Lake-level rise in the late Pleistocene and active subaquatic volcanism since the Holocene in Lake Kivu; East African Rift

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Abstract

The history of Lake Kivu is strongly linked to the activity of the Virunga volcanoes. Subaerial and subaquatic volcanoes, in addition to lake-level changes, shape the subaquatic morphologic and structural features in Lake Kivu's Main Basin. Previous studies revealed that volcanic eruptions blocked the former outlet of the lake to the north in the late Pleistocene, leading to a substantial rise in the lake level and subsequently the present-day thermohaline stratification. Additional studies have speculated that volcanic and seismic activity threaten to trigger a catastrophic release of the large amount of gases dissolved in the lake. A bathymetric mapping and seismic profiling survey covered the volcanically active area of the Main Basin at a high resolution unprecedented for Lake Kivu. New geomorphologic features identified on the lake floor can accurately describe related lake-floor processes for the first time. The late Pleistocene lowstand is observed at 425 m depth, and

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volcanic cones, tuff rings, and lava flows observed above this level indicate both subaerial and subaquatic volcanic activity during the Holocene. The geomorphologic analysis yields new implications on the geologic processes that have shaped Lake Kivu's basin, and the presence of young volcanic features can be linked to the possibility of a lake overturn.

Keywords:

rift lake, lake-level rise, subaquatic volcano, subaquatic lava flow, limnic eruption, high-resolution bathymetry

1 **1. Introduction**

2 Together with Lakes Nyos and Monoun in Cameroon, Lake Kivu is re-
3 puted for being one of the three 'exploding lakes' in Africa (Halbwachs et al.,
4 2004; Kling et al., 1987). Lakes Monoun and Nyos erupted in the 1980s re-
5 sulting in nearly 2000 fatalities. A limnic eruption, commonly referred to as a
6 lake 'explosion', is a result of oversaturation of dissolved gases that accumu-
7 late in the stratified depths of the lake. Lake Kivu's large size and population
8 density extends the risk of a limnic eruption to millions of people. The lake's
9 permanently stratified deepwater contains vast quantities of dissolved CO₂;
10 300 km³ at STP (standard temperature and pressure), and CH₄; 60 km³
11 at STP (Schmid et al., 2005). These gases would erupt from the stratified
12 depths, and consequently mix the lake, if their combined partial pressures
13 were to exceed the hydrostatic pressure of the lake at all depths (Schmid
14 et al., 2005). The CO₂ is a product of magmatism and active volcanism in
15 the lake basin (Degens et al., 1973; Tedesco et al., 2010), whereas the CH₄
16 is produced by microbial activity in the anoxic sediment (Tietze et al., 1980;
17 Pasche et al., 2011; Bhattarai et al., 2012). CO₂ and CH₄ gas accumulate
18 in the deepwater below the main density gradient centred at 255 m depth.
19 This gradient is sustained by fresh groundwater inputs from above, and salt
20 and CO₂ entering with hydrothermal inputs at greater depths (Schmid et al.,
21 2005; Ross et al., 2014).

22 Volcanic activity has played an important role in the evolution of Lake
23 Kivu, especially that within the Virunga Volcanic Province (VVP) (e.g.,
24 Smets et al., 2010b; d'Oreye et al., 2011). The cities Goma (DRC; the
25 Democratic Republic of Congo) and Gisenyi (Rwanda), located on the north-
26 ern shoreline of the lake (Figure 1), are constructed almost entirely on lava
27 flows. Furthermore, the presence of phreatomagmatic cones along the north-

ern shoreline document the historical occurrences of eruptive events at the lake boundary (Capaccioni et al., 2003). Haberyan and Hecky (1987) estimated that Lake Kivu’s northern outflow to Lake Edward was dammed by the lavas of the VVP sometime in the late Pleistocene. This blockage was followed by lake-level rise that eventually drained south along a border fault to Lake Tanganyika at $\sim 10\text{-}14$ ka, which is occupied by the present-day ‘Ruzizi River’ in Figure 1 (Degens et al., 1973).

Several studies have been made on the sediments of Lake Kivu and Lake Tanganyika that date the geological developments contributing to the current state of Lake Kivu. An overview of the proposed events is given from the varying data in the publications, and summarized in Table 1. Core dating on the sediment taken by Degens et al. (1973) in the 1970s was based on radiocarbon measurements of bulk organic carbon, and therefore could be anomalously old due to hydrothermal inputs of CO_2 . In summary, Lake Kivu’s stratification (state of meromixis) was initiated with its lake-level rise at $\sim 10\text{-}14$ ka (Table 1). Its alteration from a thermal to a thermohaline stratification was inferred to have been initiated within the last 6 kyrs according to Hecky and Degens (1973). The researchers postulated that the change was presented with the onset of hydrothermal spring activity within the stratified depths of the lake (Degens and Kulbicki, 1973). At least two periods are described in literature where there has been a breakdown of the thermohaline stratification linked to intense hydrothermal activity from volcanism within the last 6 kyrs (Table 1). Recent acquisition of sediment cores, and radiocarbon dating at subaquatic-tephra layers in Lake Kivu have determined a date of ~ 0.8 ka for the last lake-overturn event. Ongoing hydrothermal spring inputs into the deepwater of the main basin were identified based on data obtained by CTD (conductivity, temperature, and depth) profiling (Ross et al., 2014). The lack of knowledge of the lake-floor processes relating to hydrothermal spring activity were the motivation to produce the high-resolution bathymetric map and seismic profiles that are presented here.

The newly constructed map of the Main Basin delineates the geomorphology in Lake Kivu at a resolution that is unprecedented for African rift lakes. Furthermore, our dataset represents the first high-resolution geophysical analysis on the sediment. The Main Basin hosts a variety of geologic structures and processes including: lava flows (e.g., Halbwachs et al., 2004; Komorowski, 2003), faults/fractures (e.g., Peeters, 1957; Degens et al., 1973; Wong and von Herzen, 1974), hydrothermal vents (Degens and Kulbicki, 1973; Schmid et al., 2005; Tassi et al., 2009), earthquake epicentres (e.g.,

66 Kavotha et al., 2003; Wong and von Herzen, 1974; d’Oreye et al., 2011;
67 Mavonga et al., 2006, 2010), and dike/magma intrusions (Wauthier et al.,
68 2010, 2012). Volcanic and seismic activities have significantly affected Lake
69 Kivu, but the question concerning the possibility of a lake overturn and expul-
70 sion of gas triggered by volcano-tectonic activity remains debated. There is
71 no historical documentation of an event causing a lake overturn, and the last
72 overturn is thought to have occurred at ~ 0.8 ka (Table 1). Here we discuss
73 new and crucial insights into the development of Lake Kivu’s basin, which
74 are interpreted from the lake-floor processes that shape its geomorphology.
75 The unprecedented look into Lake Kivu’s sublacustrine environment provides
76 a more accurate assessment of its current geologic state, with implications
77 for the possibility of a lake overturn event as a result of subaquatic volcanic
78 activity.

79 **2. Geological setting of the Lake Kivu basin**

80 *2.1. Location and morphology*

81 Lake Kivu, shared by the Democratic Republic of Congo (DRC) and
82 the Republic of Rwanda (Rwanda) (Figure 1), is a rift lake in the Western
83 Rift branch of the East African Rift system (EARS). The EARS is an intra-
84 continental region of lithospheric extension and is volcanically and seismically
85 active (Ebinger, 1989). Most of the African great lakes sit within the EARS,
86 with the exception of Lake Victoria. The Western Rift branch is a seismically
87 active corridor that encircles the Archaean Tanzania craton (Ebinger, 1989)
88 and extends from Lake Albert in the north to Lake Malawi in the south. The
89 different rift basins that make up the Western Rift branch are separated from
90 each other by accommodation zones (Ebinger, 1989). Lake Kivu is centred on
91 the rift axis and is the highest of the western rift lakes, located at an altitude
92 of 1463 m asl (e.g., Degens et al., 1973). It has a surface area of 2386 km²,
93 a volume of 549 km³, and is 486 m in depth. The irregular shoreline of
94 Lake Kivu strongly differs from the other rift lakes in the western branch of
95 the EARS. The rift lakes generally present smoother and elongated shapes
96 that reflect the strong control of rift-basin border faults on their morphology.
97 Affirmation of the lake damming in the north is apparent from the orientation
98 of the paleo-hydrographic network, including the lake depth that increases
99 from the south to the north (Boutakoff, 1939).

100 The present-day lake comprises the Main Basin, consisting of two half-
101 graben valleys separated by the Idjwi Island horst, and four smaller basins

102 (Figure 1): Kabuno Bay, Kalehe Basin, Bukavu Bay, and the Ishungu Basin
 103 (Schmid and Wüest, 2012). The Main Basin is filled with a maximum sedi-
 104 ment accumulation of ~ 1.5 km at the west side of the basin (Wood, 2014).
 105 Studies correlating sedimentation rates with seismic stratigraphy in the Main
 106 Basin have calculated the age of Lake Kivu to be between the mid-Pleistocene
 107 (Haberyan and Hecky, 1987) and the Pliocene (Degens et al., 1973). How-
 108 ever, recent research along the eastern half graben of the Main Basin has
 109 revealed thicker sediments (Wood, 2014), which might extend Lake Kivu’s
 110 development to the Miocene.

111 *2.2. Volcanic and seismic activity*

112 Four volcanic regions constitute the Western Rift branch including the
 113 VVP. The volcanic activity in the VVP dates back to the mid-Miocene, but
 114 a majority of the volcanic edifices are expected to be younger than 100 ka BP
 115 (Rogers et al., 1998). The VVP is situated in a transfer zone characterized
 116 by several fault and fissure networks with different orientations. The main
 117 volcanic edifices are located at the intersection of these networks. Nyamu-
 118 lagira and Nyiragongo (Figure 1), are the two currently active volcanoes of
 119 the VVP and are distinguished by their effusive eruptions of lava flows and
 120 fountains. The cone and eruptive fissure distributions in the Nyamulagira
 121 and Nyiragongo volcanic fields generally coincide with the NE and NNW-
 122 trending fracture networks of the VVP. The silica-undersaturated composi-
 123 tion of their lava yields low viscosities and induces high-velocity flows (e.g.,
 124 Tazieff, 1977). As a consequence, their lava flows are thin and typically ex-
 125 tend tens of kilometres away from the source vent. Nyamulagira is Africa’s
 126 most active volcano with eruptions every 1-4 years (Smets et al., 2010a). Its
 127 lava field covers a surface area of over 1100 km² and contains more than 100
 128 flank cones. Nyiragongo is permanently active and hosts an active lava lake
 129 in its main crater. The last two fissural eruptions in 1977 and 2002 induced a
 130 draining of the lava lake. The 2002 eruption of Nyiragongo is hypothesized to
 131 be a result of rift tectonics on the basis of isotopic analysis of lava flows (Ko-
 132 morowski, 2003) and ground deformation modeling (Wauthier et al., 2010,
 133 2012). A best-fit model to explain the co-eruptive ground deformations was
 134 produced by Wauthier et al. (2012), and suggests the intrusion of a deep
 135 dyke extending southward beneath Lake Kivu.

136 In addition to volcanic activity, the Western Rift branch of the EARS
 137 is also defined by significant seismic activity. Herrincks (1959) listed ~ 800
 138 seismic events in the period between 1909 and 1954 in DRC, Rwanda, and

139 Burundi. The 2002 eruption of Nyiragongo occurred with several pre- co- and
140 post-eruptive earthquakes. Altogether, more than 100 tectonic earthquakes
141 of $M \geq 3.5$ were recorded on the southern flank of Nyiragongo and in the north-
142 ern part of Lake Kivu. The most recent and fatal Bukavu-Cyangungu (M
143 5.9) earthquake in February 2008 had a deleterious impact on infrastructure
144 and triggered landslides in the area (d’Oreye et al., 2011).

145 The volcanic and seismic events that have been recorded within Lake
146 Kivu have not been conducive to a limnic eruption at the current state of
147 dissolved CO_2 and CH_4 in the stratified layer. The 2002 eruption of Nyi-
148 ragongo induced lava flows that extended up to 100 m deep into Lake Kivu.
149 It was concluded that water temperatures were altered only at a local scale
150 (Lorke et al., 2004), i.e. without any threat to the lake stability. Similar
151 conclusions were reached for the 1938-40 eruption of the neighbouring Nya-
152 mulagira, where lava flows entered Kabuno Bay (Verhoogen, 1948). However,
153 the possible consequence of an eruption or a mass movement from within the
154 deepwater of the lake remains more difficult to elucidate.

155 3. Data acquisition and analysis

156 Bathymetric mapping of Lake Kivu’s Main Basin was conducted with a
157 50 kHz swath system (SeaBeam 1050) in October 2010 over an area of 224
158 km^2 and in water depths from 8 to 486 m (Figure 2a). Depths given in the
159 analysis and interpretation here within refer to water depth relative to the
160 present-day lake level. The map survey covered the area of the basin known
161 to host most of the geological activity as mentioned in the introduction. The
162 surveying system was mounted to the hull of the boat RV Gloria, and a bathy-
163 metric sounding was derived for each beam with up to 126 receive-beams at
164 120° athwartship swath. The survey was conducted at an average speed of
165 2 m/s during daylight hours. Positions were provided by a dual-differential
166 GPS-aided inertial navigation system. Boat movements were compensated
167 by a OCTANS 3000 motion sensor. Sound velocity profiles were determined
168 as a function of water density, which was calculated as a function of tem-
169 perature and salinity using the Chen-Millero equation (Chen and Millero,
170 1986). Temperature and salinity parameters were acquired via CTD profil-
171 ing with a RBR[®] XR-620 probe. The sound velocity at the transducer head
172 was measured continuously for accurate beam-steering. Data acquisition was
173 managed with Hydrostar Online, and HDP was used for post processing and
174 calibration. Data was exported as gridded xyz data from HDP and spikes

175 were removed with Fledermaus (version 7) for final editing. In the deeper
176 part of the basin, typically at depths greater than 300 m, the sediment is
177 characterized by a layer of low amplitude reflections at the sediment-water
178 interface that are thought to represent highly water-saturated sediments.
179 E.g. Figure 5, Figure 6, and Ross et al. (2013). The final processing of the
180 bathymetric data was limited by the beam penetrating this layer on occasion,
181 which produced a pronounced grain in the bathymetric map upon interpola-
182 tion. This is particularly evident at the locations of the outer beams where
183 the angle of incidence is the highest.

184 Visualization for the analysis of the merged data sets was conducted
185 with GMT 4.5.8 using the functions described within Wessel and Smith
186 (1991). The multibeam data were averaged using the *blockmean* function
187 onto 5 m grid spacing and then surface interpolation was conducted using
188 the *nearneighbour* algorithm on 5 and 10 m grids. For the geomorphologic
189 interpretation of the bathymetry, the 5 and 10 m interpolated data was evalu-
190 ated. A slope angle map was produced from the 10 m interpolated grid using
191 the *grdgradient* and *grdmath* functions to depict the first derivative of the
192 bathymetric data. Depth contours helped to highlight vertical steps at the
193 scale of the basin, and were determined using the function *grdcontour*. Pro-
194 file curvatures, defined as the degree of vertical convexity (negative values)
195 or concavity (positive values) of the relief along the line of maximum slope
196 were extracted from the bathymetric map using the DEM (Digital Elevation
197 Model) Surface Tools developed by Jenness (2012) and the Spatial Analyst
198 Extension for the ESRI ArcGIS® software. Both slope and profile curva-
199 tures highlight the horizontal limits of specific relief features such as fault
200 escarpments, volcanic cones, landslides, and terraces. The function *grdimage*
201 was used to produce the planar views of the interpolated 5 and 10 m grids
202 following the application of an averaging filter using the function *grdfilter*.
203 Finally, a combined bathymetric map was produced using a low-resolution
204 bathymetric map provided by Lahmeyer (1998), the present-day lake level
205 contour extracted from the 30 m ASTER Global DEM (GDEM) provided
206 by NASA, and by applying the function *grdblend*. The result was used to
207 calculate the area and volume at different depth contours for interpretation.

208 Cross-sectional analysis of the lake-floor subsurface was conducted from
209 high-resolution 2D reflection seismic data of the sediment fill. A 3.5 kHz
210 pinger single-channel system was used to digitally record data of the shallow
211 subsurface down to ~ 10 m depth in SEG-Y format with a flat gain. Pro-
212 cessing involved bandpass filtering (1600-6000 Hz) and a constant-velocity

213 migration for selected areas. Seismic data were interpreted in Kingdom Suite
214 Software.

215 Temperature anomalies from a background lake-temperature profile were
216 investigated to discern hydrothermal and cold groundwater inflows into Lake
217 Kivu (Ross et al., 2014). Locations where the most significant temperature
218 differences from the background profile were observed, over specific depth
219 ranges, are marked on the bathymetric map. The location of groundwater
220 discharge relative to the local morphology was interpreted to help conclude
221 possible geologic features.

222 4. Morphologic and Stratigraphic Analysis

223 The bathymetric map covers approximately a tenth of the area of Lake
224 Kivu (Figure 1). The area is subdivided into three regions in order to de-
225 scribe the various geologic features and structures based on their specific
226 geomorphological characteristics (Figure 2a). Geologic features are defined
227 within the limits of resolution and detail determined by the data, and in the
228 context of their volcanic and lacustrine setting. Generally, the morphology in
229 Lake Kivu’s Main Basin is shaped by hummocky volcanic relief in the north
230 and northwest regions; Precambrian rocks extending from the shore into the
231 basin from the Mbuji Peninsula, the Idjwi Island horst, and the drowned
232 river valleys; and enhanced sediment deposition covering deltaic regions, and
233 the flatter regions of the deep basin (Figure 2b). Furthermore, many of these
234 features are demarcated by four depth contours delineated in Figure 2b. The
235 425 m contour outlines the base of the isolated volcanic features, and the
236 extension of the Idjwi Island horst in the south. The 395 m contour outlines
237 the base of the volcanic platform in the northwest, and the eroded river val-
238 leys in the east. Lastly, the 315 and 250 m contours outline the base of the
239 steeply sloped walls at the northern limit of Mount Goma, and the tuff rings
240 and maars in the volcanic region.

241 4.1. Region 1

242 Region 1 (Figure 3) is characterized by the southward extension of the
243 VVP and is dominated in general by volcanic features. Numerous cones, tuff
244 rings and maars, and areas of hydrothermal and cold groundwater discharge
245 distinguish this region (Figure 3). A concentrated area of cones is outlined
246 in Figure 3, with the larger of these features marked as ‘small cones’ on
247 the map. The numerous maars or tuff rings are recognized by circle-shaped

248 crater-rims excavated into the sediment, and flat internal floors with rims
 249 built above the sediment, respectively (Figure 3). These features align in
 250 the same NE-SW orientation as the 'satellite cones' (Figure 3) linked to the
 251 Nyiragongo volcano. This linkage suggests their eruption to have occurred
 252 along a similar fault lineament. A large phreatomagmatic satellite cone,
 253 Mount Goma, is found at the eastern limit of Figure 3. Mount Goma's steep
 254 walled summit crater is used as a port for the city of Goma, whereas its main
 255 edifice is submerged and gently sloped. A total of three prominent subaquatic
 256 volcanic features are distinguished in Region 1: the first is a broad dome
 257 found south of the concentrated cones and surrounded by maars and tuff
 258 rings ('A' on Figure 4a); the second feature is the apparent lava flow located
 259 just south of this cone ('B' on Figure 4a); and the third is the cone directly
 260 west of Mount Goma, which lacks a summit crater ('C' on Figure 4a). In
 261 addition to the volcanic features, areas of concentrated groundwater inflows
 262 are observed in, and are specific to Region 1. Hydrothermal groundwater
 263 is observed in the area dominated by the maars/tuff rings and small cones,
 264 whereas cool groundwater flows can be found where the limit of the Mount
 265 Goma edifice abuts the steeply sloped wall to the northwest. Groundwater
 266 flows are likely discharged from high-velocity conduits, such as faults and
 267 fractures, that are not visible at the resolution of the map. Region 1 is
 268 further divided into frames i to iii to describe geologic features.

269 4.1.1. *Frame i*

270 The volcanic field in Frame i is linked to the southern extension of the
 271 phreatomagmatic cones of the Bulengo-Rusayo zone (d'Oreye et al., 2011).
 272 The most striking feature in Frame i is a ~ 3 km wide and ~ 200 m high dome
 273 that cuts the NE-SW alignment of maars and tuff rings ('A' on Figure 4a i).
 274 The slopes of the dome are on the order of 3 to 12° and appear smooth and
 275 constructional. Furthermore, the dome appears to be superimposed onto tuff
 276 rings on its western and eastern margin, indicating that the dome is younger
 277 than these features. The smooth lobes to the southeast are separated from
 278 the dome by a small valley, and the base of the dome disappears where the
 279 lobes terminate at ~ 395 m depth. The dome-like characteristic implies that
 280 it has a magmatic origin and could indicate a subaquatic volcanic eruption.
 281 However, there are no distinct morphologic traits to suggest that the dome
 282 was formed either as a subaquatic or subaerial feature.

283 The northern part of Frame i is composed of several grouped, or aligned,
 284 cones with diameters ranging from 10 to 100 m ('small volcanic cone' on

Figure 3). These features are interpreted to be created by the explosive interaction of lava and water-saturated sediment, defined as rootless cones by Hamilton et al. (2010). The cones are built over a large and irregular plateau, which likely corresponds to lava flows with fronts ending at a depth of ~ 250 m (Figure 4a i). Additional rootless cones can be seen in the concentrated area marked in Figure 3, which reach the dome and maar features in the map to depths of 315 m. The cones observed deeper than the 250 m contour are likely formed from older lava flows that were covered by subsequent sedimentation.

The maars and tuff rings located in Frame i are outlined at the base of their external flanks by both the 250 and 315 m contour (Figure 4a i). Maars, tuff rings, and phreatomagmatic edifices are created by the explosive interaction between lava and the groundwater aquifer or surface water (Zimanowski and Büttner, 2003). The wide and shallow morphology of the maars and the presence of tuff rings are indicative of shallow-water steam explosions (Lorenz, 1986). Consequently, we can assume a lake level equivalent to the ~ 315 and 250 m contours at the time of their formation. The high-resolution seismic profiles of the tuff rings along profile S3 and in the northeast corner in Figure 5 (seismic profile not shown) reveals layered sediments reaching an average depth of 4.2 m (std. dev. 1.1). The sedimentation in the crater is absent of unconformities and is assumed to be subsequent to the tuff formations and lake-level rise, i.e., eliminating the possibility of erosion, volcanic sediment accumulation, and isolated lakes (Figure 5 S3). Using the sedimentation rates of 30 cm per 1000 yrs calculated by Degens et al. (1973), we can assume their formation to be ~ 14 ka. This is a rough estimate of age, where recent sedimentation rates of up to 46 cm per 1000 yrs calculated by Ross et al. (2013) would yield an age of ~ 9 ka.

The terraces observed in the seismic sections in Frame i along profiles S1 and S2 are interpreted as paleoshore indicators, yielding evidence to the possibility of a lake-level rise that would have submerged the tuff rings. The terraces are on average 100 to 200 m across with heights of 5 to 10 m (Figure 5 S1 and S2). The horizontal alignment of the terraces between profiles S1 and S2 imply a lake-level transgression from the ~ 425 to 395 m contour. The presence of lake-level indicators has been observed in the seismic profiles of Lake Edward and Lake Malawi, although they represent delta progradation controlled by climatically-induced lake-level changes that are not comparable to the seismic facies observed here (McGlue et al., 2006; Lyons et al., 2011). However, studies from other large lakes documented paleoshoreline terraces of equal dimensions formed by 'in-place drowning' of beach ridges,

323 and formation of a new ridge at landward positions from rapid water-level
 324 rises (Gilli et al., 2005). An alternative theory is that the terraces are a
 325 product of block faulting in the lake. This would require fault lineaments
 326 in the N-S to NE-SW orientation, which aligns to the maars and tuff rings
 327 in the lake and to faults (e.g., Ebinger, 1989) and eruptive fissures observed
 328 onshore.

329 4.1.2. *Frame ii*

330 Mount Goma is a polygenetic, phreatomagmatic volcanic cone (Denaeyer,
 331 1975) consisting of an emerged steep summit crater and a submerged gen-
 332 tle sloping flank (Figure 4a ii). The summit of the volcano is composed of
 333 relatively recent ash deposits overlying at least two series of tuff layers (De-
 334 naeyer, 1975). The 2 to 10° slopes of the subaquatic flank of Mount Goma
 335 are characterized by wave-like features with 15 to 20° slopes, situated mainly
 336 along the western half of the flank (Figure 4a ii). The wave features reach a
 337 depth of ~430 m and appear as 'steps' in the seismic profile (Figure 5 S4).
 338 The steps observed in the seismic profiles could be interpreted as shoreline
 339 terraces; however, their inclination and morphology in the bathymetric map
 340 are not clearly linked to shoreline erosion. Phreatomagmatic cones typically
 341 release pyroclastic sediments, which in turn can form compression ridges as
 342 the sediment moves down slope (Clague, 2009). Therefore, these features
 343 could be explained by the stabilization of these pyroclastic deposits, reminis-
 344 cent of that found along the much deeper and larger slopes of the Hawaiian
 345 Ridge (Moore et al., 1989). An alternative theory is that the wave features
 346 are formed from turbidity currents produced when volcanoclastic products
 347 of the phreatomagmatic eruption entered the lake. This is comparable with
 348 the sediment-waves observed downslope of delta fronts in lakes, formed from
 349 turbid rivers entering the lake (e.g., Urgeles et al., 2011).

350 A 20 to 40° steep sloped wall delimits the northern end of Lake Kivu's
 351 Main Basin, and abuts the phreatomagmatic cone of Mount Goma ('D' on
 352 Figure 4a ii). This wall corresponds to a superposition of lava from Nyi-
 353 ragongo, whose lava flows frequently extend into Lake Kivu including flows
 354 from the 2002 eruption (Figure 3). Subaquatic penetration of lava flows along
 355 steeper slopes is expected to break up into debris flows, which cascade down
 356 the slopes without extending the land, or collapse the sediment-laden slopes
 357 (Morgan et al., 2003; Sansone and Smith, 2006; Mitchell et al., 2008). It is
 358 plausible that the lava wall above 250 m depth was created by lava pouring
 359 into the lake from nearby volcanic eruptions during the period of lake-level

rise, and eventually formed the dam blocking the flow to the north. From a depth of 250 to 315 m, the lake-bottom slope becomes more gentle and irregular and likely consists of the superposition of debris-flow deposits from the lava wall 'D' and lava flows from the eruption of subaquatic volcanic features in the area (e.g. 'B' and 'C', Figure 4a ii).

At the foot of Mount Goma a dendritic shaped relief occurs between 395 and 425 m contour ('B' on Figure 4a ii). The fractal outlines of this relief imply its formation as an underwater lava flow, similar to some of the lava flows observed offshore of Pico Island, Azores (Mitchell et al., 2008). Reduced driving-stress is imminent upon lava flowing into water, due to increased buoyancy and cooling-induced viscosity. Furthermore, viscosity, local topography, and flow thickness determine whether the lava will penetrate deeper into the lake or form a delta (Mitchell et al., 2008). There is no association of the relief observed in Figure 4a ii with a lava-delta upslope, suggesting that the flow originated at its location. With respect to the low gradient of the lake bottom directly upslope of the lava, it is also possible that the effusion of lava into the lake occurred when the lake-level was at the ~ 390 m contour. However, the seismic profile of the lava flow in Figure 5 S4 and additional seismic profiles that directly cross-cut the lava flow (seismic profiles not shown), indicate a lack of sediment cover. This could suggest a relatively recent age of the flow, or is potentially related to poor seismic penetration due to the hummocky surface of the flow.

Isolated volcanic cones are observed in Figure 4a ii and are aligned NE-SW with two tuff rings observed in the southeast corner of Frame i. These features probably materialized along a major fault lineament that extends the lake basin from its northern limit to the Idjwi Horst in the south. The cone located ~ 1 km south of the shoreline ('C' on Figure 4a ii) appears constructional, and is reminiscent of subaquatic volcanic cones observed in the ocean that are formed under the confining pressure of a water column (e.g., Clague et al., 2000). However, the 'eroded cones' located ~ 2 km to the SW (including 'E' on Figure 4a ii) appear to have a deteriorated topography, which could indicate their subaerial formation.

4.1.3. Frame iii

The Main Basin of Lake Kivu is separated from the basin viewed in Figure 4a iii by a ~ 500 m wide, submerged channel. This channel is delimited in the west by the Precambrian rock relief of the Mbuji Peninsula, which extends into the lake to a depth of 125 to 140 m. The basin has the same

397 maximum depth as Kabuno Bay, ~ 140 m, that is located directly to the
 398 northwest (Figure 1) and separated from the Main Basin by a saddle at 12 m
 399 depth. Furthermore, the basin in Frame iii is filled to a depth outlined by the
 400 126 m contour with warm, salty, and Fe^{2+} -rich water, which is identical to
 401 that below the main density gradient in Kabuno Bay. This suggests that the
 402 two basins have a hydrogeologic connection, or are fed by the same ground-
 403 water source (Ross et al., 2014). A $\text{N}26^\circ\text{E}$ oriented depression is located
 404 in the northern part of the frame, and is bordered by a ~ 1 km NNE-SSW
 405 escarpment ('F' on Figure 4a iii). The origin of this feature is not known,
 406 however, it could be where Fe^{2+} -rich groundwater is discharging into the
 407 basin. Distinct depression features have commonly been observed as areas
 408 of focused groundwater flow in other lakes and in the ocean (Whiticar, 2002;
 409 Bussmann, 2011).

410 A pronounced hummocky relief dominates this basin, such as that out-
 411 lined to the east of the depression feature. This relief is probably linked to
 412 lava flows from the Nyamulagira volcano. Lava flows from the 1912 eruption
 413 of Nyamulagira are responsible for creating the narrow channel and saddle
 414 that separates the Main Basin from Kabuno Bay (Pouclet, 1976).

415 The southwest part of the bathymetry in Frame iii reveals a $\text{N}33^\circ\text{E}$ ori-
 416 ented trench calculated from the slope map, which is characterised in its
 417 southwestern end by an escarpment oriented towards the SE ('trench' on
 418 Figure 4a iii). This feature is interpreted to be an open fracture or fault with
 419 an orientation similar to fault strikes encountered in this part of the rift (e.g.,
 420 Ebinger, 1989; Yamba and Boven, 1998; Corti et al., 2007), and could also
 421 be a conduit for groundwater discharge.

422 *4.2. Region 2*

423 The Sebeya River meanders through Gisenyi, Rwanda, and is the largest
 424 river feeding into Lake Kivu and discharges at its northeastern corner to form
 425 a delta ('G' on Figure 4b). The catchment of the basin consists of volcanic
 426 and schist terrain, from where rainfall events can induce highly turbid, dense,
 427 hyperpycnal flows (Muvundja et al., 2009). At the front of the river mouth, a
 428 delta fan and subaquatic channels are observed in the bathymetry in Region
 429 2, Figure 4b. A step in the delta slope occurs at the 250 m depth, which
 430 likely represents a former shoreline suggested by the pronounced 'submerged
 431 channel'. This step could also represent a scar from a collapse of the delta
 432 front. However, an associated avalanche deposit is not clearly detectable
 433 downslope.

434 South of the river, along the eastern shoreline, numerous submerged
 435 paleo-valleys sculpted in the Precambrian basement occur. A seismic profile
 436 reveals several mass-flow deposits at a depth ~ 450 m, which are identified by
 437 their semi-circular lobe structures and transparent seismic facies that pinches
 438 out downslope (Figure 6 S5). The shallowest deposit is located ~ 4 m below
 439 the modern lake floor and outlined in Figure 6 S5. Mass flows are potentially
 440 triggered by ground shaking during seismic activity, which are common in
 441 the Kivu Basin (e.g., d'Oreye et al., 2011).

442 *4.3. Region 3*

443 The escarpment extending from the Idjwi Island horst structure is ob-
 444 served at the southern limit of Region 3, Figure 4c at a depth of ~ 460
 445 m. Lake Kivu sediments are thickest in the vicinity of the horst along the
 446 eastern basin, which probably represents the oldest part of the lake (Wood,
 447 2014). At the northern end of the escarpment, a seismic profile reveals an in-
 448 sequence normal fault with a downfaulted eastern part that occurs ~ 100 m
 449 east of the escarpment (Figure 6 S6). Additional normal faults are observed
 450 in the seismic profiles along a NE-SW alignment, which likely propagates to
 451 the northern extent of the lake east of the phreatomagmatic cone of Mount
 452 Goma. This is coherent with the lineament of subaquatic volcanic features
 453 observed in Figure 4b ii that extends northwards to the western limit of
 454 Mount Goma. It is possible that these faults are the high-velocity conduits
 455 from where cold groundwater discharges into the lake (Figure 3). Cold and
 456 dilute groundwater discharge, in addition to hydrothermal discharge, is nec-
 457 essary to maintain the temperature and salinity gradients observed in the
 458 deepwater of Lake Kivu (Schmid et al., 2012; Ross et al., 2012).

459 **5. Lake-level and event reconstruction**

460 Recent seismic surveys in the eastern part of the Main Basin of Lake
 461 Kivu (Wood, 2014), suggest that the lake could be as old as the Miocene.
 462 However, the oldest date obtained from the sediment record up to now is
 463 ~ 13.5 ka (Hecky and Degens, 1973). This date was acquired 40 years ago by
 464 radiocarbon analysis of bulk organic matter, which was collected from the
 465 bottom of an 8 m core taken at a depth of ~ 425 m in the Main Basin (Degens
 466 et al., 1973). An additional core collected by the same team revealed coarse
 467 rounded pebbles at a depth of ~ 315 m, which implied a lake-level lowstand
 468 with an inferred age of 12-14 ka. These results were consistent with seismic

469 surveys conducted by Wong and von Herzen (1974), which revealed a thin
470 veneer of sediment covering depths of less than 300 m in the southern part
471 of the lake. The authors concluded that prior to the lake-level rise and
472 subsequent draining via the Ruzizi River ~ 10 ka the lake-level lowstand was
473 ~ 300 m, and that the lake was hydrologically closed so that the lake level
474 fluctuated depending on rates of precipitation and evaporation.

475 The high-resolution bathymetric survey and seismic profiles presented
476 here suggest the presence of at least four paleoshorelines at 425, 395, 315,
477 and 250 m depth. A timeline for the transgression of these shorelines and
478 the relative dates of volcanic activity in Lake Kivu is outlined in this section.
479 The timeline is inferred from our data in addition to the published analysis
480 of the cores gathered by this investigation and previous research conducted
481 on Lake Kivu and Lake Tanganyika, which is summarized in Table 1.

482 5.1. *Paleoshorelines*

483 The paleoshoreline at 425 m depth outlines the isolated volcanic features
484 and the northward extension of the Idjwi Island horst, and is the lower limit
485 of the submerged erosional valleys in the east (Figure 2b). From the analysis
486 of the high-resolution bathymetric map, the morphology at depths greater
487 than 425 m is dominated by enhanced sediment deposition (Figure 2b). This
488 implies an older age of the basin below 425 m depth where prolonged sed-
489 iment deposition appears to even out the morphology. Furthermore, the
490 seismic penetration into the sediment is largest at depths greater than 425
491 m implying a thicker sedimentary basin below this depth (e.g. Figure 6 S6).
492 The eroded river valleys in the east and the volcanic platform in the west are
493 clearly outlined by the 395 m contour. The morphology appears relatively
494 smooth between the 425 and 395 m contour when compared to shallower
495 water depths, which suggests a more substantial sediment cover below the
496 395 m depth. From this assessment, we infer that the 425 m paleoshoreline
497 represents the lake level prior to the volcanism in this region initiated at
498 100 ka Rogers et al. (1998). At this time volcanism in the western VVP
499 was active and consequently began to form the dam linked to the lake-level
500 rise. The lake level prior to 100 ka could have fluctuated greatly. However,
501 lake-level rise would have been limited by the height of the lava dam in the
502 north. Following the late Pleistocene and before the active initiation of the
503 Nyiragongo volcanism at ~ 15 ka, the lake level is assumed to have risen to
504 the 395 m paleoshoreline. A lake-level rise from the 425 to 395 m depth is
505 documented by shoreline terraces observed in Figure 6 S1 and S2.

506 We postulate that the 425 m paleoshoreline represents the late Pleistocene
 507 lowstand that was suggested to have been at 300 m by Hecky and Degens
 508 (1973). As a consequence of the active volcanism ~ 15 ka, the lake-level rise
 509 is expected to have become more rapid with the increasing dam elevation in
 510 the north. The dam formation in conjunction with the onset of the Holocene,
 511 during which wetter climates prevailed, rapidly led to the eventual overflow
 512 via the Ruzizi River ~ 10 ka. The inundation of the active volcanoes and the
 513 onset of the deep lake marked the beginning of the hydrothermal activity
 514 and subsequent thermohaline stratification in Lake Kivu (Table 1).

515 5.2. *Volcanism*

516 Two additional paleoshorelines are postulated at the 315 and 250 m con-
 517 tours. These depths outline the base of tuff rings and maars in the western
 518 part of the basin. The sedimentation recorded within 4 tuff rings that are
 519 formed from shallow-water eruptions implies that these features culminated
 520 during the period of lake-level rise ~ 9 -14 ka (Figure 4b). The 250 m con-
 521 tour also outlines the steep lava wall delimiting the Main Basin to the north,
 522 further signifying volcanic activity in connection with a rapid lake-level rise.
 523 Additional evidence of volcanic activity at ~ 10 ka was observed in a 3.2 m
 524 long core provided by Hecky and Degens (1973), which terminated in a nearly
 525 2 m thick ash layer. The core was taken in proximity to the volcanic plat-
 526 form, at a point southeast of the isolated tuff cone observed in Figure 4a i.
 527 This ash layer likely coincides with the formation of the maars and tuff rings
 528 at the 315 or 250 m depth, where shallow-water eruptions typically disperse
 529 considerable quantities of ash (Zimanowski and Büttner, 2003). Elemental
 530 analysis conducted by Degens and Kulbicki (1973) and Haberyan and Hecky
 531 (1987) suggested that intense hydrothermal activity inside Lake Kivu was
 532 initiated abruptly at ~ 5 ka (or at 138 cm core depth). Furthermore, Felton
 533 et al. (2007) observed an increase in all elements in the sediment of Lake Tan-
 534 ganyika at ~ 4 ka BP to the top of the core. Although they attributed this
 535 rise in elements to renewed overflow at the Ruzizi outlet (Table 1), it could be
 536 correlated with increased hydrothermal activity in Lake Kivu. Similar char-
 537 acteristics of increased volcanic activity were observed again by Haberyan
 538 and Hecky (1987) at ~ 1 ka (32 cm core depth), and constrained to ~ 0.8 ka
 539 by the present investigation (Ross et al., 2013). Lower intensity volcanism,
 540 or hydrothermal pulses, were recorded in the sediment at ~ 0.6 and 0.4 ka
 541 and could represent the most recent subaquatic volcanic activity determined
 542 in Lake Kivu (Ross et al., 2013).

543 Observations of subaquatic volcanic activity recorded in the sediment of
 544 Lake Kivu imply that the subaquatic volcanic features observed in the high-
 545 resolution bathymetry, including the volcanic dome in Figure 4a i ('A'), and
 546 the subaquatic volcanic cone and the apparent lava flow in Figure 4a ii ('C'
 547 and 'B'), were formed sometime after the onset of the Holocene but were not
 548 significantly active prior to ~ 4 -5 ka. Furthermore, we can postulate from the
 549 morphological analysis and seismic profiles that the phreatomagmatic cone
 550 Mount Goma has been active following the lake-level rise at the beginning
 551 of the Holocene, but is likely much older than the lava flow that crosscuts
 552 its pyroclastic sediment waves ('B' on Figure 4a). Hydrothermal activity
 553 is a typical indicator of active subaquatic volcanism (Ohsawa et al., 2002;
 554 Morgan et al., 2003; Clague, 2009) and is concentrated in Lake Kivu at the
 555 volcanic platform in the west in Figure 3; within the basin in Figure 4a iii,
 556 and at the dome feature observed in Figure 4a i. Hydrothermal activity
 557 following the lake-level rise in the early Holocene, and still observed in the
 558 Main Basin today, maintains the current state of meromixis in Lake Kivu
 559 (Schmid et al., 2005). Not surprisingly, hydrothermal activity appears to be
 560 linked to geological activity in Lake Kivu, and intense hydrothermal events
 561 are correlated to volcanism and lake-mixing events (Haberyan and Hecky,
 562 1987; Ross et al., 2013, 2012).

563 **6. Conclusion on threat of a lake overturn**

564 It is generally presumed that volcanic and seismic activity occurring
 565 within Lake Kivu would lead to an overturn of the lake and subsequent
 566 expulsion of its dissolved gases. The marked volcanic events at ~ 5 and 0.8
 567 ka (Table 1) are linked to a lake-mixing event such as a limnic eruption. In
 568 particular, drastic changes in the flora and fauna were recorded immediately
 569 following the event at ~ 5 ka (Haberyan and Hecky, 1987) that indicated sig-
 570 nificant geochemical changes in the water column. Furthermore, a distinct
 571 subaquatic-tephra layer and a change in deepwater carbonate saturation are
 572 recored in the sediments at ~ 0.8 ka (Ross et al., 2013). An indication of
 573 toxic deepwater having been mixed into surface water is inferred from the
 574 genetically much younger cichlid fish in Lake Kivu than those in lake Edward
 575 and Victoria (Bezault et al., 2011), despite the implied Miocene age of Lake
 576 Kivu. However, there is still no conclusive evidence to link geologic features
 577 observed on the lake floor to an occurrence of a lake overturn triggered by
 578 volcanism or rift tectonics. Preliminary estimates of the effect of a plume

579 formed by a subaquatic magmatic intrusion suggested that such an event
 580 would not be sufficient to trigger a catastrophic gas release at the present
 581 stratification and gas concentrations (Schmid et al., 2003), but a thorough
 582 hazard assessment still needs to be conducted. The high-resolution bathy-
 583 metric mapping of the Main Basin of Lake Kivu reveals an unprecedented
 584 look at geologic features in this area, including volcanic features that might
 585 have been active at least within the last ~ 1 kyrs. The young age of these
 586 features in addition to the 2002 volcano-tectonic eruption of Nyiragongo,
 587 and its probable dyke intrusion beneath the lake, suggest that a sublacus-
 588 trine eruption in Lake Kivu is a possible threat. A brief summary of the
 589 volcanism associated with the dynamics of Lake Kivu is as follows:

- 590 **i** late Pleistocene - active volcanism in the western VVP
- 591 **ii** 10 ka - damming of Lake Kivu by active volcanism, leading to overflow of
 592 the lake at the Ruzizi River to Lake Tanganyika
- 593 **iii** 5 ka - intense deep hydrothermal activity linked to volcanism in Lake
 594 Kivu, and possibly a lake overturn
- 595 **iv** 1-0.8 ka - intense hydrothermal activity linked to volcanism in Lake Kivu,
 596 and breakdown of thermohaline stratification
- 597 **v** 0.6-0.4 ka - moderate hydrothermal activity linked to perturbation of ther-
 598 mohaline stratification
- 599 **vi** 10 years ago - flank eruption of Nyiragongo linked to a possible dyke that
 600 propagates beneath Lake Kivu

601 Earthquakes have also been documented within Lake Kivu, where their
 602 occurrence could lead to mass-movements in the sublacustrine basin, includ-
 603 ing that observed in Figure 6 S5. Such mass-movements, if substantially
 604 large, could cause tsunami waves and seiches (e.g., Schnellmann et al., 2002)
 605 that eventually could trigger the expulsion of gases from the lake. Further-
 606 more, the increasing concentrations of the dissolved gases and the possibility
 607 of their inherent combined partial pressures overcoming the hydrostatic pres-
 608 sures of the lake is a commonly referenced threat (e.g., Tietze, 1991, 1992;
 609 Schmid et al., 2003). The present increase of the gases in the water column is
 610 linked to an increase in the groundwater discharge into the lake (Schmid et al.,
 611 2005, 2012). Since hydrothermal groundwater discharge has a causative effect

on the thermohaline stratification, and is a product of volcanism in the lake, then monitoring changes in hydrothermal activity could yield early warning signs of a gas outburst. Hydrothermal activity concentrated along volcanic features in the sublacustrine environment can be monitored with CTD profiling, or moored sensors (Baker et al., 2012). Additionally, the bathymetric map presented here can be used to monitor developments in the Kivu sublacustrine basin, where changes to volcanic features over time can be observed by modern acoustic techniques (e.g., Caress et al., 2012; Chadwick Jr et al., 2012; Watts et al., 2012). In conclusion, the new high-resolution geomorphologic dataset will greatly aid in the assessment of a potentially catastrophic eruption. Additional studies on the subaquatic volcanoes observed in this map should focus on the possible activity of the young geomorphological features, including features 'A' and 'B' in Figure 4a.

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Table 1: Inferred lake-level changes and hydrothermal activity over the past 14 kyrs

Events	*Date (ka)	lake-level (m)
Nyiragongo and Nyamulagira become active and dam the outflow to Lake Edward ^{1,2,3,4,5,6}	14 - 10	-300 or lower
Initiation of hydrothermal activity and stratification linked to overflow at Ruzizi to Lake Tanganyika ^{2,4,5,8}	10 - 8	present to +100
Possible brief cessation in overflow and breakdown of stratification in period ^{2,6}	8 - 6	present to -100
Sudden change in lake chemistry attributed to extreme hydrothermal event - followed by initiation of thermohaline stratification ^{1,2,4,6}	5	near present
Extreme hydrothermal event - breakdown of stratification ^{1,2,3,9}	1 - 0.8	near present
Renewed thermohaline stratification, lake chemistry progressively reaches its current state ^{1,2,4,9}	1 - 0	present

* Dates are rounded to the nearest thousand and are based upon radiocarbon analysis of sediment, with the exception of those derived from Ross et al. (2013). A range implies a duration, or more than one date calculated for the same event.

¹ Degens et al., 1973

² Hecky and Degens, 1973

³ Stoffers and Hecky, 1978

⁴ Degens and Kulbicki, 1973

⁵ Haberyan and Hecky, 1987

⁶ Felton et al., 2007

⁷ Wong and von Herzen, 1974

⁸ Olson and Broecker, 1959

⁹ Ross et al., 2013

Figure 1: Overview map of Lake Kivu's basin. The inset represents the location of the Kivu Basin within Africa. Main figure: Gray scale, 30 m ASTER Global DEM (GDEM) data displaying Nyiragongo and Nyamulagira volcanoes north of the lake (ASTER GDEM is a product of METI and NASA). Yellow discontinuous line is the border separating the Democratic Republic of Congo to the west and Republic of Rwanda to the east. Black contour lines represent the 200 and 400 m depths in the lake. Coloured map overlaying the lake represents the 224 km² area of high-resolution bathymetry data acquired within the Main Basin, and used for analysis of observed features.

(a) The 10 m resolution bathymetric map sectioned into 3 regions (1, 2 and 3); used for morphological analysis. Tracklines from the bathymetric survey are delineated in red.

(b) The interpretive map displaying the general morphological features including hummocky volcanic relief, Precambrian rock structures, drowned river valleys, enhanced sediment deposition, and the paleoshorelines represented by depth contours; 425, 395, 315, and 250 m. Satellite cones, eruptive fissures and the lava flows from the 2002 eruption of Nyiragongo are marked on the subaerial terrain. The escarpment of the Idjwi Island horst is outlined on the southeastern part of the map.

Figure 2: High resolution bathymetric map

Figure 3: Region 1: Volcanic features including rootless cones (outlined by the polygon 'concentrated cones'), tuff rings and maars, hydrothermal and cold groundwater discharge locations. The subaerial features include the satellite cones (the phreatomagmatic cone 'Mount Goma' is named), eruptive fissures, and the lava flows from the 2002 Nyiragongo eruption. The figure is sectioned into 3 frames (i, ii, and iii) for further morphologic and structural analysis. Three subaquatic features observed here are marked on the map in Figure 4a: dome 'A', lava flow 'B', and volcanic cone 'C'.

(a) Frames i, ii, and iii from Region 1 in Figure 3 at 5 m resolution. Features from the morphologic and structural analysis are marked on the map and annotated in the legend. The 250, 315, 395 and 425 m contours are outlined on each map. The seismic profiles S1 through S6 are depicted in Figure 5. Frame i: The volcanic platform. Frame ii: The submerged flank of Mount Goma with 'sediment waves' along its western margin. Frame iii: The isolated basin connected to the main basin of the lake. The 126 m contour outlines the height to where warm groundwater fills the basin. The north arrow is found in Frame iii, and the horizontal and water depth scales are located on each frame.

(b) Region 2: The drowned river valleys and delta 'G'.

(c) Region 3: Idjwi Island horst structure, outlined by the 'escarpment'.

Figure 4: Region 2 and 3. The north arrow is found in Region 2, and the horizontal and water depth scales are located on each map. The 250, 315, 395 and 425 m contours are outlined on the maps. The seismic profiles S5 and S6 are depicted in Figure 6.

Figure 5: Seismic profiles; S1, S2, S3 and S4, located in Frames ii and iii. The vertical and horizontal scales (in meters) are located in the lower corner of each profile. Annotations are explained within the text.

Figure 6: Seismic profiles: S5 and S6, located in Region 2 and 3. The vertical and horizontal scales (in meters) are located in the lower corner of each profile. Annotations are explained within the text. An upper and lower mass-movement observed in the seismic profile S5 are outlined and shaded in green.