates the difficulties related to reconstructing past slope stability conditions for the moment during which failure took place. Therefore, it is believed that the best understanding of slope failure processes can be gained by investigating relatively young (Holocene) landslide events in regions in which the depositional patterns have not changed significantly since failure.

Lake basins can offer this opportunity, as “recent” landslides of different sizes are common components of their sedimentary infill (Karlin et al., 2004; Moernaut and De Batist, 2011; Waldmann et al., 2011; Strasser et al., 2013; Kremer et al., 2015; Praet et al., 2016). Their small dimensions and shallow water depths allow for very-high-resolution datasets and the critical layers that acted as sliding surfaces during slope failure can be accessed with relatively simple and cost-effective methods (e.g. Stegmann et al., 2007; Strasser et al., 2011; Wiemer et al., 2015). Nevertheless, lacustrine slope stability studies and characterization of in-situ pore pressures are still very scarce.

In the present case study, we evaluate the influence of excess pore pressure and focused fluid escape on subaquatic slope stability in a Chilean glacigenic lake. We characterize the morphology, stratigraphy, in-situ shear strength and pore pressure state of an intact sublacustrine platform surrounded by failed slopes. Based on these data, we further discuss excess pore pressure development and landsliding in formerly glaciated basins or along formerly glaciated continental margins.

2. Setting and previous studies

2.1. General setting

Lake Villarrica (39°15'S; 72°02'W; 214 m asl) is a large (21 km x 9 km) glacigenic lake with a maximum depth of 167 m (Fig. 1, 2). It is located at the western piedmont of the volcanically-active Chilean Andes in a region that has experienced

strong and prolonged ground shaking during several subduction megathrust earthquakes (Moernaut et al., 2014; Van Daele et al., 2015). The most notorious “recent” seismic events happened in AD 1960 (M_w 9.5) along the Valdivia Segment and in AD 2010 (M_w 8.8) along the Maule Segment (Fig. 1). The Villarrica lake basin originated from glacial valley overdeepening and the formation of successive frontal moraine belts during the Late Quaternary glaciations (Laugenie, 1982). Actually, the lake basin consists of a deep central basin and a shallower area in its southwestern part, which comprises several rock basement highs, moraine ridges and sedimentary sub-basins (Fig. 2, 3A; Moernaut et al., 2009). The main river in- and outflow are located at its eastern and western extremities, respectively.

In the southwestern area, the irregular basement morphology is directly overlain by a complex and thick sedimentary infill (up to ~120 m thick), interpreted as being deposited during the early stage of deglaciation (Moernaut et al., 2009). This thick infill is covered by a more continuous and thinner drape (10-20 m) of hemipelagic sediments, deposited in an open, post-glacial lake environment (Heirman, 2011). The recent background sediments can be classified as diatomaceous ooze, with a small fraction of terrestrial organic matter, dispersed volcanic ash and terrigenous clays/silts. These sequences are frequently interrupted by thin tephra-fall layers and turbidites related to lahars and earthquakes (Van Daele et al., 2014, 2015; Moernaut et al., 2014, 2015). Seismic-stratigraphic analysis revealed that several voluminous landslides ($1-6 \times 10^6 \text{ m}^3$) took place on the sedimentary slopes of the southwestern extremity of the lake and were deposited on top of each other in the sub-basins (Moernaut et al., 2009).

2.2 Previous studies and open questions

Our study target is a small sedimentary “platform” which is almost completely surrounded by headwall scarps of up to 10 m high (Fig. 3). Previous studies documented that a large landslide took place at ~1.6 kyr BP, which remobilized about $6.3 \times 10^6 \text{ m}^3$ of sediment (Moernaut et al., 2009; Wiemer et al., 2015). The landslide propagated up to 5.5 km eastwards of the headscarp (mass-transport deposit MTD-a1). About 1-2 km to the north, another voluminous mass-transport deposit (MTD-a2; $3.6 \times 10^6 \text{ m}^3$) is located at the same seismic-stratigraphic level as MTD-a1. Moernaut et al. (2009) interpreted these landslides as being triggered simultaneously and suggested strong seismic shaking as the most likely trigger mechanism. Two smaller landslides in the study area were attributed to the AD 1960 earthquake (Moernaut et al., 2015).

The main sliding surface of event MTD-a1 consists of a continuous high-amplitude reflector (R2 on Fig. 4). The headwall geometry of the slide is relatively complex and forms a bottle-neck geometry in some places. Wiemer et al. (2015) explain this morphological complexity by retrogressive failure within quick clays, and speculate that focused fluid escape may have locally weakened the sedimentary sequence. However, these authors had not mapped the headscarp in detail and did not systematically explore the sub-surface using seismic profiles to evaluate if subsurface fluid flow exists, and how this eventually may have influenced slope failure processes.

Geotechnical and sedimentological studies were conducted by Wiemer et al. (2015) on the intact (unfailed) platform (test site CPT1.2 / VILLAR1) in order to characterize the sedimentary sequences involved in past slope failure: Sediment cores consist of three main depositional units (Fig. 4), which represent the Holocene and early post-glacial sediment drape. The uppermost unit, U1, is composed of sandy-silty diatomaceous ooze, U2 of silty diatomaceous ooze and U3 of detrital fine silt/clay. U3 is a quick clay with a very high sensitivity (> 300) caused by a low remolded strength. One-dimensional compression tests and normalized undrained shear strength data show that U1 and U2 are “apparently overconsolidated” (terminology after Sultan et al., 2000) and that the overconsolidation ratio (OCR) decreases downcore from U1 (OCR = 8.05 at 2.0 m meters below lake floor (mblf)) to U2 (OCR = 2.0 at 6.4 mblf), and to U3 (OCR = 1.2 at a projected 9 mblf). Hydraulic conductivity (k_f) is 1.1×10^{-7} , 1.0×10^{-7} and 6.1×10^{-8} cm/s for U1, U2 and U3, respectively. The permeability of U3 is in the typical range for clay or silty-clay, and about one order of magnitude less than U1 and U2 (Wiemer et al., 2015).

Wiemer et al. (2015) characterized the unfailed platform at three sites by free-fall piezocone penetrometer tests (CPTU) to determine the in-situ undrained shear strength (s_u) of the sediment sequence. Based on these values, they reconstructed slope stability and failure under static gravitational and earthquake loading conditions. One of the key points in the slope stability model is that U2 exhibits slightly decreasing values of in-situ undrained shear strength down to the lowermost U3, where values are as low as at about 0.5 m below the lake floor. This is clearly contrasting with the generally-expected linear increase of undrained shear strength with depth (Karlsson and Viberg, 1967; Bartetzko and Kopf, 2007). It was proposed that this decreasing trend results from excess pore pressures. However, without direct measurements of pore pressure, such hypothesis remains speculative.

The present study investigates the possible presence of excess pore pressure and focused fluid escape in Lake Villarrica's slope sediments, and how this could have affected slope stability. The geometry of the headwall and the shallow subsurface are documented in great detail by integrating an extensive dataset of new multibeam bathymetry and pinger seismic profiles. This allows for the identification of fluid escape features and reveals how these relate to the stratigraphic architecture of the subsurface. Sparker seismic data visualize the deeper sedimentary infill, allowing the reconstruction of sedimentation patterns since the beginning of the last deglaciation. Moreover, in-situ piezocone dissipation tests are introduced for estimation of ambient pore pressure values. The in-situ shear strength data of Wiemer et al. (2015) is spatially complemented with i) new CPTU data obtained in different settings (i.e. failed slope, unfailed slope, fluid escape areas and deep basin), and ii) a sediment core located very close to the studied headscarp.

3. Methods and Materials

3.1. Bathymetry and sub-surface imaging

High-resolution bathymetry data was acquired in January 2013 using an R2Sonic 2024 multibeam system (200-400 kHz) of the survey company BENTOS (Santiago, Chile). Navigation was recorded by real time kinematic (RTK) corrected GPS with horizontal accuracy of a decimeter (OmniSTAR G2). Swath angle and pulse frequency was adapted on-line in function of water depth. Data processing (corrections for sound velocity, vessel motion and orientation, sensor offsets, spike removal, etc.) and visualization was carried out using CARIS HIPS and SIPS 7.1 and standard GIS software. All presented bathymetric maps have a horizontal grid size of 3 m.

Dense networks of reflection seismic data (SI-Fig. 1) were acquired in 2001, 2007 and 2011. Navigation and positioning was based on stand-alone GPS, and the data was logged using an iXblue Delph Seismic system. In order to image the entire infill of the lake RCMG's multi-electrode Centipede sparker (400-1500 Hz) was used as seismic source and a single-channel high-resolution streamer as receiver. For a high-resolution characterization of the shallow subsurface, a Kongsberg GeoAcoustics Geopulse sub-bottom profiler (3.5 kHz, pinger) was deployed from a towed cataraft system. Seismic-stratigraphic interpretation was done using IHS Kingdom Suite after applying a frequency band-pass filter (more details in Moernaut et al., 2009). Water and subsurface depths for the uppermost seismic units were calculated using an acoustic velocity of 1500 m/s. As this is an underestimation for the lower seismic units typically encountered in glacialic

lakes (Finckh et al., 1984), we report the thicknesses for the general seismic stratigraphy in milliseconds Two-Way Travel Time (TWT).

3.2. Sediment cores

Several hammer-driven piston cores were taken in December 2011 from an anchored platform on the intact and failed slope (Fig. 3). The composite cores consist of two overlapping drives of successive sections of 3 m each. The closed cores were scanned with a Geotek multi-sensor core logger at ETH Zürich to obtain continuous along-core measurements of the density and magnetic susceptibility of the sediments. Detailed sedimentological and geotechnical parameters of these sediments (grain-size, macroscopic description, water content, undrained shear strength, Atterberg Limits, sensitivity, void ratio) were obtained at the Centre for Marine Environmental Sciences (MARUM), Bremen, and described in detail in Wiemer et al. (2015). In the present study, we build a conceptual model based on undrained shear strength derived from in-situ data (section 3.3), as this may better represent the actual field conditions compared to the data obtained during lab tests.

3.3. In-situ geotechnical measurements

The MARUM Free-Fall Piezocone Penetrometer (CPTU; Stegmann et al., 2006) was deployed in December 2011 at the coring sites, and along E-W and N-S oriented transects. A CPT lance of 5-8 m length was used from an anchored platform and a shorter lance (2-3 m) from a small research vessel. Cone resistance (q_c), sleeve friction (f_s) and pore pressure (u) were measured to maximum 8.1 m below lake bottom. Different penetration depths were reached by varying the free-fall height and thus initial penetration velocity. Undrained shear strength was derived from pore-pressure and strain-rate corrected cone resistance using the following geotechnical solution:

$$s_u = (q_{t, \text{quasi-stat}} - \sigma_{vo}) / N_{kt}$$

where s_u is the undrained shear strength, $q_{t, \text{quasi-stat}}$ is the pore pressure and strain-rate corrected cone resistance which is comparable to standard pushed CPT cone resistance, σ_{vo} is the total overburden stress, and N_{kt} is an empirically

determined cone factor (Lunne et al., 1997), here set to 14 as was done in Wiemer et al. (2015). The total overburden stress (σ_{vo}) was calculated as:

$$\sigma_{vo} = \gamma * z$$

where γ is the unit weight of the sediment and z is the subbottom depth. γ was taken as 13.9 kN/m³ based on the mean value of bulk density measurements. For details on MARUM CPTU data processing and post-processing, including strain-rate correction see Steiner (2013).

A dissipation test monitors the temporal evolution of the pore pressure signal after probe penetration –and its inherent displacement of sediment and fluids– towards ambient formation pressures (e.g. Seifert et al., 2008; Stegmann et al., 2011). Thirteen pore pressure dissipation tests were carried out. The dissipation tests had an average duration of ~40 min and ranged from 27 to 49 minutes (Table 1). The shape of dissipation curves is generally dependent on the hydraulic conductivity and the behavior of the sediments and pore fluids during and after probe insertion. Over-consolidated fine-grained soils –as in our study area (Wiemer et al., 2015)– can show dilatant shearing behavior and give rise to sub-hydrostatic pore water pressure recorded right behind the cone (u2 position) during penetration (Seifert et al., 2008). This produces “non-standard” dissipation curves (Fig. 5) in which the initial excess pore pressure values at the beginning of dissipation are smaller than in a standard dissipation curve, and sometimes even negative. These suppressed values can either rise to a peak and subsequently slowly decrease toward ambient conditions like in a standard curve, or slowly rise to a plateau without recording any peak in pore pressure (Sully et al., 1999; Stegmann et al., 2006; Seifert et al., 2008; Chai et al., 2012). In our study, the classification of Sully et al. (1999) was used to describe the different dissipation curves observed (Fig. 5).

At some sites, the hydrostatic pressure at the sediment-water interface presented slightly different values with consecutive CPT drops. This indicates a delayed response of measured pore pressure during free fall in the water column. For fast dynamic deployments, reduced pressure can result from the Bernoulli Effect, i.e. suction caused by the lake water moving relative to the CPT probe (termed “skin effect” by Seifert et al., 2008). Moreover, delayed pressure response can be induced by partial clogging of the pore pressure filter with fine sediments during successive drops. Therefore, as reference value for all drops performed at a specific site, we use the hydrostatic pressure at the lake bottom

measured during the first drop at that site. As no clogging is present then, this value forms the highest recorded lake bottom pressure at that specific site. For the dissipation tests, the reference hydrostatic pressure was obtained by adding the hydrostatic pressure for the respective penetration depth to the reference hydrostatic pressure at the lake bottom.

Different correction and extrapolation methods have been proposed to deduce the ambient pore water pressure from incomplete dissipation tests (e.g. Stegmann et al., 2006; Flemings et al., 2008; Sultan et al., 2009; Long et al., 2007; Sultan and Lafuerza, 2013). By comparing dissipation profiles with theoretical considerations, Sultan and Lafuerza (2013) suggest that the second derivative of pore pressure versus the logarithmic of time needs to be positive in order to make reliable extrapolations. In other words, the concavity of the curve at the end of dissipation needs to be up (concavity > 0). This is the case for most of our tests (see “concavity” column in Table 1), with exception of two cases where the end of curve is rather straight (concavity = 0 in Table 1). For incomplete dissipation profiles, the gradient of the ultimate part of the curve was extrapolated using the “inverse time” approach, which was tested and applied in different marine settings (Flemings et al., 2008; Sultan et al., 2009). We found that the “inverse time” approach is only applicable in case of a decreasing pore pressure at the end of the test (partial type I, II and IV profiles), and not for partial type V profiles (Fig. SI-2). As previously reported by Long et al. (2007), it seems that the $1/t$ method can systematically overestimate the ambient pore pressure value, whereas the $1/\sqrt{t}$ method systematically underestimates it (Fig. SI-2). The following procedure was applied in this study: The hydrostatic pressure (u_h) was subtracted from the “ambient” pore pressure to calculate the overpressure (u^*) at specific stratigraphic intervals. The overpressure ratio (λ^*) (Fig. 5; Flemings et al., 2008) represents the amount of excess pore pressure relative to the hydrostatic vertical effective stress (σ'_{vo}), which is calculated from the density data obtained on the sediment cores. For sites without core data, the average density values for the different units (U1-U3) were used and applied to a reconstructed lithological profile based on the seismic profiles (Table 1). The overpressure ratios are reported (λ^*) (Table 1) as either: i) a single value, in case of a complete dissipation (any type); ii) a minimum value, i.e. the last data point of a rising partial dissipation curve (type V); or iii) a range, in case of a decreasing partial dissipation curve (type I, III and IV). Minimum and maximum limits of the extrapolated pore pressure are based on the $1/\sqrt{t}$ and $1/t$ techniques, respectively.

4. Results

4.1. Sedimentary infill of Lake Villarica

Sparker seismic profiles reveal that the southwestern part of Lake Villarrica is characterized by a complex sedimentary infill (Fig. 6). The sedimentary infill consists of four major seismic units (SU-A to SU-D) deposited on top of the acoustic basement. Detailed identification of this basement is locally hampered by multiple reflections and limited acoustic energy reaching large sub-surface depths. The lowermost seismic unit SU-A comprises parallel sub-horizontal to slightly undulating reflections, which exhibit vertical acoustic wipe-outs and large lateral amplitude changes. It is the thickest unit (up to 140 ms) and has an infilling depositional geometry. Its upper boundary is an irregular surface which locally truncates the unit's reflections, and which is thus interpreted as an erosional unconformity. In the southwestern extremity of the lake, this unconformity and the acoustic basement is overlain by a massive unit SU-B which is characterized by a chaotic acoustic facies and an irregular eastwards-dipping upper boundary composed of several highs and depressions. SU-B has its maximum thickness in its western part (up to 90 ms). Both units SU-A and SU-B are overlain by SU-C, which has a similar acoustic facies as SU-A and partly levels out the pre-existing (i.e. top SU-B) morphological irregularities. SU-C is divided in two sub-units: i) SU-C1 which is only present in the basin and comprises reflections characterized by large lateral changes in amplitude, and ii) SU-C2, which drapes the entire area and shows less variability in reflection amplitudes. SU-C2 is covered concordantly by the continuous, parallel and high-amplitude reflections of SU-D. SU-D reflections are locally intercalated by lenses of chaotic-to-transparent facies interpreted as mass-transport deposits (Moernaut et al., 2009). In the shallower, southwestern part of the lake, units SU-C2 and SU-D were excavated at many places during slope failure events (Moernaut et al., 2009). In the basinal area, both SU-C and SU-D are each around 25-30 ms thick.

We interpret the origin of these seismic units by comparing them with seismic-stratigraphic studies of other glacial lakes worldwide (e.g. Van Rensbergen et al., 1998; 1999; Eyles et al., 2000; Heirman et al., 2011; Pinson et al., 2012; Ndiaye et al., 2014). The location, morphology, and seismic facies of SU-B resemble in all aspects the so-called lowermost "Unit 1" in these studies. It is interpreted as a complex and heterogeneous assemblage of glacier-edge deposits and landforms, such as subaquatic outwash fans, frontal dump moraines, glacier push moraines, etc. These glacial assemblages were rapidly deposited during the last glacial maximum and during the onset of deglaciation, and, in South-Central Chile, form a wide variety of sedimentary facies ranging from deformed glacio-lacustrine clays to gravels and boulders (Denton et al., 1999). Its westernmost location just offshore the frontal moraine belts (Fig. 2) confirms its intimate association with glacier-proximal processes during the initial phase of deglaciation.

The units SU-C and SU-D are interpreted as a typical succession of depositional units related to a retreating glacier in the lake's fluvial catchment (Heirman et al., 2011 and references therein). Unit SU-C represents glacio-lacustrine sediments deposited in a proglacial setting. When the glacier was still in close contact to the studied basin, coarse turbidites were deposited in the basin (SU-C1). Thereafter, the glacier front retreated out of the lake basin, into the Andean Cordillera. Meltwater-laden rivers provided large volumes of fine-grained glacially-eroded sediments (SU-C2) to the distal elevated parts of the proglacial lake. When temperatures culminated at the initiation of the Holocene, the catchment became essentially ice-free, and hemipelagic lacustrine sedimentation started draping the lake bottom and formed SU-D. The seismic-stratigraphic interpretation of the upper two units is confirmed by sediment core VILLAR1 (Fig. 4) taken on an unfailed portion of the slope. This core contains 7.3 m of diatomaceous ooze (U1 and U2) corresponding to the open lacustrine SU-D. The underlying detritic fine silt and clay (U3 in the core) correlates to unit SU-C2. Many sub-vertical fractures with small offsets are present in the rhythmically-layered sediments of U3. This strongly fractured pattern was also encountered in other cores in the lake and in outcrops of similar glacio-lacustrine sediments at the border of a glacigenic lake in the region (Lake Rupanco; Fig. 7). One AMS radiocarbon date (Fig. 4B) on a leaf, and two well-studied regional tephra layers observed in VILLAR1 (Wiemer et al., 2015; Fontijn et al., 2016) place the transition of unit SU-C2 (U3) to unit SU-D (U2) at around 12.7 ka BP. Tephrastatigraphy constrains a sedimentation rate of 0.4 to 0.8 mm/yr for the hemipelagic Holocene sediments (Fontijn et al., 2016). Deglaciation of the Chilean Lake District piedmont started around 17.5-17.1 ka BP at most places (McCullogh et al., 2000). Applying this date to the boundary of SU-B and SU-C2 (U3), we obtain a sedimentation rate of about 3 mm/yr for unit SU-C2 (U3) at VILLAR1, which is significantly higher than for SU-D (U1 and U2).

The main difference between the sedimentary infill of Lake Villarrica and the typical seismic stratigraphy of glacigenic lakes (Ndiaye et al., 2014 and references therein) is the presence of a thick unit SU-A below the glacier-proximal unit SU-B. Given its similar seismic facies as SU-C, it is regarded as a glacio-lacustrine unit deposited in a proglacial environment during former interstadials within the last glacial period. During the Last Glacial Maximum, SU-A was overridden by the glacier, scouring and pushing upper SU-A sediments towards the west, and dumping glacier debris in the southwestern part of the lake, in this way forming the glacier-proximal unit SU-B. Apparently, the readvancing glacier did not fully erode the pre-existing lake sediments, a phenomenon which has not been documented often, except in e.g. the Chilean Lake

Puyehue (see Heirman et al., 2011), the northern Swiss Alpine Foreland (Dehnert et al., 2012) or the Inn Valley in the Austrian Alps (Spötl et al., 2013).

4.2. Headwall scarps

The multibeam bathymetry data (Fig. 3, 8) provide the exact location and geometry of the headwall scarps of the large subaqueous landslides in the SW part of the lake. In the northern Area 1 (Fig. 8), the headwall scarp has several relatively straight segments in map view. Only in the shallowest parts, it locally cuts back upslope. In the southern Area 2, the headwall scarp forms a much more irregular trace. It cuts back and forth in the stratigraphic sequences and delimits an unfailed platform, which is almost completely surrounded by headwall scarps. Just north of the platform, a narrow bottle-neck shaped morphology is present and served as outlet for the remobilized and remolded landslide mass, which moved down from the western slopes (Fig. 3B; Wiemer et al., 2015). More eastwards, the headwall develops into a sidewall which has a very straight and smooth trace.

Due to the gravitational driving force for landsliding, headscarps are often located at downward steepening slope breaks in the stratigraphic layering, separating more stable and more unstable portions of the slope (Moernaut and De Batist, 2011). Figure 8B presents a paleobathymetric map of reflector R2, which forms the main sliding surface of the slope failures (“SFPT”; Wiemer et al., 2015). Interpolation between the 2D seismic profiles was used where R2 was disturbed or removed by deeper-seated slope failure (Fig. 8C-E). The headscarp only mimics the R2 contour lines in certain areas (e.g. in Area 1). In the western shallow region of Area 2, the headscarp trace does not follow the contour lines at all, and around the unfailed platform it runs even perpendicular to them, indicating that other parameters than slope gradient determine the headscarp location.

4.3. Seismic and morphologic evidence of focused fluid escape

Local reflection discontinuities and vertical wipe-out features without lateral offsets are present in the otherwise acoustically-stratified seismic facies. In some places, reflection continuity is disrupted for several tens of meters (e.g. “breached horizon” on Fig. 8C and 8E). By visual analogy to other studies in lacustrine basins (Chapron et al., 2004;

Moernaut et al., 2009) and marine basins (Cartwright and Santamarina, 2015), we interpret these features as fluid escape conduits indicating areas where focused fluid escape took and/or takes place.

In Area 1 (Fig. 9A), the bathymetric data show several morphological depressions with a concentric to elongated form. These are 10-80 m wide and up to 1-4 m deep and their bottom is overall horizontal but with many irregularities. Seismic profiles reveal fluid escape features in the form of acoustic wipe-outs and reflection discontinuity in the sedimentary sequences underlying these depressions. Based on this spatial association, we interpret the depressions as pockmarks. Their rather irregular form allows classifying them as “composite pockmarks” (cfr. Hovland et al., 2002).

In Area 2 (Fig. 9B), several distinct regions of fluid escape within the failed slope were interpreted from the seismic profiles. On a bathymetric slope gradient map, these regions stand out as overall horizontal zones of several tens of meters (green patches on Fig. 9C) located within a regionally eastward dipping lake bottom. These fluid escape areas show small-scale irregularities. Supported by the seismic data, we consider this specific morphological association (overall flat with small-scale irregularities) as an indicator of areas of fluid escape in sloping regions. This approach is further applied to locations which are not covered by seismic profiles. In the basin, this method does not work as this specific morphology may also represent mass-transport deposits. In the shallower westernmost part of Area 2, the sub-horizontal zones correspond to erosional unconformities on the seismic profiles. Interestingly, the mapped fluid escape areas are mainly situated above highs in the upper boundary of SU-B (Fig. 8C, 8D, 8E; Fig. 9D), the seismic unit composed of glacier-proximal deposits. Where the top of SU-B is flat, no fluid escape features were found. Locally, this leads to an apparent relief inversion (compare Fig. 9B with 9D).

4.4. In-situ undrained shear strength in the upper 2 m

The CPTU transects on and around the unfailed platform reveal significant spatial differences in undrained shear strength (s_u) of the upper 1-2 m of U1 sediments (Fig. 10). The undrained shear strength of the post-failure drape on top of the sliding plane is about 60-70% of that of the upper sediments on the unfailed platform. CPTU drops within the areas of fluid escape on the failed slope provide s_u values as low as 10-35% of those on the unfailed platform.

Undrained shear strength of normally consolidated sediment generally increases linearly with effective overburden stress. Underconsolidated sediments are characterized by lower undrained shear strength than expected from their burial depth. Therefore, the normalized undrained shear strength ratio (s_u / σ'_{vo}) is often used as a first-order in-situ indicator for the state of consolidation. Normally-consolidated organic and inorganic soils have a normalized undrained shear strength (s_u / σ'_{vo}) of 0.2-0.3 (Karlsson and Viberg, 1967; Skempton, 1969). Normally-consolidated diatomaceous sediments may reach s_u / σ'_{vo} as high as 0.4 (Tanaka et al., 2012) or even 0.7 (Diaz-Rodríguez, 2011). As outlined in Wiemer et al. (2015) we assume an intermediate gradient of $s_u / \sigma'_{vo} = 0.3-0.4$ for normally-consolidated U1-U2 diatomaceous ooze, which contains a small fraction of terrestrial organic matter, volcanic ash and terrigenous clays/silts. In this view, the relatively steep s_u trend (normalized s_u of ~ 1.7) on the unfailed platform indicates that these sediments are severely “apparently overconsolidated”. This is further supported by one-dimensional consolidation tests on core VILLAR1 which reveal an overconsolidation ratio (OCR) of 8.05 for the upper U1 (i.e. at 1.98 m subsurface) (Fig. 4). In contrast, similar sediment types deposited on the failed slope show less apparent overconsolidation (normalized s_u of ~ 1.2), whereas sediments deposited within the fluid escape areas seem to be normally consolidated or even underconsolidated (Fig. 10B).

4.5 In-situ undrained shear strength of the upper 8 m of unfailed platform sediments

CPTU measurements reaching ~ 8 m subsurface at three positions on the unfailed platform show a striking downward decrease in s_u for U2 (Fig. 4; Wiemer et al., 2015). Here we add s_u data from an additional CPTU test (CPT4) and a long sediment core (VILLAR8), both obtained in the vicinity of the main headwall scarp (Fig. 11). There, U1 and the upper part of U2 were evacuated (“sliding surface U2a” on Fig. 10) during the slope failure event around 1.6 ka BP. Since then, about 1.8 m of U1 sediment accumulated on top of the exposed sliding surface. The grain size distribution in the different units (U1-U3) is very similar to the grain size distribution obtained from core VILLAR1 (Fig. 4). Void ratio is highest in U2 and slightly increases with depth. Out of all study sites, U2 at site CPT4 exhibits the steepest downward decrease in s_u and the lowest s_u values at depth (down to 4 kPa). The close location of cores VILLAR1 and VILLAR8, the small difference in water depth and their quasi identical sedimentology suggest that the spatial variability in s_u is controlled by other factors than sediment type and depositional history.

4.6. Pore pressure based on dissipation tests

The 13 CPTU dissipation tests (Table 1) are grouped within the different types of pore pressure response documented by Sully et al. (1999) for overconsolidated fine-grained soils (Fig.5, Fig. 12). Type I represents the “standard” dissipation profile with an immediate peak due to insertion of the probe and a monotonic decay of pore pressure with time (Fig. 12D). This style was only found in the deep basin at the reference site CPT6, which is outside our main study area (location see Fig.3). The “unloading” type II was not encountered in our measurements as this style is characteristic for pore pressure filters installed on the cone surface (“u1 position”). Type III (3 tests) and type IV (3 tests) correspond to increasing pore pressures shortly after penetration and subsequent decay like a “standard” curve. The pore pressures immediately at the end of penetration are above (type III) or below the hydrostatic value (type IV). Type V response (6 tests) is characterized by initial negative excess pore pressures which monotonically rise to a “plateau” without recording any distinct peak in pore pressure.

Only two dissipation tests reached a stable value at the end of dissipation (Table 1). These tests were taken within sandy-gravelly tephra layers, which likely have a larger hydraulic conductivity. Shorter time for full dissipation can be expected compared to the less permeable silty background sediments of U2 or to the fine silty-clay of U3, where extrapolation of the dissipation test was needed. Most of the investigated stratigraphic levels at the different sites are characterized by formation overpressures ($\lambda^* > 0$). On the main study site CPT1.2, the measurement at 7 m subsurface ($\lambda^* = 0.42$) is by far the most reliable, as it is obtained from a complete dissipation. The 5.5 m level shows the highest measured λ^* value (0.78) and pore pressure was still rising at the end of the 40 minutes long test.

For investigating the spatial distribution of excess pore pressure, we compared the overpressure (λ^*) values obtained in the lower part of the stratigraphy (lower U2 and within U3) at different sites (Fig. 13). On the unfailed platform, only sites CPT1.1 and CPT1.2 show λ^* values below 0.5. An intermediate value of 0.53-0.8 is obtained at CPT 1.3, located on the unfailed platform close to the headwall scarp. Sediments underlying the failed slope sequence (CPT2, 3, 4) and a fluid escape area (CPT5) are characterized by the highest overpressure values ($\lambda^* = 0.8-0.9$). At reference site CPT6 in the basin, almost no overpressure was found ($\lambda^* = \sim 0.04-0.05$).

5. Discussion

5.1. Measurements of overpressure

Estimations of fluid overpressure are often based on measurements of porosity and one-dimensional consolidation tests on core samples (Dugan and Sheahan, 2012). In the present study, the downcore increasing void ratio of U2 in VILLAR8 (Fig. 11B) deviates strongly from what would be expected by consolidation under hydrostatic conditions, i.e. a rather logarithmic decrease of void ratio with depth. It thus seems that the normal consolidation process is hampered, possibly due to the presence of excess pore pressure in this unit. Quantification of overpressure using void ratio is not feasible due to the dominance of diatoms in U2. Diatom frustules retain large portions of water in their structure (intra-skeleton porosity) and, therefore, a change in diatom abundance (or type) may produce variations in the measured void ratio, regardless of possible changes in the degree of consolidation. In such sediments, CPTU dissipation tests may produce more reliable estimates of overpressure. The most important limitation is the relatively short duration of the tests compared to the time needed for complete dissipation of the probe insertion pressure, especially in fine-grained sediment of low hydraulic conductivity. Moreover, the possible dilative behavior upon CPT penetration in overconsolidated sediments (e.g. U1 and U2) can lead to non-standard dissipation curves that often require longer measurements for making meaningful extrapolations. In case of Type III and IV profiles, dissipation time must have –at least– passed the peak in measured pore pressure, whereas extrapolation with type V profiles is always erroneous. Another source of error on the calculation of λ^* is the estimation of vertical overburden stress based on γ -density measurements on piston cores. Piston coring may induce sediment disturbance and gas expansion could result in incompletely filled core liners. In this study we consider this error as minimal, as sediment cores were of excellent quality indicated by intact horizontal laminations, absence of sediment gaps and a good core-to-seismic correlation. In conclusion, we acknowledge that it is difficult to constrain the error bars on the absolute overpressure values based on these tests and calculations, but as the data is considered to be of good quality (see also Wiemer et al., 2015), we are confident that it can be used to explore the general trends in excess pore pressures between sites and between stratigraphic levels.

5.2. Shear strength and consolidation state at the unfailed platform and pockmarks

The sediments of U1 on the unfailed platform and failed slope are “apparently overconsolidated” (Fig. 10; Wiemer et al., 2015). Progressive “seismic strengthening” (Locat and Lee, 2002; Sawyer and DeVore, 2015) due to recurrent earthquake shaking and intermittent excess pore pressure dissipation may have taken place in this seismically very active region.

However, seismic strengthening does not explain why the sediments covering the failed slope show less apparent overconsolidation (normalized s_u of ~ 1.2) compared to the unfailed plateau (normalized s_u of ~ 1.7), as both areas experienced the same seismic history. The difference in the degree of overconsolidation seems to be related to water depth; i.e. the unfailed platform is located at a water depth of 19-25 m whereas the failed slope region at 30-40 m. We hypothesize that sediments on the unfailed platform could have been more affected by wave action processes during strong easterly storms (“Puelche”; Meruane, 2005). Indeed, multiple episodes of wave-induced cyclic loading can densify sediments if the generated excess pore pressures can dissipate between events (Sassa and Sekiguchi, 1999). We therefore hypothesize that wave action causes the development of severe apparent overconsolidation in the relatively shallow water sediments in Lake Villarrica. In contrast, sediments within the fluid escape areas are normally consolidated (Fig. 10). It thus seems that disturbance of the sediment structure (i.e. remolding) by focused fluid escape resets the initially overconsolidated sediments to a normal consolidation state. This is expressed in a drastic loss of in-situ s_u by ~ 40 -80% compared to the adjacent areas 50-100 m away on the failed slope (blue vs. green data on Fig. 10).

5.3 Influence of focused fluid flow on headwall scarp geometry

For spatially uniform sediment sequences in lacustrine and marine settings, headwall scarps mostly develop along downslope-steepening slope breaks in the stratigraphic horizons (e.g. Strozyk et al., 2010; Moernaut and De Batist, 2011). Lower downslope gravitational stress acting upslope of the slope break, results in more stable sediment sequences. In Lake Villarrica, headwall scarps are associated to slope breaks (Fig. 8B) where the underlying sequences are devoid of fluid escape features. In map view, these scarps are relatively straight and smooth, as their position is mainly controlled by the geometry of the stratigraphic sequences (i.e. R2). Locally, irregular segments of the headwall scarp crosscut the stratigraphic contour lines. This occurs near areas of focused fluid escape where the underlying sediments show wipe-outs and breached horizons on the seismic profiles (Fig. 8, 9). This spatial association between fluid escape and irregular headwall scarps suggests that focused fluid escape locally weakened the sediments (see section 5.2), making such areas more prone to failure (cfr. tearing apart a pre-perforated piece of paper).

5.4. Overpressure and hydrofracture development

Fractures were found within the youngest glacio-lacustrine unit (U3), deposited when the glacier already retreated far back into the catchment. Hence, these fractures are not linked to stresses and meltwater pumping associated to glaciers overriding former lacustrine sequences (e.g. Philipps et al., 2007). Our data does not allow revealing the exact geometry and genesis of the U3 fractures (Fig. 7) but the documented elevated pore pressures point towards hydraulic fracturing as possible mechanism. Hydraulic fracturing can take place when pore pressure in the sediments exceeds the least principal stress (i.e. horizontal) and the tensile strength of the sediments. Effective stress drops significantly and mode I fractures can form under tensile stresses applied on lake sediments that are stiff enough to undergo brittle failure. Fault displacement in U3 with vertical throws up to several centimeters to a decimeter suggest that processes other than hydraulic fracturing were also involved in the deformation process, such as e.g. the downslope gravitational stress acting on the 1-4° dipping slopes.

5.5. Influence of overpressure on slope stability

Fluid overpressure reduces the effective overburden stress available for the consolidation of sediment, and thus, leads to underconsolidated sediment sequences. To investigate if fluid overpressure determined the consolidation state of the Lake Villarrica sediments, we evaluated the overpressure ratio (λ^*) in light of the normalized undrained shear strength ratio (s_u / σ'_{vo}) (Fig. 14). Normal consolidation of the diatomaceous ooze is associated with a s_u / σ'_{vo} of 0.3-0.4 (section 4.4). Large scatter is expected as different stratigraphic levels are included with -in some cases- strongly contrasting lithology (Fig. 13), and thus s_u values. Due to the low amount of data points ($n = 4$), we opted to add data from below the failed slope. To do so, we reconstructed the maximum overburden thickness that these sediments experienced, i.e. at the moment of slope failure. This was done based on stratigraphic reconstruction using seismic profiles and average density values for U1 and U2. Fig. 14 shows that stratigraphic intervals with λ^* of 0.5-0.9 are characterized by underconsolidated sediment, whereas λ^* of about 0.4 associates with normally consolidated sediments. The lowest λ^* of ~0.15 corresponds to overconsolidated sediments situated higher up in the stratigraphy (Fig. 13D). In general, there is a rough negative relationship between s_u / σ'_{vo} and λ^* which suggests that normal consolidation is inhibited at locations where significant overpressure is present. Accordingly, the strongly decreasing s_u with depth at sites CPT1.3 and CPT4 (Fig. 13E) may be related to the high λ^* (0.7-0.9) at depth. Overpressure must have been present there since the early stage of sediment burial in order to keep consolidation and s_u at very low values. In such case, water content of the U3 sediments remained high (80% “dry” water content), leading to very low remolded shear strength (~0.02 kPa; Wiemer et al., 2015). This partly

clarifies the high sensitivity and quick clay behavior of U3, in addition to its high-void-ratio flocculated fabric. We do not have dissipation tests in the upper 4 m of sediments, but according to our findings on the deeper sequences, we interpret that the very high s_u / σ'_{vo} values (0.5-2.5; Wiemer et al., 2015) indicate that no significant overpressures exist in the upper U2 and U1 sediments (< 4 mbf).

To evaluate the effect of earthquake shaking on slope stability, Wiemer et al. (2015) calculated a pseudo-static factor of safety (FoS) for sites CPT1.1, 1.2 and 1.3, based on the in-situ s_u values and the infinite-slope approach. For a seismic intensity of VII, a FoS < 1 in U2 is only reached at site CPT 1.3, just above the SFPT sliding surface. The FoS at that stratigraphic level is about two times higher at site CPT1.1 and CPT1.2 (FoS = 2). Based on the new pore pressure data presented here, it is hypothesized that the lower λ^* values (< 0.5) in CPT1.1 and CPT1.2 resulted in a less-reduced consolidation state and thus in a more stable sequence. This explains why this area did not fail during the earthquake-triggered landslide, which excavated the areas north, east and south of this unfailed platform. In this regard, site CPT1.3 must have been close to failure during the earthquake shaking, as indicated by the low FoS and its location at only 40 m of the generated headwall scarp.

The influence of overpressure on slope stability is not limited to long-term preconditioning of a slope sequence by obstructing the normal consolidation process and thus the development of an underconsolidated state. For example, after the complete (normal) consolidation of a sequence, a sudden increase of overpressure (e.g. by an earthquake) can also affect slope stability. Static failure in granular material on a gentle slope ($< 2^\circ$) can occur when λ^* is very high, i.e. around 0.8-0.94 depending on the sediment critical state strength (Urlaub et al., 2015). The effective stresses are lowered and the Mohr Coulomb effective stress failure criterion can be considered using the drained shear strength as resisting force to the static shear stress induced by gravity. The s_u of sediments does not react to post-consolidation pore pressure changes (Leynaud et al., 2009), except when pore pressure is high enough (λ^* near unity) to induce liquefaction of susceptible cohesionless sediments (see section 5.6.1).

5.6. Evaluation of the origin of overpressure

Many processes can lead to excess pore pressure in sedimentary sequences (Dugan and Sheahan, 2012). Here we evaluate the possibility of several overpressure mechanisms in order to single out the most likely one causing overpressure in the slope sequence of Lake Villarrica.

5.6.1. Earthquake shaking

Strong earthquake shaking can induce significant overpressures in shallow sediments, which can take days to years to dissipate (Biscontin et al., 2004). For example, earthquake-induced overpressures have been documented in the submarine slope sediments near the Sunda Trench (offshore Sumatra) about seven months after the M_w 9.1 Sumatra-Andaman in 2004 (Sultan et al., 2009). The survey on Lake Villarrica was conducted about 2 years (22 months) after the 27 February 2010 earthquake (M_w : 8.8) in Central Chile, which had a seismic intensity of VI¼ at Lake Villarrica (Moernaut et al., 2014) and produced several liquefaction-related phenomena (sand boils, lateral spreading, etc.) at the lake shore. The question arises whether the overpressures encountered in our in-situ measurements are possible remnants of excess pore pressures generated during the 2010 Maule earthquake. Wiemer et al. (2015) used dynamic triaxial tests to decipher the response of the different lithological units in the Lake Villarrica slope sequences to cyclic loading under an overall compressive regime. In this way, strain and pore pressure build-up during earthquake shaking was simulated for estimated in-situ stress conditions. These tests show that the diatomaceous ooze (U1, U2) and tephra layers at our study sites generate a significant amount of overpressure (30-60% of the confining pressure at strain failure; Wiemer et al., 2015) when subjected to cyclic loading simulating an earthquake with a peak ground acceleration of 0.25 g or seismic intensity of ~VIII. Pore pressure buildup in the quick clays of U3 was much slower (about 8% of confining pressure at strain failure). Consequently, it seems highly unlikely that earthquake shaking induced the high overpressure ratios (λ^* : 0.7-0.9) found in the U3 sediments. Also, it would be hard to explain the differences in λ^* in only short distances (~100-200 m) as earthquake shaking impacted these sites in a similar manner.

The boundary between the tephra layer “SFPT” and the overlying U2 sediments acted as one of the main sliding planes for the studied slope failures. Wiemer et al. (2015) hypothesized that this developed due to an earthquake-induced pore pressure pulse in the SFPT and –due to its high permeability– a rapid transfer of pore pressure to the overlying U2 sediments. This sudden pore pressure rise may have significantly weakened the U2 sediments at this interface, allowing large-scale sliding on a discrete failure plane. Remarkably, the highest overpressure ratios ($\lambda^* = 0.78+$) at CPT 1.2 were

found just 0.5 m above the SFPT, whereas other stratigraphic levels at this site have considerable lower overpressure values ($\lambda^* = 0.13-0.42$) (Fig. 13A; Table 1). Therefore, we hypothesize that this elevated in-situ pore pressure above the SFPT corroborates the hypothesis of Wiemer et al. (2015), as this indeed may be a long-lived relict of earthquake-induced overpressure in the SFPT tephra layer and postseismic fluid migration to the overlying U2 sediments.

5.6.2. Rapid sedimentation

Disequilibrium compaction due to rapid sedimentation (>1 mm/yr) of compressible, low permeability ($< 10^{-16}$ m²) material is considered as one of the main causes of shallow overpressure (Dugan and Germaine, 2008; Urlaub et al., 2015). The low sedimentation rates (0.4-0.8 mm/yr) in U1-U2 at our study site in Lake Villarrica make syn-depositional overpressure generation unlikely. In U3, sedimentation rates (~ 3 mm/yr) may be high enough and permeability ($\sim 10^{-17}$ m²; derived from hydraulic conductivity) may be low enough to produce syn-depositional overpressure. However, we do not expect large spatial variability in the type and rate of glacio-lacustrine U3 deposition in the SW distal area of the proglacial lake, where fine material slowly settled from meltwater plumes. In such setting, overpressure by disequilibrium compaction should be distributed rather equally over the considered slope segment. This contradicts the existence of distinct fluid escape areas rooted below U3 and the strong spatial variation in the documented overpressure ratios. We therefore infer that syn-depositional overpressure may have contributed to some extent to overpressure generation in U3, but is not the main mechanism.

5.6.3. Unloading by slope failure

Many studies postulate that pockmark fields within and near a landslide scar were created shortly after slope failure took place (e.g. Reiche et al., 2011; Bøe et al., 2012). Ebullition of dissolved gases in the pore spaces occurs due to a sudden drop in overburden pressure when unloading the sediments. These gas bubbles can vent to the sea bottom and abruptly create pockmark fields. Remobilization of slope sediments can also induce steepened pore-pressure gradients adjacent to scarps which can induce fluid flow towards these scarps (You et al., 2012). However, in Lake Villarrica, the occurrence of isolated pockmarks (Fig. 9A) far away from the landslide scarps implies that unloading via slope failure is not the dominant mechanism for focused fluid escape.

5.6.4. Free gas

Elevated pore pressures and non-standard dissipation profiles have often been attributed to the presence of free gas in offshore sediments (Hirst and Richards, 1977; Stegmann et al., 2006). No evidence of degassing (cracks and holes) was observed in the Lake Villarrica sediment cores. In typical glacio-lacustrine sediments such as in U3, this would also not be expected as a very low organic matter contents simply impedes methanogenesis. Seismic profiles do not show evidence of free gas in the form of acoustic turbidity or enhanced reflections in the studied unfailed platform. Near-surface acoustic turbidity was detected in some pockmarks of the westernmost shallow part of the lake, and may indicate free gas in the sediments (SI-Fig. 3). Given the coastal setting of these pockmarks, we think that shallow biogenic gas may have been produced by trapping of coastal organic material due to locally reduced bottom currents in the overall dynamic littoral system (cfr. Bussmann et al., 2011). With exception of these westernmost depressions, we conclude that free gas is not present in our study area and does not account for the elevated pore pressures measured.

5.6.5. Fluid transfer from below

None of the previously mentioned overpressure generation processes can explain the spatial variability in overpressure ratio and the existence of discrete areas of focused fluid escape. The association of focused fluid escape with the topographic highs in the underlying SU-B (Fig. 8, 9D) suggests that fluids are sourced from below the studied U1-U3 sequence. SU-B is considered as the reservoir unit of the expelled fluids at the lake floor, whereas the fine-grained SU-C2 (U3) represents a permeability barrier ("cap layer"). Indeed, the latter unit contains an elevated clay content of about 35% (Fig. 4), which results in a relatively low permeability ($\sim 10^{-17} \text{ m}^2$). SU-C2 is thinner on top of the topographic highs in SU-B (Fig. 8C-E). This leads to a locally reduced vertical overburden stress and a steeper pore pressure gradient, which allows the development of fluid focusing towards the highs in SU-B (i.e. topographic focusing). Additionally, horizontal compressive stresses ("lateral earth pressure") are lowest above these highs due to the radial pattern of the downslope component of gravitational stress. Consequently, the combination of enhanced overpressures due to flow focusing and enhanced tensile stress above the SU-B topographic highs make these the preferential location for hydrofracturing to take place.

Unit SU-B probably consists of a heterogeneous assemblage of rapidly-deposited mixture of coarse sediments deposited at the glacier edge and remobilized fine-grained lake sediments (section 4.1, Fig. 15). Ice-proximal sediments (till) can exhibit a wide range of permeabilities (10^{-19} to 10^{-13} m^2 ; Domenico and Schwartz, 1990). Low permeability results from the inclusion of compacted fine grained glacio-lacustrine sediments bulldozed by the glacier, and sediment overconsolidation due to glacial loading. More permeable fluid flow pathways can be formed by meltwater pumping, locally impeding the consolidation process (Boulton and Dobbie, 1993) and creating permeable hydrofracture networks filled with coarse sediments (Phillips et al., 2007). In Lake Villarrica, the permeability of SU-B must be lower than of SU-C2 in order to act as a reservoir unit for overpressured fluids and for topographic focusing to take place. This assumption is feasible given that the measured permeability of the cap layer SU-C2 (10^{-17} m^2) falls in the lower part of the possible range for glacial till.

We hypothesize that severe overpressures in SU-B may have been generated during its deposition by i) glacial loading of the adjacent SU-A and associated injection of fluids into SU-B (Fig. 15B; e.g. Siegel et al., 2014), and/or ii) rapid dumping of SU-B ($\sim 0.1\text{-}1 \text{ m/yr?}$) near the glacier front without full drainage of pore water during progressive burial. Both mechanisms lead to a strongly underconsolidated state, especially in the lower part of SU-B. Once the glacier retreats from this position, both drivers for overpressure abruptly cease, and normal consolidation of SU-B initiates. This provokes the upwards expulsion of a large amount of excess pore water. The SU-C2 permeability barrier modulates fluid pathways towards topographic highs, leading to locally high overpressure ratios. These induce hydrofracturing of SU-C2 and partial release of overpressured fluids, followed by a closing of the fractures due to increased effective overburden stress. When overpressure builds up again by continuous fluid migration from below, the hydrofractures reopen and the escaping fluids reuse these pathways. In this manner, fluid venting at the lake floor reoccurs in the same areas, partially inhibiting sedimentation and creating long-lived pockmark features.

A main uncertainty in this conceptual model is that the normal consolidation of SU-B and upwards migration of excess pore water needs to be relatively slow in order to supply fluids to the root zones of the fluid escape areas during the Holocene. If not, dissipation would result in low remaining overpressures nowadays (e.g. Hustoft et al., 2009), which disagrees with the high λ^* detected in U3 and lower U2, and the Holocene fluid expulsion features on the seismic profiles. For resolving the exact role of SU-B in the generation and transfer of overpressure, one needs to fully characterize the depositional architecture of SU-B by a large amount of deep sediment cores, geotechnical tests and (pseudo-) 3D seismic

data. Such field data should be integrated with numerical simulations (Vanneste et al., 2014) in order to thoroughly evaluating different hypotheses.

5.7. General implications

Lake Villarrica's depositional sequence is comparable to many glacial lake basins worldwide in which a fine-grained glacio-lacustrine cover overlies coarser and more heterogeneous glacier-proximal deposits (see section 4.1). Fluid overpressure in "glacial units" is hypothesized to play a key role in the initiation of landslides in Alpine lakes (Strasser et al., 2007; Stegmann et al., 2007), but a comprehensive understanding is lacking as no other lacustrine studies were dedicated on this topic. Several fjords and high-latitude ocean margins may contain a similar succession of depositional units and therefore we expect that overpressured sequences may exist where massive rapidly-deposited glacier-proximal units are covered by glacio-marine clays. This concept has been proposed to control the creation of large subaquatic landslides in fjords (e.g. Hjelstuen et al., 2009) and at formerly-glaciated passive margins (Leynaud et al., 2009 and references therein), although ground-truthing by in-situ pore pressure measurements is generally missing. In Southern Alaskan fjords, it is hypothesized that thick, underconsolidated slope sequences have been produced by enhanced sediment input during the Little Ice Age. These may have facilitated the development of massive tsunamigenic failures at several fjord head deltas (Haeussler et al., 2014), moraines and sub-basin slopes (Brothers et al., 2016) during the giant AD 1964 (M_w 9.2) megathrust earthquake. For the Storegga Slide offshore Norway, the alternation of rapidly-deposited glacial sediments (tills, debris flows) with fine-grained contouritic marine deposits may have created high excess pore pressure ratios facilitating the development of this giant landslide on a very gentle slope (Kvalstad et al., 2005). Accordingly, alternating rapid and slow sedimentation during glacial-interglacial cycles may have been reflected in the recurrence of large slides at the Norwegian margin (Solheim et al., 2005). However, not all submarine landslides at glacial continental margins are governed by this type of preconditioning, i.e. slides occurring in a low-sedimentation-rate regime underline the influence of other processes such as contour currents, tectonic faulting, and overpressure build-up related to gas hydrate destabilization (Elger et al., 2015).

Due to the many technical challenges involved in the acquisition of in-situ pore pressure data, we propose that the identification of fluid escape features and massive glacial-proximal (type SU-B) units on seismic profiles may be used as a

first predictive tool to identify regions where overpressure could be present. These areas can then serve as dedicated targets for more detailed in-situ measurements in order to better characterize submarine landslide potential and risks.

6. Conclusions

We characterized the morphology, seismic stratigraphy and geotechnical properties of a subaquatic slope in the glacial Lake Villarrica (South-Central Chile) where large translational landslides took place. We identified areas of focused fluid escape and detected spatial variability in in-situ undrained shear strength and ambient overpressure. This study came up with the following inferences and conclusions concerning the slope sequences in Lake Villarrica, which may be applicable at other formerly-glaciated basins and continental margins:

1. Focused fluid escape may locally lower the sediment consolidation state (by remolding) and thus weaken the sediments. Cone penetrometer tests can be used as a rapid tool to identify areas of fluid escape based on their reduced normalized undrained shear strength. This is especially useful where seismic-reflection or morphologic expression of fluid escape processes is unclear.
2. Reduced sediment strength due to fluid escape facilitates retrogression of landslide headwall scarps. This can lead to a highly irregular headwall trace and an increased landslide volume compared to areas where focused fluid escape does not occur.
3. Overpressure below glacio-lacustrine sediments may have created networks of hydrofractures through which focused fluid escape repeatedly took place.
4. The sedimentary sequences near the slide scarp and near fluid escape areas exhibit high overpressure ratios ($\lambda^* = 0.7-0.9$). Significantly lower λ^* values of 0.3-0.4 are present in an unfailed platform where no fluid escape features were found. This suggests that fluid escape features are a direct result of overpressure gradients in the sequence, and thus can be used as a predictive tool.

5. High overpressure ratio near the slide scarp corresponds with low normalized undrained shear strength, suggesting that overpressures were present since early burial. This inhibited normal consolidation which resulted in underconsolidated sequences with high water content and the development of a very high sensitivity to remoulding. Lower overpressure ratios and normally to overconsolidated sediments in the other parts of the unfailed platform explain why this slope segment remained stable during past earthquake-triggered landslide events. Given a simple pseudo-static slope stability model, it seems that a seismic intensity of VIII is required for initiating slope instability there ($FoS < 1$).

6. The influence of overpressure on slope stability in Lake Villarrica is three-fold: i) overpressure gradients drive focused fluid escape, locally remolding and weakening the sediments, ii) persistent overpressure during burial generates underconsolidated sequences with low undrained shear strength, and thus, a relatively lower Factor of Safety during earthquake shaking compared to normally-consolidated sediments, iii) very high overpressure ratios reduce effective stress and facilitate static failure.

7. Seismic shaking, high sedimentation rate and unloading due to slope failure are unlikely mechanisms to explain the variability in overpressure ratio and formation of discrete fluid escape areas. Moreover, no evidence of free gas was found near the failed slope and thus fluid overpressure must have a different origin.

8. Spatial correlation of focused fluid escape and topographic highs of a massive glacier-proximal unit suggests that the latter acted as a reservoir for overpressured fluids. The glacio-lacustrine drape is characterized by a low permeability and functioned as a permeability cap layer, forcing overpressure migration to these topographic highs. Rapid deposition at the glacier front and subsurface meltwater pumping by glacial loading are two possible mechanism for generating overpressure and underconsolidation in the glacier-proximal unit. Subsequent normal consolidation and thus upwards expulsion of excess pore water during postglacial times may form the driver for the actual overpressures in the more shallow sediments.

9. This conceptual model for overpressure development, release and modulation in the sedimentary infill of a glacigenic lake may be applied to other formerly-glaciated lacustrine and marine environments as these may have undergone a very similar depositional history. The composition and geometry of the glacier-proximal unit determines where the largest overpressure will accumulate in younger sediments and thus where large landslides and pockmarks can develop. This

intimate relation between deglaciation depositional history and landslide development suggests that detailed seismic-stratigraphic mapping of glacier-proximal units and fluid escape features may be used as a first indicator of possible overpressured sequences and subaquatic landslide potential.

Acknowledgments:

We thank the Chilean survey company BENTOS for the cost-effective logistical and technical support of the multibeam survey on Lake Villarrica. We further acknowledge the support of Francisco Martin, Mario Pino, Koen De Rycker, Alejandro Peña, Manuel Novoa, Philipp Kempf, François Charlet and Robert Brümmer during the geophysical and geotechnical field campaigns, and of Michael Hilbe for multibeam processing. IHS Kingdom and OmniSTAR are acknowledged for their educational grant programs providing seismic interpretation software and high-accuracy satellite positioning data, respectively. This research is funded by the Swiss National Science Foundation (grant 133481), a research grant of the Research Foundation Flanders (FWO-Vlaanderen) and the MARUM - Center for Marine Environmental Sciences via the Deutsche Forschungsgemeinschaft (Bonn, Germany). Jasper Moernaut further acknowledges research funding via the Chilean Fondecyt Project nr. 1150346. We gratefully thank Daniel Brothers and an anonymous reviewer for their constructive comments on an earlier version of this manuscript.

References:

- Bartetzko, A., Kopf, A., 2007. The relationship of undrained shear strength and porosity with depth in shallow (< 50 m) marine sediments. *Sedimentary Geology* 196, 235–249.
- Berndt, C., Costa, S., Canals, M., Camerlenghi, A., De Mol, B., Saunders, M., 2012. Repeated slope failure linked to fluid migration: The Ana submarine landslide complex, Eivissa Channel, Western Mediterranean Sea. *Earth and Planetary Science Letters* 319–320, 65–74.
- Biscontin, G., Pestana, J.M., Nadim, F., 2004. Seismic triggering of submarine slides in soft cohesive soil deposits. *Marine Geology* 203, 341–354.
- Bøe, R., Bellec, V.K., Rise, L., Buhl-Mortensen, L., Chand, S., Thorsnes, T., 2012. Catastrophic fluid escape venting-tunnels and related features associated with large submarine slides on the continental rise off Vesterålen–Troms, North Norway. *Marine and Petroleum Geology* 38, 95–103.
- Boulton, G.S., Dobbie, K.E., 1993. Consolidation of sediments by glaciers: relations between sediment geotechnics, soft-bed glacier dynamics and subglacial ground-water flow. *Journal of Glaciology* 39(131), 26–44.
- Brothers, D.S., Ruppel, C., Kluesner, J.W., ten Brink, U.S., Chaytor, J.D., Hill, J.C., Andrews, B.D., Flores, C., 2014. Seabed fluid expulsion along the upper slope and outer shelf of the U.S. Atlantic continental margin. *Geophysical Research Letters* 41, 96–101. doi:10.1002/2013GL058048
- Brothers, D.S., Haeussler, P.J., Liberty, L., Finlayson, D., Geist, E., Labay, K., Byerly, M., 2016. A submarine landslide source for the devastating 1964 Chenega tsunami, southern Alaska. *Earth and Planetary Science Letters* 438, 112–121.
- Bryn, P., Berg, K., Forsberg, C.F., Solheim, A., Kvalstad, T.J., 2005. Explaining the Storegga Slide. *Marine and Petroleum Geology* 22, 11–19.

- Bünz, S., Mienert, J., Bryn, P., Børg, K., 2005. Fluid flow impact on slope failure from 3D seismic data: a case study in the Storegga Slide. *Basin Research* 17, 109–122.
- Bussmann, I., Schlömer, S., Schlüter, M., Wessels, M., 2011. Active pockmarks in a large lake (Lake Constance, Germany): Effects on methane distribution and turnover in the sediment. *Limnology and Oceanography* 56(1), 379–393.
doi:10.4319/lo.2011.56.1.0379
- Cartwright, J., Santamarina, C., 2015. Seismic characteristics of fluid escape pipes in sedimentary basins: implications for pipe genesis. *Marine and Petroleum Geology* 65, 126–140. doi: 10.1016/j.marpetgeo.2015.03.023.
- Chai, J., Sheng, D., Carter, J.P., Zhu, H., 2012. Coefficient of consolidation from non-standard piezocone dissipation curves. *Computers and Geotechnics* 41, 13–22.
- Chapron, E., Van Rensbergen, P., De Batist, M., Beck, C., Henriot, J.P., 2004. Fluid-escape features as a precursor of a large sublacustrine sediment slide in Lake Le Bourget, NW Alps, France. *Terra Nova* 16, 305–311.
- Dehnert, A., Lowick, S.E., Preusser, F., Anselmetti, F.S., Drescher-Schneider, R., Graf, H.R., Heller, F., Horstmeyer, H., Kemna, H.A., Nowaczyk, N.R., Züger, A., Furrer, H., 2012. Evolution of an overdeepened trough in the northern Alpine Foreland at Niederweningen, Switzerland. *Quaternary Science Reviews* 34, 127–145.
- Denton, G.H., Lowell, T.V., Heusser, C.J., Schlüchter, C., Andersen, B.G., Heusser, L.E., Moreno, P.I., Marchant, D.R., 1999. Geomorphology, stratigraphy, and radiocarbon chronology of Llanquihue drift in the area of the southern Lake District, Seno Reloncavi, and Isla Grande de Chiloé, Chile. *Geografiska Annaler* 81A, 167–229.
- Domenico, P.A., Schwartz, F.W. , 1990. *Physical and Chemical Hydrogeology*, John Wiley & Sons, New York, 824 p.
- Diaz-Rodríguez, J.A., 2011. Diatomaceous soils: monotonic behavior. In: *International Symposium on Deformation Characteristics of Geomaterials*, Seoul, Korea.

Dugan, B., Flemings, P.B., 2000. Overpressure and fluid flow in the New Jersey Continental Slope: implications for slope failure and cold seeps. *Science* 289, 288–291.

Dugan, B., Germaine, J.T., 2008. Near-seafloor overpressure in the deepwater Mississippi Canyon, northern Gulf of Mexico. *Geophysical Research Letters* 35, L02304. doi:10.1029/2007GL032275

Dugan, B., Sheahan, T.C., 2012. Offshore sediment overpressures of passive margins: mechanisms, measurement, and models. *Reviews of Geophysics* 50, RG3001.

Elger, J., Berndt, C., Krastel, S., Piper, D.J.W., Gross, F., Spielhagen, R.F., Meyer, S., 2015. The Fram Slide off Svalbard: a submarine landslide on a low sedimentation-rate glacial continental margin. *Journal of the Geological Society* 172(2), 153–156). doi:10.1144/jgs2014-055

Eyles, N., Boyce, J.I., Halfman, J.D., Koseoglu, 2000. Seismic stratigraphy of Lake Waterton, a sediment-starved glaciated basin in the Rocky Mountains of Alberta, Canada and Montana, USA. *Sedimentary Geology* 130, 283–311.

Field, M.E., Gardner, J.V., Jennings, A.E., Edwards, B.D., 1982. Earthquake-induced sediment failures on a 0.25° slope, Klamath River delta, California. *Geology* 10, 542–546.

Finckh, P., Kelts, K., Lambert, A., 1984. Seismic stratigraphy and bedrock forms in perialpine lakes. *Geological Society of America Bulletin* 95, 1118–1128.

Flemings, P.B., Long, H., Dugan, B., Germaine, J., John, C.M., Behrmann, J.H., Sawyer, D., IODP Expedition 308 Scientists, 2008. Pore pressure penetrometers document high overpressure near the seafloor where multiple submarine landslides have occurred on the continental slope, offshore Louisiana, Gulf of Mexico. *Earth and Planetary Science Letters* 269, 309–325.

Fontijn, K., Rawson, H., Van Daele, M., Moernaut, J., Abarzúa, A.M., Heirman, K., Bertrand, S., Pyle, D.M., Mather, T.A., De Batist, M., Naranjo, J.-A., Moreno, H., 2016. Synchronisation of sedimentary records using tephra: A postglacial

tephrochronological model for the Chilean Lake District. *Quaternary Science Reviews* 137, 234–254. doi: 10.1016/j.quascirev.2016.02.015

Gee, M.J.R., Uy, H.S., Warren, J., Morley, C.K., Lambiase, J.J., 2007. The Brunei slide: a giant submarine landslide on the North West Borneo Margin revealed by 3D seismic data. *Marine Geology* 246(1), 9–23.

Haeussler, P.J., Parsons, T., Finlayson, D.P., Hart, P., Chaytor, J.D., Ryan, H., Lee, H., Labay, K., Peterson, A., Liberty, L., 2014. New imaging of submarine landslides from the 1964 earthquake near Whittier, Alaska, and a comparison to failures in other Alaskan fjords. In: Krastel, S., et al. (Eds.), *Submarine Mass Movements and Their Consequences, Advances in Natural and Technological Hazards Research* 37, Springer, pp. 361–370. doi: 10.1007/978-3-319-00972-8_32

Heirman, K., De Batist, M., Charlet, F., Moernaut, J., Chapron, E., Brümmer, R., Pino, M., Urrutia, R., 2011. Detailed seismic stratigraphy of Lago Puyehue: implications for the mode and timing of glacier retreat in the Chilean Lake District. *Journal of Quaternary Science* 26(7), 665–674.

Heirman, K., 2011. A wind of change: Changes in position and intensity of the southern hemisphere westerlies during oxygen isotope stages 3, 2 and 1. PhD thesis, Ghent University, 227 pp.

Hirst, T.J., Richards, A.F., 1977. In situ pore-pressure measurement in Mississippi delta front sediments. *Marine Geotechnology* 2(1–4), 191–204.

Hjelstuen, B.O., Haflidason, H., Sejrup, H.P., Lysa, A., 2009. Sedimentary processes and depositional environments in glaciated fjord systems – Evidence from Nordfjord, Norway. *Marine Geology* 258(1–4), 88–99.

Hovland, M., Gardner, J.V., Judd, A.G., 2002. The significance of pockmarks to understanding fluid flow processes and geohazards. *Geofluids* 2, 127–136.

Hustoft, S., Dugan, B., Mienert, J., 2009. Effects of rapid sedimentation on developing the Nyegga pockmark field: Constraints from hydrological modeling and 3-D seismic data, offshore mid-Norway. *Geochemistry, Geophysics, Geosystems* 10(6), Q06012. doi: 10.1029/2009GC002409

Karlin, R.E., Holmes, M., Abella, S.E.B., Sylwester, R., 2004. Holocene landslides and a 3500-year record of Pacific Northwest earthquakes from sediments in Lake Washington. *Geological Society of America Bulletin* 116, 94–108.

Karlsson, R., Viberg, L., 1967. Ratio of c/p in Relation to Liquid Limit and Plasticity Index, with special reference to Swedish Clays. Swedish Geotechnical Institute Report 23, 43–47.

Kremer, K., Hilbe, M., Simpson, G., Decrouy, L., Wildi, W., Girardclos, S., 2015. Reconstructing 4000 years of mass movement and tsunami history in a deep peri-Alpine lake (Lake Geneva, France-Switzerland). *Sedimentology* 62, 1305–1327.

Kvalstad, T.J., Andresen, L., Forsberg, C.F., Berg, K., Bryn, P., Wangen, M., 2005. The Storegga slide: evaluation of triggering sources and slide mechanics. *Marine and Petroleum Geology* 22, 245–256.

Lacoste, A., Vendeville, B.C., Mourgues, R., Loncke, L., Lebacqz, M., 2012. Gravitational instabilities triggered by fluid overpressure and downslope incision – insights from analytical and analogue modeling. *Journal of Structural Geology* 42, 151–162.

Lafuerza, S., Sultan, N., Canals, M., Frigola, J., Berné, S., Jouet, G., Galavazi, M., Sierro, F.J., 2009. Overpressure within upper continental slope sediments from CPTU data, Gulf of Lion, NW Mediterranean Sea. *International Journal of Earth Sciences* 98, 751–768.

Lafuerza, S., Sultan, N., Canals, M., Lastras, G., Cattaneo, A., Frigola, J., Costa, S., Berndt, C., 2012. Failure mechanisms of Ana Slide from geotechnical evidence, Eivissa Channel, Western Mediterranean Sea. *Marine Geology* 307–310, 1–21.

Lastras, G., Canals, M., Urgeles, R., Hughes-Clarke, J.E., Acosta, J., 2004. Shallow slides and pockmark swarms in the Eivissa Channel, western Mediterranean Sea, *Sedimentology* 51, 837–850.

Laugenie, C., 1982. La région des lacs, Chili méridional. PhD thesis, Université de Bordeaux III, Bordeaux, France.

Leynaud, D., Mienert, J., Vanneste, M., 2009. Submarine mass movements on glaciated and non-glaciated European continental margins: A review of triggering mechanisms and preconditions to failure. *Marine and Petroleum Geology* 26, 618–632. doi:10.1016/j.marpetgeo.2008.02.008

Long, H., Flemings, P.B., Germaine, J.T., 2007. Interpreting in situ pressure and hydraulic properties with borehole penetrometers in ocean drilling: DVTTP and Piezoprobe deployments at southern Hydrate Ridge, offshore Oregon. *Journal of Geophysical Research* 112(B4), B04101. doi:10.1029/2005JB004165

Lunne, T., Robertson, P.K., Powell, J.J.M., 1997. Cone Penetration Testing in Geotechnical Practice, 312 pp., Spon Press.

McAdoo, B., Pratson, L.F., Orange, D.L., 2000. Submarine landslide geomorphology, US continental slope. *Marine Geology* 169, 103–136.

McCulloch, R.D., Bentley, M.J., Purves, R.S., Hulton, N.R.J., Sugden, D.E., Clapperton, C.M., 2000. Climatic inferences from glacial and palaeoecological evidence at the last glacial termination, southern South America. *Journal of Quaternary Science* 15, 409–417.

Mello, U.T., Karner, G.D., and Anderson, R.N., 1994. A physical explanation for the positioning of the depth to the top of overpressure in shale-dominated sequences in the Gulf Coast Basin, United States. *Journal of Geophysical Research - Solid Earth* 99(B2), 2775–2789. doi:10.1029/93JB02899

Meruane, C., 2005. El efecto del viento Puelche sobre la hidrodinámica y calidad de aguas del Lago Villarrica. M.Sc. Thesis, Universidad de Chile, 123pp.

Moernaut, J., De Batist, M., Heirman, K., Van Daele, M., Pino, M., Brümmer, R., Urrutia, R., 2009. Fluidization of buried mass-wasting deposits in lake sediments and its relevance for paleoseismology: Results from a reflection seismic study of lakes Villarrica and Calafquén (South-Central Chile). *Sedimentary Geology*, 213, 121–135.

Moernaut, J., De Batist, M., 2011. Frontal emplacement and mobility of sublacustrine landslides: Results from morphometric and seismostratigraphic analysis. *Marine Geology* 285(1–4), 29–45.

Moernaut, J., Van Daele, M., Heirman, K., Fontijn, K., Strasser, M., Pino, M., Urrutia, R., De Batist, M., 2014. Lacustrine turbidites as a tool for quantitative earthquake reconstruction: New evidence for a variable rupture mode in south central Chile, *Journal of Geophysical Research - Solid Earth* 119, 1607–1633, doi:10.1002/2013JB010738.

Moernaut, J., Van Daele, M., Strasser, M., Clare, M.A., Heirman, K., Viel, M., Cardenas, J., Kilian, R., Ladrón de Guevara, B., Pino, M., Urrutia, R., De Batist, M., 2015. Lacustrine turbidites produced by surficial slope failure: a mechanism for continuous and sensitive turbidite paleoseismic records. *Marine Geology*, in press. doi: 10.1016/j.margeo.2015.10.009

Naudts, L., Greinert, J., Artemov, Y., Staelens, P., Poort, J., Van Rensbergen, P., De Batist, M., 2006. Geological and morphological setting of 2778 methane seeps in the Dnepr paleo-delta, northwestern Black Sea. *Marine Geology* 227, 177–199.

Ndiaye, M., Clerc, N., Gorin, G., Girardclos, S., Fiore, J., 2014. Lake Neuchâtel (Switzerland) seismic stratigraphic record points to the simultaneous Würmian deglaciation of the Rhône Glacier and Jura Ice Cap. *Quaternary Science Reviews* 85, 1–19.

Panieri, G., Camerlenghi, A., Cacho, I., Sanchez, C., Canals, M., Lafuerza, S., Herrera, G., 2012. Tracing seafloor methane emissions with benthic foraminifera: Results from the Ana submarine landslide (Eivissa Channel, Western Mediterranean Sea). *Marine Geology* 291–294, 97–112.

- Phillips, E., Merritt, J., Auton, C., Gollledge, N., 2007. Microstructures in subglacial and proglacial sediments: understanding faults, folds and fabrics, and the influence of water on the style of deformation. *Quaternary Science Reviews* 26, 1499–1528.
- Pinson, L.J.W., Vardy, M.E., Dix, J.K., Henstock, T.J., Bull, J.M., Maclachlan, S.E., 2012. Deglacial history of glacial lake Windermere, UK: implications for the central British and Irish Ice Sheet. *Journal of Quaternary Science* 28(1), 83–94. doi:10.1002/jqs.2595.
- Piper, D.J.W., Cochonat, P., Morrison, M.L., 1999. The sequence of events around the epicentre of the 1929 Grand Banks earthquake: initiations of debris flows and turbidity current inferred from sidescan sonar. *Sedimentology* 46, 79–97.
- Praet, N., Moernaut, J., Van Daele, M., Boes, E., Haeussler, P.J., Strupler, M., Schmidt, S., Loso, M.G., De Batist, M., 2016. Paleoseismic potential of sublacustrine landslide records in a high-seismicity setting (south-central Alaska). *Marine Geology*, in press. doi:10.1016/j.margeo.2016.05.004
- Reiche, S., Hjelstuen, B.O., Hafliðason, H., 2011. High-resolution seismic stratigraphy, sedimentary processes and the origin of seabed cracks and pockmarks at Nyegga, mid-Norwegian margin. *Marine Geology* 284(1–4), 28–39.
- Riboulot, V., Cattaneo, A., Sultan, N., Carziglia, S., Ker, S., Imbert, P., Voisset, M., 2013. Sea-level change and free gas occurrence influencing a submarine landslide and pockmark formation and distribution in deepwater Nigeria. *Earth and Planetary Science Letters* 375, 78–91.
- Sassa, S., Sekiguchi, H., 1999. Wave-induced liquefaction of beds of sand in a centrifuge. *Géotechnique* 49(5), 621–638.
- Sawyer, D.E., DeVore, J.R., 2015. Elevated shear strength of sediments on active margins: Evidence for seismic strengthening. *Geophysical Research Letters* 42, 10216–10221, doi:10.1002/2015GL066603.

- Seifert, A., Stegmann, S., Mörz, T., Lange, M., Wever, T., Kopf, A., 2008. In situ pore-pressure evolution during dynamic CPT measurements in soft sediments of the western Baltic Sea. *Geo-Marine Letters* 28, 213–227. doi: 10.1007/s00367-008-0102-x
- SHOA (Servicio Hidrográfico y Oceanográfico de la Armada de Chile), 1987. Lago Villarrica, scale 1:40000, Santiago, Chile.
- Siegel, J., Lizarralde, D., Dugan, B., Person, M., 2014. Glacially generated overpressure on the New England continental shelf: Integration of full-waveform inversion and overpressure modeling. *Journal of Geophysical Research: Solid Earth* 119, 3393–3409. doi: 10.1002/2013JB010278
- Skempton, A.W., 1969. The consolidation of clays by gravitational compaction. *Quarterly Journal of the Geological Society* 125, 373–411. doi:10.1144/gsjgs.125.1.0373
- Solheim, A., Berg, K., Forsberg, C.F., Bryn, P., 2005. The Storegga Slide complex: Repetitive large-scale sliding with similar cause and development. *Marine and Petroleum Geology* 22 (1–2), 97–107.
- Spötl, C., Reimer, P.J., Starnberger, R., Reimer, R.W., 2013. A new radiocarbon chronology of Baumkirchen, stratotype for the onset of the Upper Würmian in the Alps. *Journal of Quaternary Science* 28(6), 552–558.
- Stegmann, S., Moerz, T., Kopf, A., 2006. Initial Results of a new Free-Fall Cone Penetrometer (FF-CPT) for geotechnical in situ characterisation of soft marine sediments. *Norwegian Journal of Geology* 86 (3), 199–208.
- Stegmann, S., Strasser, M., Anselmetti, F., Kopf, A., 2007. Geotechnical in situ characterization of subaquatic slopes: The role of pore pressure transients versus frictional strength in landslide initiation. *Geophysical Research Letters* 34(7), L07607. doi: 10.1029/2006GL029122
- Stegmann, S., Sultan, N., Kopf, A., Apprioual, R., Pelleau, P., 2011. Hydrogeology and its effect on slope stability along the coastal aquifer of Nice, France. *Marine Geology* 280, 168–181.

Steiner, A., 2013. Sub-sea floor characterization and stability of submarine slope sediments using dynamic and static piezocone penetrometers. University of Bremen, Bremen. PhD Thesis, 325 pp.

Strasser, M., Stegmann, S., Bussmann, F., Anselmetti, F.S., Rick, B., Kopf, A., 2007. Quantifying subaqueous slope stability during seismic shaking: Lake Lucerne, Central Switzerland, as model for ocean margins. *Marine Geology* 240, 77–97

Strasser, M., Hilbe, M., and Anselmetti, F.S., 2011. Mapping basin-wide subaquatic slope failure susceptibility as a tool to assess regional seismic and tsunami hazards: *Marine Geophysical Research*, 32(1–2), 331–347.

Strasser, M., Monecke, K., Schnellmann, M., Anselmetti, F.S., 2013. Lake sediments as natural seismographs: A compiled record of Late Quaternary earthquakes in Central Switzerland and its implication for Alpine deformation. *Sedimentology* 60, 319–341. doi: 10.1111/sed.12003.

Strozyk, F., Strasser, M., Förster, A., Kopf, A., Huhn, K., 2010. Slope failure repetition in active margin environments: Constraints from submarine landslides in the Hellenic fore arc, eastern Mediterranean. *Journal of Geophysical Research* 115, B08103, doi:10.1029/2009JB006841.

Sully, J.P., Robertson, P.K., Campanella, R.G., Woeller, D.J., 1999. An approach to evaluation of field CPTU dissipation data in overconsolidated fine-grained soils. *Canadian Geotechnical Journal* 36, 369–381.

Sultan, N., Cochonat, P., Dennielou, B., Bourillet, J.F., Savoye, B., Colliat, J.L., 2000. Surconsolidation apparente et pression osmotique dans un sédiment marin (Apparent Overconsolidation and Osmotic pressures in marine sediments). *Comptes Rendue de la Académie des Sciences, Science de la Terre et des planètes* 331(5), 379–386. doi: 10.1016/S1251-8050(00)01419-1

Sultan, N., Cattaneo, A., Sibuet, J.-C., Schneider, J.-L., and the Sumatra Aftershocks team, 2009. Deep sea in situ excess pore pressure and sediment deformation off NW Sumatra and its relation with the December 26, 2004 Great Sumatra-Andaman Earthquake. *International Journal of Earth Sciences* 98, 823–837.

Sultan, N., Savoye, B., Jouet, G., Leynaud, D., Cochonat, P., Henry, P., Stegmann, S., Kopf, A., 2010. Investigation of a possible submarine landslide at the Var delta front (Nice slope – SE France). *Canadian Geotechnical Journal* 47, 486–496.

Sultan, N., Lafuerza, S., 2013. In situ equilibrium pore-water pressures derived from partial piezoprobe dissipation tests in marine sediments. *Canadian Geotechnical Journal* 50(12), 1294–1305.

Tanaka, M., Watabe, Y., Tomita, R., Kamei, T., 2012. Effects of Diatom Microfossils Content on Physical Properties of Clays, in: *Proceedings of the Twenty-second International Offshore and Polar Engineering Conference*, edited, Rhodes, Greece, June 17-22, 2012.

Urlaub, M., Talling, P.J., Zervos, A., Masson, D., 2015. What causes large submarine landslides on low gradient ($<2^\circ$) continental slopes with slow (~ 0.15 m/kyr) sediment accumulation? *Journal of Geophysical Research - Solid Earth* 120, 6722–6739, doi:10.1002/2015JB012347.

Van Daele, M., Moernaut, J., Silversmit, G., Schmidt, S., Fontijn, K., Heirman, K., Vandoorne, W., De Clercq, M., Van Acker, J., Wolff, C., Pino, M., Urrutia, R., Roberts, S.J., Vincze, L., De Batist, M., 2014. The 600 yr eruptive history of Villarrica Volcano (Chile) revealed by annually laminated lake sediments. *Geological Society of America Bulletin* 126(3–4), 481–498. doi: 10.1130/B30798.1

Van Daele, M., Moernaut, J., Doom, L., Boës, E., Fontijn, K., Heirman, K., Vandoorne, W., Hebbeln, D., Pino, M., Urrutia, R., Brümmer, R., De Batist, M., 2015. A comparison of the sedimentary records of the 1960 and 2010 great Chilean earthquakes in 17 lakes: implications for quantitative lacustrine paleoseismology. *Sedimentology* 62(5), 1466–1496. doi: 10.1111/sed.12193.

Van Rensbergen, P., De Batist, M., Beck, C., Manalt, F., 1998. High resolution seismic stratigraphy of late Quaternary fill of Lake Annecy (Northwestern Alps): evolution from glacial to interglacial sedimentary processes. *Sedimentary Geology* 117, 71–96.

Van Rensbergen, P., De Batist, M., Beck, C., Chapron, E., 1999. High-resolution seismic stratigraphy of glacial to interglacial fill of a deep glacigenic lake: Lake Le Bourget, Northwestern Alps, France. *Sedimentary Geology* 128(1–2), 99–129.

Vanneste, M., Sultan, N., Garziglia, S., Forsberg, C.F., L'Heureux, J.-S., 2014. Seafloor instabilities and sediment deformation processes: The need for integrated, multi-disciplinary investigations. *Marine Geology* 352, 183–214.

Vendeville, B.C., Gaullier, V., 2003. Role of pore-fluid pressure and slope angle in triggering submarine mass movements: natural examples and pilot experimental models. In: Locat, J., Mienert, J. (eds.), *Submarine Mass Movements and their Consequences*, First International Symposium. Kluwer Academic Publishers, Netherlands, pp. 137–144.

Waldmann, N., Anselmetti, F.S., Ariztegui, D., Austin, J.A., Pirouz, M., Moy, C.M., Dunbar, R., 2011. Holocene mass-wasting events in Lago Fagnano, Tierra del Fuego (54S): implications for paleoseismicity of the Magallanes-Fagnano transform fault. *Basin Research* 23(2), 171–190. doi: 10.1111/j.1365-2117.2010.00489.x

Wiemer, G., Moernaut, J., Stark, N., Kempf, P., De Batist, M., Pino, M., Urrutia, R., Strasser, M., Kopf, A., 2015. The role of sediment composition and behavior under dynamic loading conditions on slope failure initiation: a study of a subaqueous landslide in earthquake-prone South-Central Chile. *International Journal of Earth Sciences* 104(5), 1439–1457. doi: 10.1007/s00531-015-1144-8.

You, Y., Flemings, P., Mohrig, D., 2012. Dynamics of dilative slope failure. *Geology* 40(7), 663–666. doi: 10.1130/G32855.1

Figure captions:

Figure 1: A) Setting of Lake Villarrica in the subduction context of South-Central Chile with indication of the main rupture areas (co-seismic slip > 5m) of the 2010 (M_w 8.8) and 1960 (M_w 9.5) megathrust earthquakes. LOFZ: Liquiñe-Ofqui Fault Zone; Triangles: active volcanoes. B) Lake Villarrica and its drainage basin (white line) with indication of active volcanoes (triangles) and main rivers (black lines).

Figure 2: Overview bathymetric map of Lake Villarrica with indication of the main rivers and moraine belts. Contour lines every 10 m. Bathymetry was calculated via interpolation between seismic data (Moernaut et al., 2009) and bathymetric data points of the Servicio Hidrográfico y Oceanográfico de la Armada de Chile (SHOA, 1987).

Figure 3: A) Multibeam bathymetric map for the SW part of Lake Villarrica (location: see Fig. 2) with indication of the main geotechnical testing sites (main CPT sites). The bottom morphology exhibits outcropping rock basement highs, a moraine ridge and landslide scarps. The mass-transport deposits (MTD) are presented in Moernaut et al. (2009). B) Close-up of lake bottom morphology near the main CPT and coring sites (VILLAR1 and VILLAR8). Four CPT sites (1.1, 1.2, 1.3 and 4) are located on an unfailed platform almost completely surrounded by headwall scarps. Detailed mapping of the headscarp is presented in Fig. 8.

Figure 4: A) 3.5 kHz pinger profile of the “unfailed platform” with the projected location of the main CPT sites and sediment core VILLAR1 (location on Fig. 3B). TWT: two-way travel time. R1 to R3 and SU-B to SU-D correspond to strong reflectors and seismic units in the seismic stratigraphy, respectively (see section 4.1). B) Sediment core VILLAR1. Lithological description and grain-size data (1 cm thick sample every 25 cm) allow differentiation of three lithological units (U1-U3). Peaks in γ -density (measured at 0.5 cm resolution) correlate with high-amplitude reflections and allow accurate core-to-seismic correlation. Strong reflectors correspond to distinct sandy tephra layers: R1 = T1 (“Pucón ignimbrite” of Fontijn et al., 2016); R2 = SFPT (“Failure Plane Tephra”) from Wiemer et al. (2015). C) In-situ undrained shear strength (s_u) obtained by CPTU testing (modified from Wiemer et al., 2015). Peaks correspond to tephra layers. U1 shows a general increase of s_u with depth, whereas U2 shows a slight decrease. s_u values drop substantially from U2 to U3.

Figure 5: Different types of pore pressure dissipation curves obtained in the present study, according to the classification of Sully et al. (1999). Type I is the standard curve whereas Type III, IV and V are typical for overconsolidated soils. The right figure indicates the different parameters in the calculation of overpressure ratio (λ^*) from a completed dissipation test.

Figure 6: Sparker seismic profile (location on Fig. 2) illustrating the complete sedimentary infill overlying the rock basement in the SW part of Lake Villarrica (see section 4.1). Dotted and dashed black lines are unconformities. SU-C2 corresponds to U3 in VILLAR1 (Fig. 4) and SU-D to U1-U2. Recurrent landsliding is evidenced by stacked mass-transport deposits (MTDs) within unit SU-D and an excavated (failed) slope bordered by scarps. Pockmarks were identified in the westernmost part of the profile (see Fig.9). Identification of the rock basement reflector is hampered by “multiple” reflections and loss of seismic wave energy with depth.

Figure 7: Brittle deformation in the glacio-lacustrine rhythmites of U3, potentially driven by fluid overpressure. The right picture is an outcrop near glacial Lake Rupanco (40.788 °S; 72.688 °W), which has a similar piedmont setting as Lake Villarrica, and shows a complex network of sub-vertical fractures in glacio-lacustrine rhythmites.

Figure 8: A) Multibeam bathymetric map with location of slide scarps (black lines) identified on multibeam data and seismic profiles. Locations of seismic profiles in subfigures C-E are indicated in blue. Areas 1 and 2 are presented in more detail on Fig. 9. The AD 1960 slope failures (grey lines) are based on Moernaut et al. (2015). White squares represent the main CPT sites. Dashed lines: assumed trace of slide scar. B) Paleobathymetry of the R2 surface, which is the main sliding surface for the studied landslides. This map is made by interpolation between pinger seismic profiles (see seismic grid in SI-Fig. 1). In Area2, the landslide scarp does not mimic the (paleo-)bathymetric contour lines. C) Pinger profile at the “unfailed platform” showing a headwall scarp, which is located at a zone of focused fluid escape. Fluid escape is identified by acoustic wipe-outs and the breaching of reflector R2. The post-landslide stratigraphy is also disturbed indicating that fluid escape (also) took place after the slope failure event. D) Pinger profile where the headwall scarp is located above a slope break in the R2 horizon. The R2 horizon is not affected by fluid escape. E) Pinger profile indicating focused fluid escape below pockmarks (see Fig.9A) and at the headwall scarp. Fluid escape areas are located above highs in SU-B.

Figure 9: A-B) Pockmarks and other expressions of focused fluid escape identified on multibeam bathymetry data and pinger seismic profiles (see legend) in Area 1 and 2 (location see Fig. 8). In Area 1, isolated pockmarks occur upslope of a relatively smooth and linear headscarp. In Area 2, fluid escape areas correspond to an irregular trace of the headwall scarp. The mapped fluid escape features on the seismic profiles (colored lines) match with the fluid escape areas (dotted closed lines) identified on the slope gradient map of the lake bottom (sub-figure C). C) Slope gradient map used for identifying fluid escape (dotted closed lines) in study Area 2. Fluid escape areas correspond to irregular, overall horizontal topography visualized as green patches. Sub-horizontal (green) areas in the upslope western parts are associated to erosion and non-deposition (see Fig. 13A). Grey circles: short CPT sites (Fig. 10). D) Paleo-bathymetric map of the top of SU-B based on interpolation of seismic profiles (see seismic grid in SI-Fig. 1). Many areas of fluid escape (red dotted lines) coincide with peaks in SU-B. The “unfailed platform” is located where the upper boundary of SU-B is relatively flat.

Figure 10: A) Map of Area2 (location on Fig. 8) showing fluid escape areas (red dashed lines) and the different stratigraphic levels used as sliding surface (update of Wiemer et al., 2015). The sites of short CPTU measurements (numbered squares) are color coded as in sub-figure B and their site numbers do not correspond to those of the long CPTU testing sites. Arrows indicate the directions of mass-transport during slope failure derived from multibeam bathymetry. B) Undrained shear strength (s_u) from short-style CPTU tests in the uppermost 2 m of sediments. Only the moving average (19 points) is presented. Measurements in the unfailed platform (red) have higher s_u values than those in the post-failure sediments of the failed slope (green). Fluid escape areas (blue) have a significantly lower s_u which plots close to the “normal consolidation line” for diatomaceous sediments.

Figure 11: A) Pinger profile near CPT4 site and core VILLAR8 (location on Fig. 3). Color coded reflectors and units as in Fig. 4. The upper part of U2 and lower part of U1 is excavated by the landslide event, which locally developed onto the “shallow sliding surface U2a” (location on Fig. 10A). B) Sediment core VILLAR8. As in VILLAR 1, lithological description and grain-size data (1 cm sample every 25 cm) clearly show three units. Density peaks correspond to strong reflections in the seismic profile. Void ratio shows increasing values with depth in U2 (excluding negative peaks associated to tephra layers). C) In-situ undrained shear strength (S_u) obtained by CPTU testing. Peaks correspond to tephra layers. U1 shows a steep linear increase of s_u with depth, whereas U2 shows a general decrease. s_u values in U3 are as low as in the lower U2.

Figure 12: Examples of CPTU dissipation tests in Lake Villarrica. u_h = hydrostatic pressure; σ_{vo} = vertical total stress. For illustrative purposes, σ_{vo} is used instead of σ'_{vo} . Four of the five types of Sully et al. (1999) were observed in our dataset (see Fig. 5). B-C) Extrapolation of incomplete dissipation curves via the “inverse-time” technique. We used the $1/t$ and $1/Vt$ approach to estimate an upper and lower bound for the ambient pore pressure state (see section 3.3 and SI-Fig. 2). F) Example of a complete dissipation curve in a tephra layer of supposed high permeability.

Figure 13: A-B) Pinger profiles with stratigraphic location of CPTU dissipation tests. Numbers represent the overpressure ratio (λ^*). Red: λ^* values above 0.5. Bold and underlined: most reliable values based on complete dissipation during the testing period. C) Map of the different sliding surfaces (see Fig. 10A) with location of shown profiles and λ^* value for the lower U2 or U3. D) West-east transect of CPTU sites where both s_u and λ^* was determined. s_u values are visually cropped at 30 kPa. U2 shows a decreasing s_u trend with depth. Higher λ^* value correspond to lower s_u (see Fig. 14) and steeper negative trends. The dotted line represents the range in which sediments are considered to be normally consolidated.

Figure 14: Overpressure ratio (λ^*) versus normalized undrained shear strength (s_u/σ'_{vo}) for the platform sequences and below the failed slope. Horizontal bars indicate the range of λ^* based on extrapolation of incomplete dissipation tests. Values of $\lambda^*>1$ are physically unlikely and probably relate to the $1/t$ extrapolation procedure, which often overestimates the ambient pore pressure (see section 3.3). Grey data points: slope geometry at the moment of failure was reconstructed based on adjacent seismic profiles in order to estimate σ'_{vo} at the considered level at the moment of failure (see section 4.7).

Figure 15: Depositional history of Lake Villarrica since the start of deglaciation, and model for the development of excess pore pressure and fluid focusing facilitating landslide development. The different depositional stages are described in section 4.1 and the overpressure and fluid flow mechanism is discussed in section 5.6.5. Stages A-D took place from ~17.5 ka BP to 12.7 ka BP when sedimentary processes are tightly related to the presence of a glacier in the lake basin (A-C) or in the fluvial catchment (D). Stage E represents the hemipelagic open lacustrine sedimentation during the Holocene, and stage F the slope failure event discussed in this article.

Table 1: CPTU pore pressure dissipation data at the different sites used in this article (see section 3.3 for methodology) with u_h = hydrostatic pressure; σ'_{vo} = hydrostatic vertical effective stress; u^* = overpressure; λ^* = overpressure ratio. The

u_h is calculated for fresh water for the depth reached by the CPTU probe. σ'_{vo} is derived from the γ -density data on sediment cores. The used density value for CPT2 (mainly U3) and CPT5 (sequence of U1-U2-U3) is 1.6 g/cm^3 and 1.5 g/cm^3 , respectively. This is based on spatial extrapolation of the seismic stratigraphy on the unfailed platform and on the average γ -density value measured for each unit in the sediment cores.

Figure 1. 1.5 columns wide

Figure 2. 1 column wide

Figure 3. 2 columns wide

Figure 4. 1 or 1.5 columns wide

B) VILLAR1

Figure 5. 1 column wide

Figure 6. 1.5 columns wide

Figure 7. 1 column wide

villa754 (CPT2)

VILLAR8 (CPT4)

Outcrop west shore Lake Rupanco

Figure 8. 2 columns wide

Figure 9. 1.5 columns wide

Figure 10. 1.5 columns wide

Figure 11. 1.5 columns wide

B) VILLAR8

C) CPT4

Figure 12. 1.5 columns wide

Figure 13. 1.5 to 2 columns wide

Figure 14. 1 column wide

Figure 15. 1.5 columns wide

Table 1

Position	Depth (m)	Time (min)	Ty pe	Measured				Extrapolated via 1/t		Extrapolated via 1/√t		Trend at last data	Concavity last data	Average extrap. λ*	Lithology at dissipation
				u _h (kPa)	σ' _{vo} (kPa)	u* (kPa)	λ*	u* (kPa)	λ*	u* (kPa)	λ*				
CPT 1.2	4.1	27	III	260.02	16.92	6.48	0.38	4.50	0.27	0.50	0.03	decreasing	upwards	0.15	U2
CPT 1.2	5.5	40	V	266.96	23.46	17.00	0.78	x	x	x	x	rising	X	x	U2
CPT 1.2	5.9	30	V	274.88	25.14	5.12	0.20	x	x	x	x	rising	X	x	T2 in U2
CPT 1.2	7.0	40	V	281.67	30.28	12.83	0.42	12.82	0.42	12.83	0.42	constant	X	0.42	Pumice in U2
CPT 1.1	7.8	45	V	266.52	34.51	10.48	0.30	x	x	x	x	rising	X	x	Pumice in U2
CPT 1.3	8.3	42	IV	331.42	36.72	40.50	1.10	29.50	0.80	19.50	0.53	decreasing	0	0.67	U2
CPT 4	8.4	35	IV	422.40	36.80	51.60	1.40	39.00	1.06	25.00	0.68	decreasing	0	0.87	U3
CPT 3	1.5	39	III	412.22	5.56	3.80	0.68	3.10	0.56	1.80	0.32	decreasing	upwards	0.44	T2 in U2
CPT 3	3.2	49	III	430.39	11.23	34.11	3.04	19.61	1.75	6.61	0.59	decreasing	upwards	1.17	U3
CPT 2	3.4	46	V	331.35	20.02	17.65	0.88	17.65	0.88	17.65	0.88	constant	X	0.88	T3 in U3
CPT 5	4.7	40	IV	233.62	22.8	32.88	1.44	24.80	1.09	16.80	0.74	decreasing	upwards	0.91	U3
CPT 6	4.9	41	I	939.00	19.3	3.00	0.16	1.80	0.10	0.00	0.00	decreasing	upwards	0.05	U1
CPT 6	2.60	34.00	V	916.00	10.1	0.70	0.07	0.50	0.05	0.30	0.03	decreasing	upwards	0.04	U1