1 **Slope stability models for rainfall-induced lahars during long-lasting eruptions**

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11

12 Abstract

13 In this study we analyse the spatial distribution of potential lahar sources during long-lasting eruptions

14 using two slope stability models (SHALSTAB and TRIGRS). The analysis is based on observed deposit

15 grain-size and thickness, rainfall data and slope angle of the lahar events, in the area surrounding the

16 Cordón Caulle volcano, Chile, that occurred during the 2011 eruption. The main phase of the eruption

17 (4-7 June) was characterized by eruptive plumes from 7 to 11 km high which dispersed most of the

18 tephra eastward toward Argentina, with a total estimated volume of about 1 km^3. Tephra fall blanketed

19 the region from ESE to ENE of the volcano with the thickness of the tephra layers between

20 approximately 1 m (15 km from vent) and 0.06 m (240 km from the vent). On 10 June 2011, a major

21 lahar occurred close to the Argentina-Chile border that reached the National road 231 (28 km from the

22 vent). Three field campaigns were undertaken to collect samples and data from tephra deposits and

23 triggering mechanisms in the lahar source area. Model input parameters were obtained from

24 geotechnical test and field measurements. Several metrics are used for model evaluation and best fit to

25 the data are obtained for simulations considering non-cohesive ash layers for SHALSTAB and a

26 cohesion of 0.5 kPa for TRIGRS. Both models are sensitive to the physical properties of the tephra

27 deposit and the hydraulic and materiel strength properties of the study area. They both also show good

28 agreement with field data but provide different information: TRIGRS provides an estimate of the timing

29 (based on a storm event) and location of a potentially unstable area, while SHALSTAB simulations

30 result in landslide susceptibility classes based on critical rainfall value. These outcomes provide

31 fundamental insights into lahar triggering during long-lasting volcanic eruptions and are crucial to the

32 compilation of lahar hazard maps and emergency management plans in the South Andes volcanic region.

33

34 Keywords: lahar, initiation mechanism, SHALSTAB, TRIGRS, Cordón Caulle volcano

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37 **1. Introduction**

1

38 Lahars are widespread phenomena in volcanic areas that can severely affect people and infrastructure
39 (Pierson et al., 1990; Pierson et al., 1992; Janda et al., 1996; Scott et al., 2005; Scott et al., 1996; Lavigne
40 et al., 2000a). Historical cases of deadly lahars such as Mt Pinatubo, Philippines (Pierson et al., 1992),
41 Nevado del Ruiz, Colombia (Pierson et al., 1990), Casita Volcano, Nicaragua (Scott et al., 2005), and
42 Panabaj, Guatemala (Charbonnier et al., 2018), demonstrate that they represent one of the most
43 significant hazards to people living in volcanic areas. The lahars generated years after the 1991 eruption
44 of Mt. Pinatubo (Janda et al., 1996) constitute the most voluminous and devastatingly widespread, as
45 well as one of the most long-lasting lahar episodes, ever recorded.

46 "Lahar" is an Indonesian term for a mixture of rock debris and water that flows downslope, beyond
47 normal streamflows (Smith, 1986; Smith and Lowe, 1991; Vallance, 2000). The term is genetic rather
48 than descriptive, encompassing a wide spectrum of sediment to water ratios and flow rheologies
49 (Manville et al., 2009). Lahars can be caused directly by eruptive activity or during post-eruptive or
50 quiescent periods. They are mainly generated by the remobilization of tephra-fallout and Pyroclastic
51 Density Current (PDC) deposits, but may also form from a previously emplaced debris avalanche
52 produced by the gravitational collapse and disintegration of the volcano edifice (Pierson et al. 2014).
53 After explosive volcanic eruptions, PDC deposits fill valleys and tephra-fallout deposits may blanket
54 the volcano flanks and, when the intensity of the explosive eruption is high enough, topographic reliefs
55 even hundreds of kilometres from the source.

56 Tephra-fallout deposits are unconsolidated, loose and highly erodible during rainstorms (Pierson and
57 Major, 2014). It is well documented that explosive eruptions can drastically alter drainage basin
58 hydrology and erosion processes (Pierson and Major, 2014). Many researchers have also reported poor
59 infiltration capacity of freshly deposited tephra following volcanic explosions, producing surface flow
60 over barren hillslopes (Collins and Dunne 1986; Leavesly et al. 1989; Major et al., 2000; Pierson et al.,
61 2013; Major and Yamakoshi, 2005; Yamakoshi and Suwa, 2000; Yamamoto 1984).

62 Most of the available lahar-hazard assessments are based on the analysis of lahar spreading areas
63 (Manville et al., 2013). Models routinely used include the empirical model LAHARZ (Iverson et al.,
64 1998) and the more complex Titan 2D model (Sheridan et al., 2005). Numerical models based on depth-
65 averaged continuum flow equations are also used for lahar hazard evaluation, such as FLO-2D (O´Brian
66 et al., 1993) at Nevado del Huila, Colombia (Worni et al., 2012), Popocatépetl, Mexico (Caballero and
67 Capra, 2014; Caballero et al. 2016) and Panabaj, Guatemala (Charbonnier et al., 2018). Haddad et al.
68 (2010) also used a Smoothed Particle Hydrodynamics (SPH) depth-integrated model for simulating
69 lahar spreading on Popocatépetl volcano. Besides, analysis of rainfall intensity and duration thresholds
70 for lahar triggering mechanisms have been performed at many volcanoes, e.g. Mount Pinatubo (Van
71 Westen and Daag, 2005), Indonesia (Lavigne et al., 2000a; Lavigne et al., 2000b; Lavigne and Suwa,
72 2004), Mexico (Capra et al., 2010), and Montserrat (Barclay et al., 2007).

73 Predicting the initiation area is crucial to assessments of rainfall triggered lahar volumes and potential
74 runout (Iverson, 1997). Rain-triggered lahars are associated with two main mechanisms: sheet and rill

erosion by Hortonian overland flow due to deposit saturation (Collins and Dunne, 1986; Cuomo et al., 2015) and rain infiltration that can induce shallow landslides (Iverson and Lahusen, 1989; Manville et al., 2000; Crosta and Dal Negro, 2003; Zanchetta et al., 2004;Volentik et al., 2009; Cascini et al., 2010). Spatially distributed simulations of unsaturated tephra-fallout deposit failure is usually performed using hydrological models coupled with infinite slope stability. Such models have been developed following the catastrophic landslides and flow events in Sarno, 1998 and similar areas in the Campania region in South Italy (Cascini et al., 2010; Frattini et al., 2004; Crosta and Dal Negro, 2003; Cuomo and Iervolino, 2016; Cuomo and Della Sala, 2016; Cascini et al., 2011; Sorbino et al. 2007, 2010). In addition, Volentik et al. (2009) proposed an infinite slope failure criterion (Iverson, 2000) for the evaluation of lahar source areas at Mt Natib volcano, Indonesia, and Mead et al. (2016) presented a study on rain triggered lahar susceptibility using shallow landslide and erosion models.

Nevertheless, even if the main parameters for lahar triggering have been identified, how failure mechanisms vary during long lasting eruptions, i.e. those lasting from a few days to several months or years, remains an open question. In fact, the effect of the variation of parameters, such as the thickness, grainsize and composition of tephra layers and of the rainfall intensity and frequency over long periods of time, on slope stability is difficult to model accurately. For this reason, analysis of well-documented case studies is crucial for understanding such complex process, and for empirical assessment of the above-mentioned thresholds.

In this paper, we present a detailed study of events that occurred during the 1 year eruption of Cordón Caulle (Chile) in 2011-2012, and we have analyse factors that may influence the triggering of rainfall induced lahars in such a variable and complex system. We take into account several different parameters, including the extension of the source areas, amount of water, slope gradient, and the physical characteristics and thickness of the source deposit. Two watersheds with recorded lahar events, located 50 km from the Cordón Caulle vent area, have been studied in the field and used for back analysis using two shallow stability models, SHALSTAB and TRIGRS. In particular, variations in the number of layers, layer thickness and composition, and rainfall intensity and duration were considered in the numerical simulations.

This combination of detailed field and laboratory studies, and back analysis of well constrained events with two distinct numerical models has thus contributed in improving our understanding of the triggering processes of these extremely dangerous phenomena and has helped us improve the related hazard assessment in the variable and complex environment provided by long-lasting volcanic eruptions.

2. The study area

2.1 The 2011 eruption of Cordón Caulle

The case study includes several lahar-prone basins near the Argentina-Chile border and in Villa La Angostura, Argentina. The area is in the Southern Andes, east of the Cordón Caulle volcano, along the

main ash dispersal axis and road N. 231 and between 28 and 50 km from the vent area (Fig. 1). Here, the southern Andes have an average elevation of about 1500 m. The landscape is characterized by glacially sculpted steep slopes, glacial and tectonic lakes, and flat valley floors of fluvial and glacio-fluvial outwash. During the Pleistocene a lobe of the Cordilleran ice sheet occupied the study area. U-shaped valleys, many cirques, moraine and erosive glacial features (roches moutonnées) are evidence of the Pleistocene glaciation. Most of the post-glaciation erosive surfaces are covered by Holocene tephra sequences with ages ranging from 10 to 1.1 ka (Singer et al., 2008; Naranjo et al., 2017). Naranjo et al. (2017) described a Holocene tephra succession, near Villa La Angostura, composed mainly of lapilli pumice, lapilli scoria, ash and paleo-soils with a total thickness of 4.6 m. The old basement consists of Cretaceous volcanic rocks (diorites, granodiorites) and outcrops on the summit of Cerro Bayo. The whole area is densely vegetated, covered by the Valdivian rain forest composed mainly of trees from the Nothofagus genus (lengas and ñires). The climate is cold with an average temperature of 10° C and annual precipitation varying between 800 and 2500 mm.

On 4 June 2011, the Puyehue-Cordón Caulle volcanic complex (Central Andes, Fig.1) erupted from multiple vents along a NW-SE fracture in the Cordón Caulle system (CC) system (Castro et al., 2013; Pistolesi et al., 2015). The main phase (4-7 June), characterized by plumes heights of 7 to 11 km, dispersed about 1 km^3 of tephra towards Argentina and deposited between 1 m (15 km from vent) and 6 cm (240 km from the vent) of tephra on the ground (Fig. 1). The eruption continued for several months with lava effusion and low intensity ash emissions. A red alert was lowered to orange in March 2012, 10 months after the main phase (Elissondo et al., 2016).

The eruptive sequence was divided into 3 main tephra units: I, II, and III (Pistolesi et al., 2015). At outcrops 25 to 50 km from the vent, the sequences appear as two, stratified, fine lapilli bearing depositional units (I and II), covered by a third unit composed of multiple fine ash layers interbedded with thin coarse ash to fine lapilli beds (III) (Fig. 2). Unit I was emplaced during the first 24-30 hours of the eruption (4 and 5 June) and unit II was emplaced during 5-7 June. Finally, unit III was mostly emplaced from 7 until 15 June, although millimetres of fine tephra also sedimented during later low level activity.

Tephra fallout produced several impacts on population, animals and vegetation (Elissondo et al., 2016). After the main explosive phase (4-7 June) a large volume of tephra mantled the steep hillslopes of the U-shaped valley in the surrounding area of the volcano (Fig. 1a).

2.2 Lahars, floods and snow avalanche events following tephra-fallout deposition (2011-2012)

In this section we describe lahar, flood and snow avalanche events, including both data obtained from field campaigns, during and after the eruption, and reports, photos and oral communications made during the volcanic eruption by the Villa La Angostura Civil Defence and Mountain rescue group.

During the wet season, loose and unconsolidated tephra was transported through steep channels, and

149　lahars reached the valley bottom. Many lahars and related flood events were observed between the

150　Argentina-Chile border and Villa la Angostura (48 km from the vent). The tephra-fallout thickness, as

151　measured on 10 June was between 30 and 40 cm near the Argentina-Chile border and 12 and 15 cm at

152　Villa La Angostura (Pistolesi et al., 2015). The eruption and subsequent tephra-fallout coincided with

153　the beginning of winter. In June, precipitation was falling as rain at 800 m, and as snow at 1,500 m.

154　Total precipitation accumulated between 7 and 10 June was 116 cm. Figure 3a and b illustrate the daily

155　rainfall and the timing of lahar and floods events near both the Chile-Argentina Border (Rincon rain

156　gauge) and Villa La Angostura between June 2011 and June 2012.

157　The most destructive lahar event occurred on 10 June 2011, during the first week after the eruption

158　onset, flowing down a channel on Cerro Las Tres Hermanas (28 km form the vent area, Fig. 1b). The

159　volcanic material coming down the creek blocked the culvert below the road and flowed on to road N.

160　231, eroding and destroying the asphalt (Red triangle "L2" on Fig. 1b and Fig. 4a). The flow cross

161　section upstream of the road was estimated at 80 m². One hundred meters upstream of road N. 231, we

162　identified 3 main lobes of lahar deposits crossing the forest. A stratigraphic section, of thickness 1.39

163　m, from an exposure in the main creek revealed 5 beds with 2 different lithofacies. Figure 5a shows the

164　stratigraphy of the lahar deposit with 2 different lithofacies, types A and B. Type A is an inversely

165　graded, massive, pumice rich, coarse sand to fine gravel, supported in a fine sand silty matrix. Type B

166　is massive, medium to fine black sand.

167　One kilometre from the first creek, "Diana waterfalls" creek crosses road N. 231 under a bridge (red

168　triangle L1 in Fig 1b). A lahar deposit is preserved at the foot of the waterfall, located 100 m upstream

169　from the road (Fig. 3b). In outcrop, the material appears to be remobilized Cordón Caulle 2011 tephra.

170　Figure 5b shows a stratigraphic section with a total thickness of 143 cm and 7 distinct beds of 3 different

171　lithofacies: C, D and E. Type C is a massive, pumice rich, coarse to fine gravel, clast supported deposit,

172　with some coarse sand. Type D is a massive, pumice rich, coarse sand with little fine gravel. Type E is

173　a massive, pumice rich, coarse sand to fine gravel, supported in a fine sand silty matrix.

174　 Both lahar deposits from "Diana waterfalls" and "Road 231" are characteristic of hyperconcentrated

175　flows (Pierson, 2005) which, in our case, is a friable deposit with very little silt and clay. The lack of

176　large boulders in the deposit is another characteristic of hyperconcentrated flows.

177　 Mud marks found on the top of the bridge (2m above the river bed) suggest the lahar reached the Blanco

178　River through the bridge.

179　At the same time lahars occurred in the Totoral basin, located approximately 7 km northeast of "Diana

180　waterfalls" creek (Fig. 1b). The lahars starting in channels on the valley slopes reached the main river,

181　causing bank erosion and overbank flows. Flooding and sediment deposition in the Totoral River valley

182　damaged houses in Paráje Rincón and destroyed trees (Fig. 4c and d).

183　On 10 June, floods were reported in Villa La Angostura. Remobilization of tephra-fallout on the channel

184　and hillslopes caused flooding and tephra accumulation in Villa La Angostura, which is situated on the

185　alluvial fan (Fig.1b).

At the end of the winter, on 19 September, a snow avalanche occurred in the Florencia river basin. The avalanche, of volume ~ 46000 m^3, started at the Florencia creek ridge crest and flowed approximately 1 km before stopping in the main river channel at 1200 m (Fauqué et al., 2011). Earlier observations in the upper catchment had found a weakness between the lapilli bearing deposit (unit I and II) and the snowpack. This discontinuity could have been the detachment surface of the slab avalanche. On 14 October, during the snow melting (spring), a natural dam, formed from previous lahars, broke in the Colorado River, and increased the normal streamflow in the lower watershed (at Villa La Angostura town) twofold.

Six months after the main eruption phase, a small lahar occurred in the Maderera Misiones catchment and reached road N. 231. Has et al. (2012) reported two landslide scars in the upper catchment at 1400 m, and a natural dam that acted as a barrier and stopped the initial flow. The lahars initiated with the dam break and flowed through the Maderera Misiones River into the lower catchment, reaching road N. 231. On route down to 1000 m, the flow incorporated material and logs from the river bed and deposited lateral levees composed of tephra, boulders and logs. The flow cross section at 1000 m was estimated at 20 m^2. The flow began to stop at the foot of the slope and the deposit here was estimated to be 10 m across and 0.8 m high. The travel distance was approximately 2.3 km, but the total lahar volume was not reported. On 30 January, after 113 mm of rainfall had accumulated in three days, floods and landslides were reported in Villa la Angostura. The remobilization of volcanic material by surface water and in the Las Piedritas River clogged the sewer, raising the water level and causing floods in the distal alluvial plain (low area of the town).

Finally, starting on 2 June 2012, 223 mm rain fell in one week (Fig. 3b) causing several flood events in the Villa La Angostura area. At the same time, heavy rain (120 mm in 24 hours measured at the Paráje Rincon station, Fig. 3a) produced flooding in the Totoral drainage basin, damaging several houses located in the alluvial plain. During our 2016 survey, we observed lahar-deposit levees in two small channels of the catchment La Ponderosa at 1360 meters.

3. Methods

3.1 Field campaign and data acquisition

Three field campaigns were conducted to examine and collect samples in the lahar prone watersheds near both the Chile-Argentina border and Villa La Angostura (Fig. 1b). Observations in the source areas around Villa La Angostura were performed at separate intervals during the eruptive period. Due to rugged terrain, narrow channel and dense vegetation, the upper catchments near the Chile-Argentina border were inaccessible. The first campaign was in September 2011 (end of winter), the second in November 2011 (spring), and the third in November 2016 (spring). During the 2016 field campaign we collected 21 undisturbed tephra-fallout samples for geotechnical tests. Sampling was performed by pushing a steel tube of height 15 or 30 cm (depending on the deposit thickness) and diameter 10 cm into

6

223   the ground. The samples were collected from preserved 2011 tephra fallout deposits in forested areas
224   near the "Diana waterfalls" watershed (Fig 1b, S1and S2) and on the east hillslope of Cerro Bayo, at
225   1530 m (Fig. 1b, S3 an S4). We also make use of reports, photographs and oral communications by the
226   Villa La Angostura Civil Defence and Mountain Rescue group regarding lahar and flood events during
227   the volcanic eruption. Immediately after the main tephra fallout and during the following week,
228   volunteers from "Comité de montaña", Civil Defence and SEGEMAR surveyed the upper catchment
229   collecting data from snow pits. Rainfall data, acquired from 3 rain gauges (Cerro Mirador, Rincón and
230   Villa La Angostura) for the period 2000-2015, has been provided by the Interjurisdictional Watershed
231   Authority from Neuquén Province (AIC). The stations are in the study area at altitudes between 700 and
232   1000 m on a west-east trend (Fig. 1b). All of the gauges supply daily rainfall. A geomorphological map
233   for 6 watersheds in Villa la Angostura was compiled in ArcGIS using on field data, satellite imagery
234   (Spot and Aster) and a digital elevation model (DEM). A DEM with a 10 m resolution provides the
235   topographic basis for the triggering modelling. The DEM was built from a stereo pair of satellite images
236   from World View 2, taken on the 24 December 2011 (CONAE, Argentina, 2011).
237
238   **3.2 Slope stability models**
239   Two stability models, namely SHALSTAB (Shallow Slope Stability Model; Montgomery and Dietrich,
240   1994) and TRIGRS (Transient Rainfall Infiltration and Grid-Based Regional Slope-Stability Model;
241   Baum et al., 2008) were applied to identify unstable areas along the slopes, and their results compared.
242   The test area comprises 2 upper catchments located in Villa La Angostura (Fig. 1b).
243
244   3.2.1 SHALSTAB model
245   SHALSTAB is a quantitative model for assessing topographic influence on shallow landslides,
246   combining a steady-state hydrological model with a simple slope stability model (Montgomery and
247   Dietrich, 1994). The authors use steady-state shallow subsurface flow equations, based on the work of
248   O´Loughlin (1986), to model hydrologic controls on h/z, where z is the soil depth, and h the water table
249   above the failure plane. The catchment is divided into topographic elements defined by elevation
250   contours orthogonal groundwater flow direction. The net rainfall (rainfall minus evapotranspiration) in
251   each topographic element becomes groundwater flow. The hydrological model thus reduces to the
252   calculation of a so-called wetness W, which is the ratio of local groundwater flux at a given steady state
253   rainfall to that at soil profile saturation
254
255    $$W = \frac{qa}{bT\sin\beta},$$ (1)
256
257   where q is the net rainfall intensity, a the upslope contributing area draining across b, which is the
258   cumulative drainage area of all topographic elements draining into an element, T soil transmissivity

259   when saturated, and β the local slope in (degrees) of the ground surface. The parameters a, b, and β are

260   obtained from the DEM used in the analysis. There is a model assumption that for $W > 1$ excess water

261   runs off as overground flow.

262   SHALSTAB is based on an infinite slope model including a Mohr-Coulomb soil failure criterion, which

263   compares the destabilizing force of gravity, to the soil shear resistance forces of friction and cohesion

264   along a failure plane parallel to the ground surface. Edge effects are neglected (Tarolli and Tarboton,

265   2006).

266   Initially, Montgomery and Dietrich (1994) considered cohesion $c\square$ to be zero, which is realistic for air-

267   fall cohesionless volcanic materials. After neglecting cohesion, the infinite slope stability model can be

268   solved for h/z, which is the proportion of the soil column that is saturated at failure

269

270    $$\frac{h}{z} = \frac{\rho_s}{\rho_w}\left[1 - \frac{\tan\beta}{\tan\phi}\right],$$ (2)

271

272   where ρ_s and ρ_w are the soil and water bulk density respectively and $\square$ is the soil friction angle.

273   Combining the slope stability hydrological models (Eqs. 1-2), we obtain the critical steady-state

274   precipitation event q/T required to cause instability for cohesionless soils

275

276    $$\frac{q}{T} = \left(\frac{b\sin\beta}{a}\right)\left(\frac{\rho_s}{\rho_w}\right)\left[1 - \frac{\tan\beta}{\tan\phi}\right].$$ (3)

277

278   Later, Montgomery et al. (1998) provided the equation for the case of soils with cohesion and the q/T

279   required to cause instability is

280

281    $$\frac{q}{T} = \left(\frac{b\sin\beta}{a}\right)\left[\frac{c'}{\rho_w g z \cos^2\beta \tan\beta} + \frac{\rho_s}{\rho_w}\left(1 - \frac{\tan\beta}{\tan\phi}\right)\right]$$ (4)

282

283   Further details for the model with cohesion can be found in Montgomery et al. (1998). Both model with

284   and without cohesion have been used in this work.

285   SHALSTAB classifies landslide susceptibility as: "unconditionally stable", "potentially unstable" and

286   "unconditionally unstable". Slopes that are stable even when $W=1$ (wet) are classified as unconditionally

287   stable, and excess pore pressure is needed to generate slope instability. Similarly, slopes predicted to be

288   unstable even when dry $(W=0)$ are unconditionally unstable. Landslide susceptibility is calculated using

289   equations 3 and 4, obtaining a range of values for q/T. Since T is much greater than q, q/T is very small,

290   meaning it is helpful to convert these values to $\log(q/T)$, which are found to range from -10 and 10.

291   Then, a range of $\log(q/T)$ is defined for each landslide susceptibility class. Critical rainfall values can

292   be calculated for each class, if the transmissivity is estimated. A fundamental premise of the model is

293   that areas found to have the lowest q/T values (least amount of precipitation required) represent the least

294   stable areas and have the greatest potential for shallow landslides. Thus, the general scope of such

295   analysis is to attribute the q/T values as "instability descriptors" to each single cell of the DEM (Digital

296   Elevation Model).

297

298   3.2.2 TRIGRS model

299   The Fortran Program TRIGRS was developed by Baum et al. (2002) for computing transient pore

300   pressure changes and subsequent changes in safety factors due to rainfall infiltration. They extended the

301   method of Iverson (2000) by implementing a complex time sequence of rainfall intensity, an

302   impermeable basal boundary at infinite depth and an optional unsaturated zone above the water table

303   (Baum et al. 2002). They also, added a simple runoff-routing scheme to disperse excess water from cells

304   where the rainfall intensity exceeds the infiltration capacity, according to the vertical Darcy flow

305   velocity (product of soil conductivity and vertical hydraulic gradient), where "the infiltration (I) at each

306   cell is the sum of precipitation (P) plus any runoff from upslope cells (R_u). When $P+R_u$ exceeds K_{sat}, the

307   excess is considered runoff and is diverted to upslope adjacent cells". TRIGRS combines models for

308   infiltration and groundwater flow due to rainfall, routing of runoff, and slope stability to calculate the

309   effects of storms on the stability of slopes over large areas.

310   The infiltration models in TRIGRS for wet initial conditions are an extension of Iverson's (2000)

311   linearized solution of the Richards equation for Darcian flow of groundwater in response to rainfall on

312   the slope(Baum et al., 2002; Savage et al., 2003, Savage et al., 2004). The solution is valid only where

313   transient infiltration is vertically downward and transient lateral flow is relatively small. The equation

314   for the transient vertical groundwater flow is

315

316     $$\frac{\partial \psi}{\partial t} C(\psi) = \frac{\partial}{\partial Z}\left[K(\psi)\left(\frac{\partial \psi}{\partial Z} - \sin \beta\right)\right], \qquad\qquad\qquad (5)$$

317

318   where ψ is the pressure head, $K(\psi)$ is the pressure- head dependant hydraulic conductivity, $C(\psi) = \mathrm{d}\theta/\mathrm{d}\psi$

319   is the specific moisture capacity, θ is the volumetric water content, β is the slope angle of the ground

320   surface, and Z is the vertical depth (Philip, 1991). For wet initial conditions the pressure dependent

321   quantities $K(\psi)$ and $C(\psi)$ become constant values, K_{sat} and C_0, where C_0 is the minimum slope of the soil

322   water retention curve, which is the relationship between θ matric suction used to describe unsaturated

323   soil behaviour (Fredlund and Xing 1994). Equation 5 then reduces to a linear diffusion equation

324

325     $$\frac{\partial \psi}{\partial t} = D_1 \frac{\partial^2 \psi}{\partial Z^2}, \qquad\qquad\qquad\qquad (6)$$

326   where $D_1 = D_0/\cos^2 \beta$ and $D_0 = K_{sat}z/C_0$.

327   An analitical solution to Eq. (6) is proposed by Baum et al. (2002) and implemented in a Fortran code.

328   Additionally, TRIGRS is applicable to unsaturated initial conditions with a two layer system consisting

329   of a saturated zone with a capillary fringe above the water table, overlain by an unsaturated zone

330   extending to the ground surface. The unsaturated zone acts like a filter that smoothes and delays the

331   surface infiltration signal at depth. The model utilises the soil-water characteristic curve for wetting of

332   the unsaturated soil as proposed by Gardner (1958) and approximates the infiltration process as one-

333   dimensional vertical flow (Srivastava and Yeh, 1991, Savage et al., 2004).

334   Following Iverson (2000), slope stability is calculated using an infinite-slope stability analysis. Incipient

335   failure of infinite slopes is decribed by an equation that balances the downslope component of

336   gravitational driving stress against the resisting stress due to basal Coulomb soil friction, incorporating

337   the influence of groundwater (Iverson, 2000). The Factor of Safety (FS) is calculated at a depth Z by

338

339   $$FS\ (Z,t) = \frac{\tan\phi'}{\tan\beta} + \frac{c' - \psi\ (Z,t)\gamma_w\tan\phi'}{\gamma_s Z\sin\beta\cos\beta} \qquad , \tag{7}$$

340

341   where t is time, c' is the effective soil cohesion, ϕ' the effective friction angle, γ_w is the unit weight of

342   groundwater and γ_s is the unit weight of soil. The pressure head $\psi\ (Z, t)$ in (7) is obtained from various

343   formulae depending on the particular conditions modelled.

344   FS is calculated for pressure heads at multiple depths (Z). The slope is predicted to be unstable where

345   FS< 1, in a state of limiting equilibrium where FS=1 and stable where FS>1. Thus, the depth Z of

346   landslide initiation is where FS first drops below 1.

347

348   3.2.3 Simulation parameters, slope stability scenarios and comparison with field data

349   We chose two representative upper basins (Colorado and Florencia), where the recent tephra-fallout

350   deposit located on the bare slopes were partially remobilised by shallow landslides, as locations where

351   to apply the SHALSTAB and TRIGRS models.

352   For both SHALSTAB and TRIGRS, the thickness of the tephra-fallout deposit was taken to be 15 cm,

353   in agreement with the corresponding isopach map (Fig. 1a) (Pistolesi et al., 2015). However, the tephra-

354   fallout deposit in this area consists of two very different layers with contrasting grain sizes: Unit I+II,

355   and Unit III of Pistolesi et al. (2015). In Villa La Angostura (48 km from vent) the typical sequence is a

356   5 cm upper ash layer, with a median grain diameter of 0.25 mm (Unit III), and a 10 cm lower layer with

357   a median grain diameter of 1 mm (Unit I and II). As the model SHALSTAB and TRIGRS are designed

358   for homogenous deposits or soil, the modelled tephra-fallout deposit was simplified, and two different

359   scenarios were considered. The first scenario (S1) considers a homogenous ash layer of 15 cm (i.e.

360   UnitI+II+III) (Fig. 4). For the second scenario (S2), we considered a 10-cm thick lapilli layer (Fig. 6).

361   For locations 28 km from the vent area we used a tephra-fallout deposit thickness of 30 and 20 cm for

362   scenarios 1 and 2 respectively.

363   The soil mechanical features, as well as the slope angle are reported in Tab. 1-2. Hydraulic conductivity

364   (K_{sat}) for lapilli was measured in the field by filling a plastic tube with an undisturbed sample and

365   saturating the material with water. We then measured the time taken for the water to drain to estimate

the hydraulic conductivity (Tab. 2). As long term and short-term pressure head responses can be described with the simplified forms of Richards equations, where the ratio $\square = Z/\sqrt{a} \ll 1$, we obtained the upslope contributing area (a) for the lahar source zones using Taborton´s (1997) D − ∞ algorithm and calculated $\square$ for a depth of 0.15 m. Values of $\square$ are between 0.015 and 0.004.

For initial conditions in TRIGRS, we considered the water table to be located at the bottom of the tephra sequence (lower boundary) and an impermeable layer below the tephra sequence, which were emplaced on cretaceous rock outcrops, debris mantled slopes and volcanic soils.

The rainfall intensity I was determine according to daily rainfall data from the "Villa La Angostura" station (AIC, Autoridad Interjuridiccional de Cuencas, 2016). The provided rainfall data were daily totals. Therefore, in order to test the effect of different rainfall intensities on landslide processes in the upper catchment, we ran (with TRIGRS) both scenarios for three different rainfall duration and intensities (Tab. 3). Since parts of the upper basin areas are forested, mainly by Nothofagus pumilio, a fraction of the rainfall is intercepted by the canopy. We thus considered a rain interception of 25 % based on the study of Martinez Pastur et al. (2011). The rainfall intensities for both bare soil area and forested area are given in Tab. 3. For SHALSTAB, a critical rainfall intensity of 5 mm/day was considered, corresponding to the mean rainfall intensity for 2011 (Sorbino et al., 2010).

To investigate the sensitivity of TRIGRS to cohesion variations in scenario 1, we perform calculations for cohesion values between 0 and 1.5 kPa, for two different tephra thicknesses 0.15 m and 0.30 m (Fig. 7). As the result obtained with zero cohesion for scenario 1 overestimated the unstable area (80% for 0.15 m thickness, 46% for 0.3 m), we took cohesion to be equal to 0.5 kPa for the soil thickness of 0.15 m and 1 kPa for 0.30 m.

The results from the different models and scenarios were compared with the field observations collected during the eruption and 5 years later in the upper catchment Florencia. Particularly, we measured the 2011 tephra sequence in the different geomorphologic units and described the remobilization by rill erosion and shallow landslides. Based on field observation, measurements of 2011 tephra deposits and description on remobilization by rill erosion and shallow landslides, a map of the observed lahar source areas was created for the Florencia river upper catchment. Finally, the unstable volume for of bare soil area was calculated in each case.

Three metrics are used to compare the observed and simulated lahar source areas: the Jaccard similarity coefficient (Levandowsky and Winter, 1971), and two metrics derived from Bayes $\square$ Theorem, model sensitivity and model precision (Charbonnier et al., 2018).

We quantified the quality of model fitting to the field evidence through the Jaccard similarity coefficient (R_J) (Kubanek et al. 2015) by dividing the intersection of the observed source areas A_{obs} and unstable modelled areas A_{sim} by their union. Observed source areas determined unstable by the model ($A_{obs} \cap A_{sim}$) are classified as true positive (TP), and those computed as unstable by the models, but not observed as source areas are false positive (FP). Conversely, observed source areas that are simulated to be stable are false negative (FN) and true negative (TN) areas are those observed to be stable in reality and the

11

403 simulation (Fig. 8).

404 The Jaccard similarity coefficient is

405 $$R_J = \frac{|A_{obs} \cap A_{sim}|}{|A_{obs} \cup A_{sim}|} \times 100 = \frac{TP}{TP + FN + FP} \times 100 \quad . \tag{8}$$

406 Charbonnier et al. (2018) identified the similarity of R_J to Bayes ☐Theorem, which describes the

407 probability of predicted unstable areas, based on the observed source areas and vice-versa. Bayes

408 ☐Theorem serves as link between simulated and observed unstable areas

409 $$P_r\left(A_{obs}/A_{sim}\right)P_r(A_{sim}) = P_r\left(A_{sim}/A_{obs}\right)P_r(A_{obs}) , \tag{9}$$

410 where, $P_r(A_{obs})$ is the probability that an area is unstable according to observations, $P_r(A_{sim})$ the

411 probability that an area is unstable according to the model, $P_r\left(A_{sim}/A_{obs}\right)$ the conditional probability

412 of A_{sim} given A_{obs} and $P_r\left(A_{obs}/A_{sim}\right)$ the conditional probability of A_{obs} given A_{sim}. Rearranging

413 equation 9, we obtain

414 $$P_r\left(A_{obs}/A_{sim}\right) = \frac{P_r\left(A_{sim}/A_{obs}\right)P_r(A_{obs})}{P_r(A_{sim})}. \tag{10}$$

415 We identify $P_r\,(A_{obs}/\,A_{sim})$ as the model precision and $Pr\,(A_{sim}/A_{obs})$ as the model sensitivity.Combing

416 Eqs. 8 and 10. we obtain

417 $$R_{MP} = \frac{|A_{obs} \cap A_{sim}|}{|A_{sim}|} \times 100 = \frac{TP}{TP + FP} , \tag{11}$$

418 and

419 $$R_{Ms} = \frac{|A_{obs} \cap A_{sim}|}{|A_{obs}|} \times 100 = \frac{TP}{TP + FN}. \tag{12}$$

420 A high model precision R_{MP} indicates a good match between the model predicted and observed unstable

421 areas, without taking into consideration observed unstable areas not simulated by the model. The model

422 sensitivity R_{MS} gives the percentage of lahar source areas that the model correctly predicts, without

423 counting the FP areas. A model simulation with high sensitivity agrees well with the observed source

424 areas, but may predict unstable areas outside the observed sources.

425

426 **4. Results**

427

428 **4.1 Geomorphology of the study area and field observations**

429 4.1.1 The watersheds of Villa La Angostura

430 Villa La Angostura is situated on an alluvial fan, at the convergence of six watersheds: Las Piedritas,

431 Colorado, Florencia, La Ponderosa, Maderera Misiones and El Muerto. Fig. 9 shows the

432 geomorphological map of all the watersheds.

433 The Las Piedritas River upper catchment is a U-shaped valley with steep lateral hillslopes. The river

434   originates at an elevation of 1450 m in an old glacial cirque and runs through the valley until 1090 m.

435   At this point, the river makes a 60 m jump (waterfall) and flows through a granitic canyon for almost

436   0.6 km before reaching the alluvial fan. The river runs between small hills to the south, before crossing

437   Villa La Angostura and reaching the Nahuel Huapi Lake. The Colorado River also originates in a small

438   glacier cirque, at an elevation of 1650 m, and is fed by many first order channels which drain the back

439   and lateral upper catchment slopes (talus debris slope and forested hillslopes). The upper Colorado basin

440   has an old cirque front slope, with many small channels draining the area. Below 1250 m the stream

441   flows through a narrow valley with several cascades and reaches the lower catchment at 890 m, near the

442   Villa La Angostura water plant. The Florencia upper catchment is a small glacial cirque with steep slopes

443   and debris flow prone channels. The river in the middle catchment is confined and has several waterfalls

444   before reaching the lower catchment and run into the Colorado River. The back and lateral hillslopes

445   have been divided in three main geomorphological units: rock outcrops, talus debris slopes and forested

446   hillslopes. Two debris-flow prone channels have been mapped, one in the Florencia and the second in

447   Colorado upper catchment. The summit and valley interfluves are glacially smoothed, with plane and

448   gentle slopes between 0° and 15°. The middle watershed is located on the regional Nahuel Huapi valley

449   flank. The proximal alluvial fan area has glacial erosional landforms and 10 m high elongated hillslopes

450   with some swamps areas filled with volcanic sediments. The distal alluvial fan is flat marshland crossed

451   by the Piedritas River, which splits into at least two channels before reaching the Nahuel Huapi Lake.

452   The La Ponderosa, Maderera Misiones and El Muerto watersheds drain the southern Cerro Bayo flank

453   (Fig. 1b). The upper catchment source area has steep slopes between 15° and 45° and range in area from

454   0.153 km^2 (La Ponderosa) to 0.62 km^2 (Madera Misiones). The uppermost part of the hillslopes pertains

455   to the Cerro Bayo ski field and has been modified with a manmade track.

456

457   4.1.2 Shallow landslides and erosion in the lahar source areas

458   Examination of the upper catchment of Villa La Angostura in spring (November 2011) found different

459   types of erosion and remobilization. Rills formed just below the crest, at slope angle of 25-35°, and

460   below rock outcrops located on the top or middle hillslope (Fig. 6d). Rills formed were erosion was

461   focused by flow-concentrating irregularities, such as boulders (Horton, 1945, Mannville et al., 2000),

462   on the talus debris slope unit. Otherwise shallow landslides occurred where the tephra deposit was lying

463   either on outcropping rock or the steep talus debris slope (Fig. 9d). The tephra was mobilized into first

464   order drainage concavities (Fig. 9e), located on the talus debris slope. Measurements, on these mobilized

465   deposits, performed in November 2011, found a 15 cm primary tephra fall deposit and 15 cm of

466   remobilized tephra.

467   Observations from December 2016 in the upper catchment of Florencia and Colorado, showed that the

468   2011 tephra deposit was almost completely eroded from the rock outcrops and bare hillslope since the

469   old black tephra deposit cropped out in many hillslope areas, as before the 2011 eruption. Nevertheless,

470   a small fraction of tephra forms the debris slope colluvium, which is a mix of tephra and rock boulders.

471   Part of the eroded tephra remains in the first order channels. Figure 10 shows the tephra deposit erosion

472   stage in the Florencia upper catchment between June 2011 and December 2016. Tephra was first

473   removed from rock outcrops, which represent 22% of the whole upper catchment (Fig. 10b). At the same

474   time tephra from hillslopes (31%) was partially removed by rill erosion, shallow landslides and

475   downslope creep mostly related to snowmelt during the spring (Fig. 10b). Five years later, tephra was

476   almost completely removed from rock outcrops and unforested hillslopes. Part of the eroded sediment

477   remained in the first order channel and in the forested areas which represent 8% and 39% of the upper

478   catchment, respectively (Fig. 10c). The tephra sequence was well preserved on the forested hillslopes.

479

480   **4.2 Modelling**

481   Instability scenarios modelled with SHALSTAB for two upper watersheds (Colorado River and

482   Florencia River) in Villa La Angostura are presented here. Figure 11 shows two maps with different

483   levels of instability computed with the SHALSTAB model (scenario 1, Tab, 1). Figures 11a and 11b

484   show the results without and with cohesion respectively. We defined four stability descriptors:

485   unconditionally unstable, unstable, stable, unconditionally stable. The results for scenario 1 with a

486   critical rainfall (q) of 5 mm/day are presented in Tab. 4. We also ran SHALSTAB for scenario 2, but

487   the amount of rain needed for instability is out of range, between 202 and 6000 mm/day, 40 and 1200

488   times the critical rainfall value calculated for the study area respectively. In this case the unconditionally

489   unstable area represents approximatively 1% of the total upper catchments.

490

491   Figures 12 and 13 illustrate the evolution of unstable areas for the Colorado and Florencia Rivers upper

492   watersheds through time calculated with TRIGRS for scenarios 1 and 2. The colour lines represent three

493   different rainfall intensities and durations (Table 3). Figures 12a, b and 13a, b show that for scenario 1,

494   a maximum fraction of unstable area was attain for all three rainfall intensities. Conversely for scenario

495   2, the maximum fraction was only reached for the greatest rainfall intensity (3.3 mm/h). It is also seen

496   that more time is needed to reach the maximum unstable area when the tephra thickness is doubled (0.30

497   m). This modified slope response depends on the transient groundwater response time Z^2/D_0. This

498   timescale is the minimum time necessary for slope-normal pore pressure to transmit from the ground

499   surface to depth Z (Iverson and Major, 1987; Haneberg 1991; Reid, 1994). Comparing results for the

500   Colorado upper watershed (Figs. 12 a-d) with the Florencia upper watershed (Figs. 13 a-d) both show

501   qualitative agreement for the different scenarios and tephra thicknesses, but there is much greater

502   potential for unstable areas in the Florencia study case. This large difference is due to the different

503   topography, since slope angles between 38.4°and 53° cover only 6% of the Colorado upper basin area,

504   but 20% in the Florencia upper basin. Table 5 list all of the values of unstable area percentage as

505   calculated with TRIGRS for the Colorado River and Florencia River upper catchments.

506   Model validation metrics (R_J, R_{MP} and R_{MS}) calculated for the upper catchment Florencia River using

507   both TRIGRS and SHALSTAB are given in Table 2. For scenario 1 (Ash deposits), both SHALSTAB

508 (without cohesion) and TRIGRS give the highest Jaccard similarity coefficient, R_J = 48% and 47%,

509 respectively. These simulations also have high model sensitivities. For scenario 2 (lapilli deposits) the

510 highest R_J was achieved for the highest rainfall intensity and a duration of 24h.

511 TRIGRS model S1 determined most of the source areas to be in the bare soil areas (Fig. 14b), but also

512 predicted a lot of the forested stable areas to be unstable (FP). This is reflected in the model precision

513 (57%). Figure 14c shows that for the case of the lapilli deposits (scenario 2) and 24h of rainfall, the

514 model underestimated the fraction of observed source areas, but was more accurate in the forested stable

515 areas (R_{MS} 62.9%). Finally, the unstable volumes in both bare soil upper catchment areas, as calculated

516 from the best fit scenario, are presented in Tab. 7.

517

518

519 **5. Discussion**

520

521 **5.1 Remarks on modelling results**

522 The direct investigation of the lahar source areas of the 2011 eruption of Cordón Caulle provided

523 invaluable insights into lahar triggering mechanisms and the amount of remobilised tephra. Tephra-

524 fallout properties (friction angle, hydraulic conductivity and diffusivity) required by numerical models

525 and obtained for the ash and lapilli sequences around Cordón Caulle, represent the first data derived for

526 this region. The 2011 tephra-fallout stratigraphy, with 2 contrasting layers (ash and lapilli), is too

527 complex to be analysed with the SHALSTAB and TRIGRS models, which are designed for homogenous

528 deposits. We therefore considered, two different scenarios. Comparing the two models, we found that

529 TRIGRS provides an estimation of the timing (based on a storm event) and location of a potentially

530 unstable area, while the topographic effects on shallow landslides can be determined with SHALSTAB.

531 For both models, results are sensitive to the physical properties of the tephra-fallout sequence and thus

532 require accurate knowledge of the hydraulic and materiel strength properties at the scale of catchment

533 area. TRIGRS requires additional parameters such as hydraulic diffusivity and information on initial

534 water table. The results obtained with TRIGRS demonstrate that shallow landsliding is possible in our

535 study area for both ash and lapilli bearing sequences. However, the ash sequence (scenario 1), with the

536 low permeability and diffusivity, has a higher predisposition for instability compared with the high

537 permeability and diffusivity lapilli sequence (scenario 2). The same maximum unstable areas are found

538 for the 3 different storms in the case of ash sequences (S1), but the time needed to reach the maximum

539 is different. For the lapilli sequences (S2), the unstable area percentage decreases as the rainfall intensity

540 decreases, and the unstable area percentage increases with time.

541 Only having daily rainfall data, instead of more detailed hourly data, was a limitation for the simulations

542 because the relationship between rainfall intensity (I) and hydraulic conductivity (K_{sat}) determines the

543 behaviour of pressure head response. The time required to develop positive pressure and further

instability (FS<1) decreases with rainfall intensity, as demonstrated in Figs. 12 and 13 where the rate of growth of the unstable area is faster for higher rainfall intensity. Therefore, we tested different rainfall intensities and duration according to daily rainfall data in the study area.

As we already described in section 3.1.2, we did not observe shallow landslides in the 2011 tephra deposit on the forested hillslope. For the TRIGRS model we accounted for rain interception (25%) in forested areas, but still prediction many unstable areas, which did not match with observations. There are many reasons for the reduction of shallow landslides in forested areas: root strength, evapotranspiration, evaporation and preferential drainage via root systems (Phillips and Watson, 1994; Sidle and Ochiai, 2006). In our case the tree and litter roots at the base of the tephra deposit could have enhanced the drainage and thus dissipate positive pore water pressure in the tephra section (Uchida et al., 2001). Several months after deposition, a new litter of leaves and new plants growing on the tephra deposits further reduced the potential for shallow landslides and sheet erosion.

5.2. Comparison with previous lahar-triggering models

Coupled hydrological and infinite-slope models have previously been used by several authors (Frattini et al., 2004; Sorbino et al., 2007; Cascini et al., 2011) to perform spatially distributed simulations of the catastrophic landslides and flow events of Sarno, 1998. The pyroclastic deposits in the landslide source areas (above Sarno town) consisted of pumice lapilli clasts and ashy layers belonging to at least 5 different eruptions from Somma-Vesuvius volcano. Frattini et al. (2004) used a lateral flux model for kinematic subsurface storm flow and a vertical flux was simulated through a simplified formulation derived from Richard´s equation (Iverson 2000). The results showed that both vertical and lateral fluxes were responsible for landslide triggering. In our study, the results from SHALSTAB demonstrated that lateral fluxes were responsible for instabilities in ash layers (S1), but not for lapilli layers (S2). Sorbino et al. (2007) applying the physically based models SHALSTAB, TRIGRS and TRIGRS unsaturated, show that the most adequate model for analysis of shallow landslide source area in Sarno was TRIGRS unsaturated. Our result instead indicates that TRIGRS and SHALSTAB models both produced good results for ash layers, but the best model for the lapilli layers was TRIGRS.

Additional strategies for studying lahar triggering exist. For example, Volentik et al. (2009) analysed the potential hazard of lahars to the Bataan nuclear power plant site (Philippines), focusing on the nature of tephra fallout and lahar generation. They used the slope stability model, based on an infinite slope form of the Mohr Coulomb failure criterion following Iverson (2000), to calculate the potential for gravitationally induced failure of the tephra deposit on the volcano slopes, thus triggering lahars. The main difference with TRIGRS (which accounts for transient rainfall effects) is that Volentik et al. (2009) considered the effect of rainfall infiltration on the pressure head, with results from a static infinite slope analysis assuming two water table conditions. If we were to use the Volentick et al. (2009) model for the case of an ash layer (scenario 1), the percentage of unstable area would have been similar, but for a lapilli layer the model would have overestimated the unstable area. The main reason for this is that the

16

581   model does not consider the deposit hydraulic conductivity and the amount of rain needed to saturate it.
582   In the case of a probabilistic lahar-hazard assessment, with no information on hydraulic conditions and
583   tephra-fallout properties at the scale of lahar processes, the choice of a simple slope stability model to
584   calculate potential lahar source region was appropriate.
585   Mead et al. (2016) also carried out an interesting study on rain-triggered lahar susceptibility using
586   shallow landslide and erosion models. They used the model developed by Iverson (2000) with an infinite
587   slope failure criterion and the reduced form of the Richard's equation. Their shallow landslide and
588   erosion models were applied to the 1995 lahar of Ruapehu volcano, New Zealand. Here, the volume of
589   materiel remobilised from the Mangatoetoenui glacier was known and the lahar well documented, but
590   the hydraulic properties and depth of ash deposits were unknown. Different scenarios were modelled
591   with the slope stability model, varying diffusivities and infiltration rates to calculate landslide materiel
592   volume estimates which were compared with the observed volume of the Mangatoetoenui lahar in order
593   to choose the best hydraulic parameters. The main difference between the approach of Mead et al. (2016)
594   and our strategy is that we compare different slope stability models (TRIGRS and SHALSTAB) and
595   two different tephra layers for the estimate of lahars initiation volumes, while Mead et al. (2016)
596   combined a slope stability model with an erosion model to calculate the lahar initial volume.
597
598   **5.3. Implication for the compilation of lahar-hazard maps**
599   During the first 6 months of the 2011 eruption of Cordón Caulle, an estimate of the unstable tephra-
600   fallout volume in the Villa La Angostura upper catchments was made by the Argentinian Geological
601   Survey (SEGEMAR), assuming that 100% of the unforested, fallout covered areas would be eroded.
602   These estimates were used by Córdoba et al. (2015) to calculate the runout and thickness of a potential
603   lahar event in Villa la Angostura. They assumed 75% and 90 % of the deposited tephra as a high initial
604   volume for the Colorado and Florencia upper catchment respectively. Our simulation results for the
605   same bare soil upper catchments were similar with SHALSTAB and lower with TRIGRS both for ash
606   layer (Table 7). In the case of Colorado bare soil upper catchment TRIGRS model simulated that only
607   27.2% and 12.9% would be unstable for scenario 1 and scenario 2 respectively (Table 7).
608   An evaluation of potential mobilised volumes is fundamental to the modelling of both lahar spreading
609   and inundation that is required for the compilation of lahar-hazard maps. Our results demonstrate the
610   importance of using physical models (e.g. TRIGRS) combined with geotechnical data (when available)
611   so to provide accurate estimates of remobilised tephra volumes. Our results also show the potential of
612   using the SHALSTAB and TRIGRS models to predict the solid fraction of a lahar triggered by rainfall.
613   As an example, our new data on rain-triggered lahars for the 2011 eruption of Cordón Caulle can be
614   used as input for lahar-hazard assessments. The results obtained with shallow landslide models using
615   rainfall intensity, duration and frequency thresholds (IDF), could be used to create a rain-triggered lahar
616   susceptibility map (Mead et al., 2016). To assess the lahar hazard we could combine the unstable area
617   results for variable rainfall IDF (rain triggered lahar susceptibility map) with a lahar run-out model. The

618 total unstable area in each river upper catchment is the potential mobilised sediment volume used as
619 input in lahar run-out models. In addition, the evaluation of unstable area obtained with physically based
620 models such as TRIGRS or SHALSTAB provide important inputs for compiling dynamic lahar-hazard
621 maps of long-lasting eruptions, that can be adapted through time according to the varying eruptive
622 dynamics (variation of tephra thickness and grainsize, rainfall, etc.).
623

624 **6. Conclusions**
625 We have presented a detailed analysis of rain-triggered lahars combining geomorphological
626 observations and geotechnical data with numerical studies based on two slope-stability models:
627 SHALSTAB and TRIGRS. We applied our methodology to events that occurred after the main phase of
628 the 2011 eruption of Cordón Caulle volcano, Chile. SHALSTAB mostly describes topographic influence
629 on shallow landsliding, while TRIGRS accounts for rain infiltration to assess unstable areas. Due to the
630 heterogeneity of the tephra-fallout deposits that generated the lahars (i.e. combination of ash and lapilli
631 layers), different initial conditions have been tested. Tests consisted of two main scenarios with different
632 soil and hydraulic parameters. In addition, different rainfall intensities and durations and two values of
633 tephra-fallout thicknesses have been used in TRIGRS for the analysis of shallow landsliding. The
634 comparison between field observations and model outcomes shows that:
635 1) both SHALSTAB and TRIGRS provide a goodness of fit of nearly 50% for ash layers (scenario 1)
636 and also a high model sensitivity. The SHALSTAB model for lapilli layers (scenario 2) was not realistic,
637 but scenario 2 simulated with TRIGRS had a fit of 40 % and high precision in the case of high rain
638 intensity;
639 2) results from both models are sensitive to the physical properties of the tephra-fallout deposit and to
640 the hydraulic properties and strength of the material in the study area. TRIGRS requires more parameters
641 than SHALSTAB, such as hydraulic diffusivity and information on initial water table;
642 3) TRIGRS provides estimates of the timing (based on a storm event) and location of a potentially
643 unstable area, while SHALSTAB provides landslide susceptibility classes based on critical rainfall
644 values.
645

646 The comparison between the different scenarios modelled with TRIGRS shows that:
647 1) an increase in rainfall intensity increases the extent of unstable areas for lapilli layers (scenario 2).
648 Conversely, the extent of unstable areas is not affected by rainfall intensities for ash layers (scenario 1);
649 2) the time taken for positive pore pressure to develop is larger for larger values of tephra-fallout
650 thickness (ground surface to depth H) and depends on the transient groundwater response time H^2/D_0.
651 The shallow landsliding process modelled with SHALSTAB and TRIGRS provides important data of
652 lahar initiation processes from recent tephra fallout sequences, emphasising in particular the importance
653 of considering the presence of different ash layers and associated geotechnical data in the susceptibility
654 of a slope.

671   **Appendix A**

672

673   Notation

674

Symbol	Description	Units
a	upslope contributing area	m^2
b	grid cell width	m
c	cohesion	kPa
c′	effective soil cohesion	kPa
C (ψ)	moisture capacity	1/kPa
D_0	hydraulic diffusivity	m^2/s
d	depth of the steady state water table	m
d_{lb}	vertical depth to the lower boundary	m
FS	factor of safety for a hillslope	
h	water level above the failure plane	m
I_z	long term (steady state) rainfall flux at the ground surface	mm/h
K	hydraulic conductivity	m/s
K_{sat}	saturated hydraulic conductivity	m/s
Kz	hydraulic conductivity in the Z direction	m/s
q	net rainfall rate	mm/day
T	soil transmissivity	m^2/s
W	wetness	
z	soil depth	m
Z	depth of landslide initiation	m
β	local slope	deg
φ	angle of internal friction	deg
φ′	affective angle of internal friction	deg
ρ_s	soil bulk density	$tonne/m^3$
ρ_w	water bulk density	$tonne/m^3$
γ_s	unit weight of soil	kN/m^3
γ_w	unit weight of water	kN/m^3
□	ground-water pressure head	kPa
θs	saturated volumtric water content	%
θr	residual volumetric water content	%
α	pore size distribution index; SWCC modeling constant	
□	Time scale ratio	

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Figure captions

Figure 1: A) Isopach map on shaded relief of the cumulative Cordón Caulle 2011 tephra fall deposit in the proximal and medial area, after Pistolesi et al. (2015). Isolines between 0.2 and 3 cm from Gaitan et al. (2011). Black dots represent the tephra outcrops measured in 2012 and 2013 field campaign (Pistolesi et al., 2015). The red square indicates the study area between the Argentina-Chile border and Villa La Angostura. B) Map of study area along the Road n. 231. The green hexagons indicate the 2016 sample locations. The blue squares are the meteorological stations: M1, Paráje Rincon; and M2, Villa La Angostura. The red triangles represent the locations of observations of 2011 and 2012 lahar deposits: L1, Diana waterfalls; L2, 231 road; and L3, Maderera Misiones. The light brown line delimitates the watershed of Villa la Angostura.

Figure 2: A) Tephra sequence on April 2012 in the town of Villa La Angostura. Unit I = 4 cm; Unit II = 3cm; Unit III = 5,5 cm. B) Tephra sequence on December 2016 located in Cerro Bayo (fig 1, S3), in a small Nothofagus forest at 1530 m, with a slope inclination of 35 ° (Fig.1b, S3). Unit I = 4 cm; Unit II = 1cm; Unit III = 6 cm; and remobilized tephra on the top.

Figure 3: Timeline between June 2011 and June 2012 showing the daily rainfall data measured at A) Rincón rain gauge, near Chile- Argentina border, and B) Villa La Angostura. Occurrences of lahar and flood events during this time are marked.

Figure 4: A) Photograph of a big ditch formed through road N. 231 during the 10 June 2011 lahar event near the Chile-Argentina border (photo Diario de Villa La Angostura, https://www.diarioandino.com.ar/). B) Lahar deposit at the foot of the "Diana" waterfall, located 100m upstream from road N. 231. C) Photograph of the 2 June 2012 flood event which damaged a house on the Totoral River alluvial plain (photo Diario de Villa La Angostura https://www.diarioandino.com.ar/) D) Photograph of the Totoral River and terrace with dead trees and flooding volcanic sediment deposit. The red line indicates the flood deposit.

Figure 5: 1) Photograph of the 10 June lahar deposit, 100 m upstream from road N. 231 (L2 in Fig. 1b). The stratigraphic section represents 5 beds of alternating type A and type B lithofacies. 2) Photograph of an exposure of lahar deposit in "Diana waterfalls" creek (L2 in Fig 1b). The stratigraphic section shows a sequence of 7 beds with 3 different types of lithofacies (C, D and E).

Figure 6: Sketch of tephra sequence and thicknesses for scenarios 1 and 2.

Figure 7: Plot of the percentage of unstable area against a range of cohesion from 0 to 1.5 kPa. Calculated with TRIGRS for scenario 1 and Colorado River upper catchment

Figure 8: Sketch illustrating the relationship between True positive (TR), False positive (FP), False negative and True negative. Simulated unstable area: A_{sim}; observed source area: A_{obs}.

Figure 9A) Geomorphological map of six catchments in Villa la Angostura. River and catchment names: I-Las Piedritas, II-Colorado, III-Florencia, IV-La Ponderosa, V-Maderera Misiones, VI-El Muerto. B) Florencia river upper catchment on November 2011. On the left-side, rills formed just below the hillside crest, transporting tephra and rock boulder. On the right-side, tephra is removed by shallow landslides on the steep hillslope. C) Remobilized tephra deposit in a first order channel depression, located in the upper Colorado catchment.

Figure 10: Schematic evolution of tephra erosion between June 2011 and December 2016 in the Florencia River upper catchment. Arrows in figures A, B and C indicate the location of the photographs.

Figure 11: Instability maps obtained with SHALSTAB for scenario 1 with a critical rainfall of 5 mm/day and A) no cohesion and B) cohesion included. Different colours correspond to ranges of $\log(q/T)$ (Un. (un)stable: Unconditionally (un)stable).

Figure 12: Time evolution of the percentage of unstable areas in the Colorado upper watershed calculated with TRIGRS for scenario 1, with layer thicknesses of a) 0.15 m and b) 0.2 m, and scenario 2 with thicknesses of c) 0.1 m and d) 0.2 m. Simulations were performed with 3 different rainfall intensities and durations (3.3 mm/h, 24h; 1.6 mm/h, 48h; 0.83 mm/h, 72h).

Figure 13: Time evolution of the percentage of unstable areas in the Florencia Rivers upper watershed calculated with TRIGRS for scenario 1, with layer thicknesses of a) 0.15 m and b) 0.2 m, and scenario 2 with thicknesses of c) 0.1 m and d) 0.2 m. Simulations were performed with 3 different rainfall intensities and durations (3.3 mm/h, 24h; 1.6 mm/h, 48h; 0.83 mm/h, 72h).

Figure 14: A) Map of observed lahar source areas in Florencia upper catchment. B, C) Instability maps, obtained with TRIGRS, for B) scenario 1 and C) scenario 2, with a constant rainfall flux of 3.3 mm/h for 24 hours The black line in B) and c) separates forested area on the lower side from bare soil area on the upper side.

1 A

72°0'0"W 71°40'0"W 71°20'0"W

Vent area
Cordon Caulle
Puyehue

Chile
Argentina

30

20

0.2

1

2

3

10

5

Nahuel
Huapi
Lake

40°40'0"S

10 5 0 10 km

1 B

71°50'0"W 71°40'0"W

Chile
Argentina

Paso Cardenal
Samore

Pireco River

30

Road 231

Cº Las Tres Hermanas

L1

S1,S2 L2

Blanco River

M1

Totoral R

Paraje
Rincon

20

Correntoso

Nahuel
Huapi
Lake

R. 231

10

Piedritas R.

Colorado R.

Florencia

Aº Misiones

Cº Bayo
S3,S4

Bonito River

M2

L3

Villa La
Angostura

40°40'0"S

Unit III

Unit II

Unit I

A

UNIT III

UNIT II

UNIT I

B

S1

Rainfall

Ash

Lapilli

15 cm
(30 cm)

S2

Rainfall

Lapilli

10 cm
(20 cm)

FN

FN

FP

TP

TP

FP

TN

Legend

sim. unstable area (A_{sim})

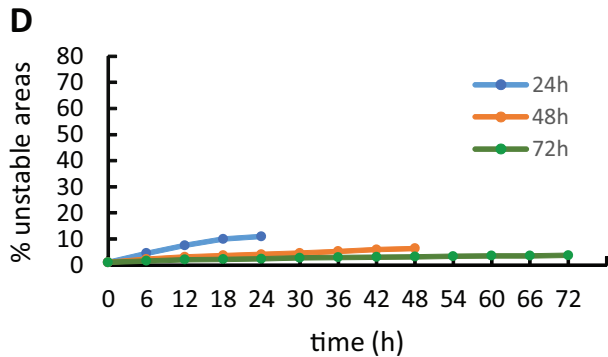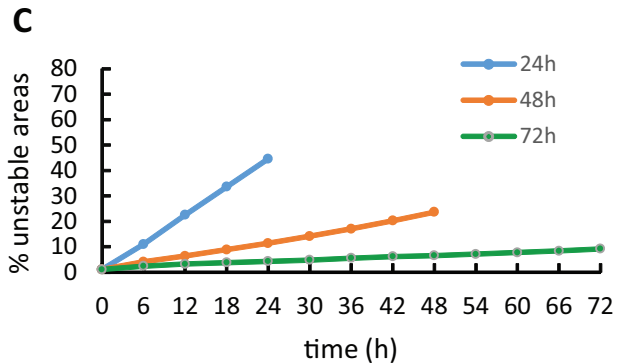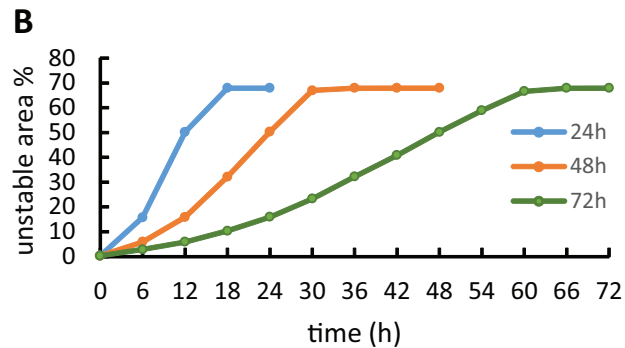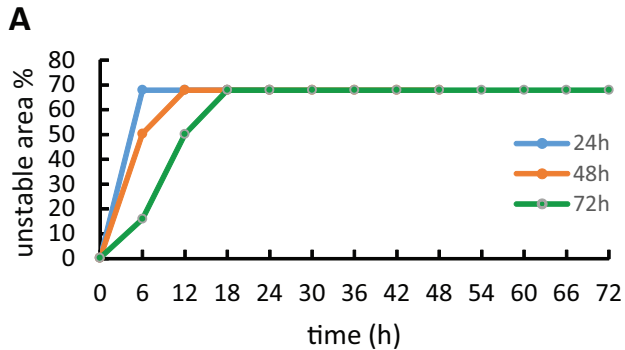obs. source area (A_{obs})

stable for obs. and sim. area
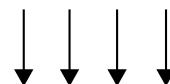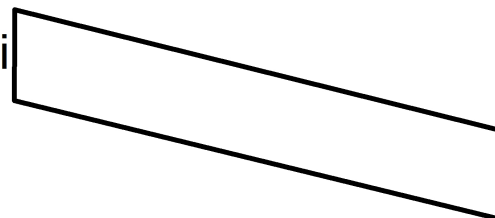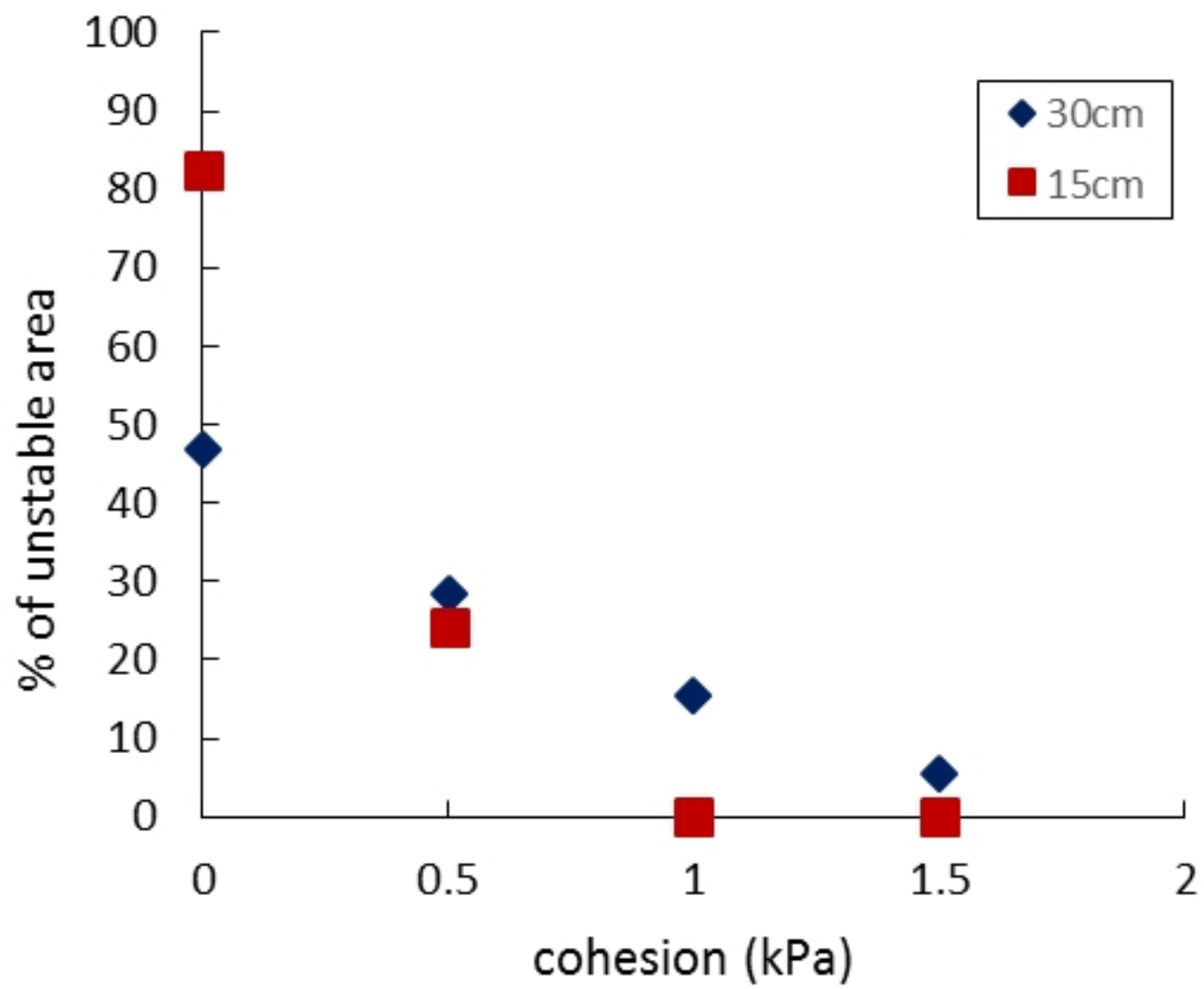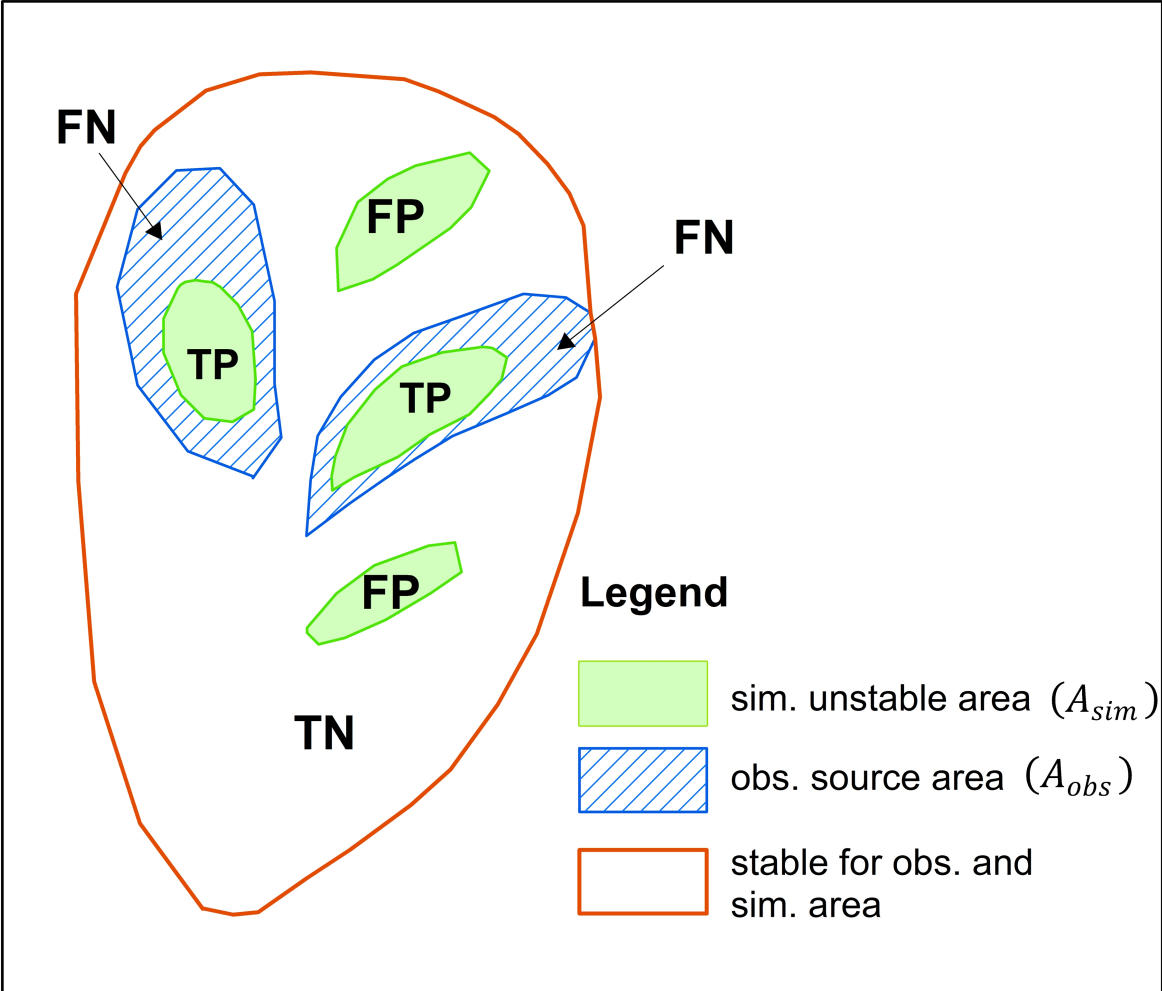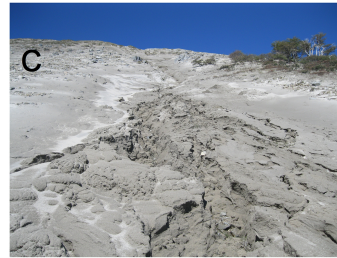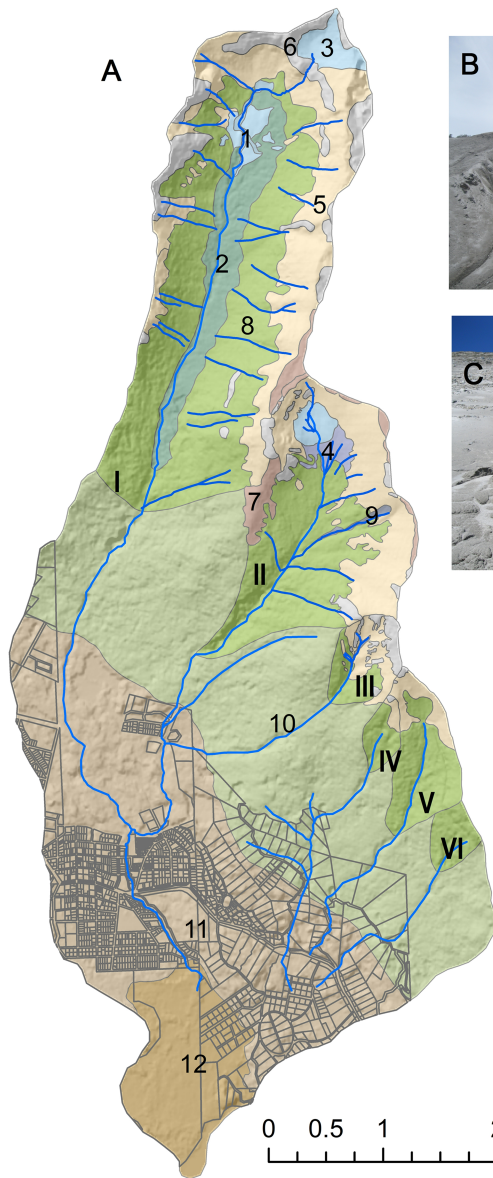
Geomorphological units

1, Alluvial plain	9, Debris flow channel
2, Forested alluvial plain	10, Regional valley flanck slope
3, Old glacial cirque depresion	11, Proximal alluvial fan
4, Old cirque front slope	12, Distal alluvial fan
5, Talus debris slope	— River
6, Rock outcrop	Villa la Angostura Plots of land
7, Glacially smoothed valley interfluves	**I-VI** Upper catchments and River
8, Forested hillslope	

Legend

log q/T

🟥	Un. unstable
🟧	< -3,1
🟨	-3,1 - -2,8
🟩	-2,8 - -2,5
🟦	-2,5 - -2,2
🟦	> -2,2
🟦	Un. stable
—	river

0   0,5   1 km

A legend:
- source areas
- tephra deposit
- tephra deposit

B legend:
- Unstable areas (FS ≤ 1)

C legend:
- Unstable areas (FS ≤ 1)

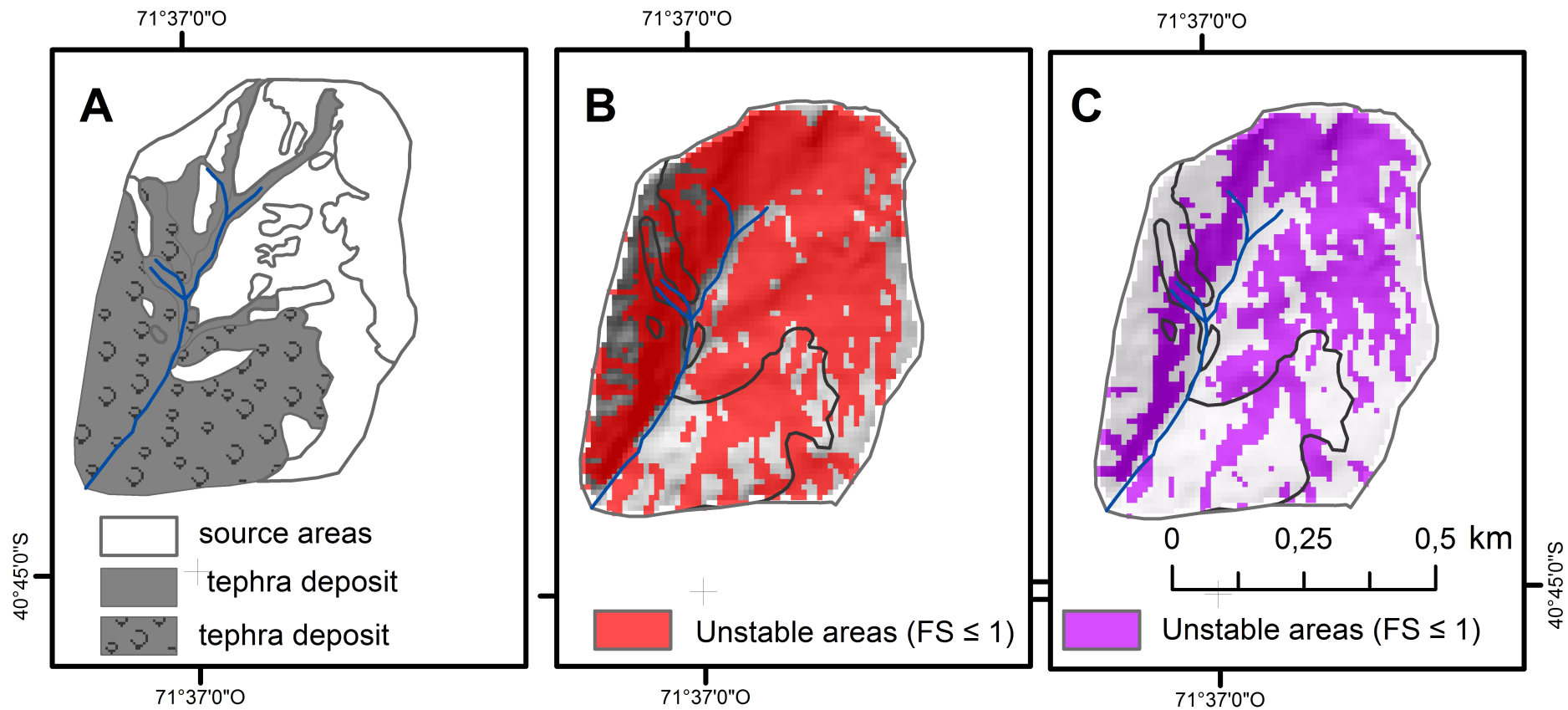71°37'0"O

40°45'0"S

0   0,25   0,5 km
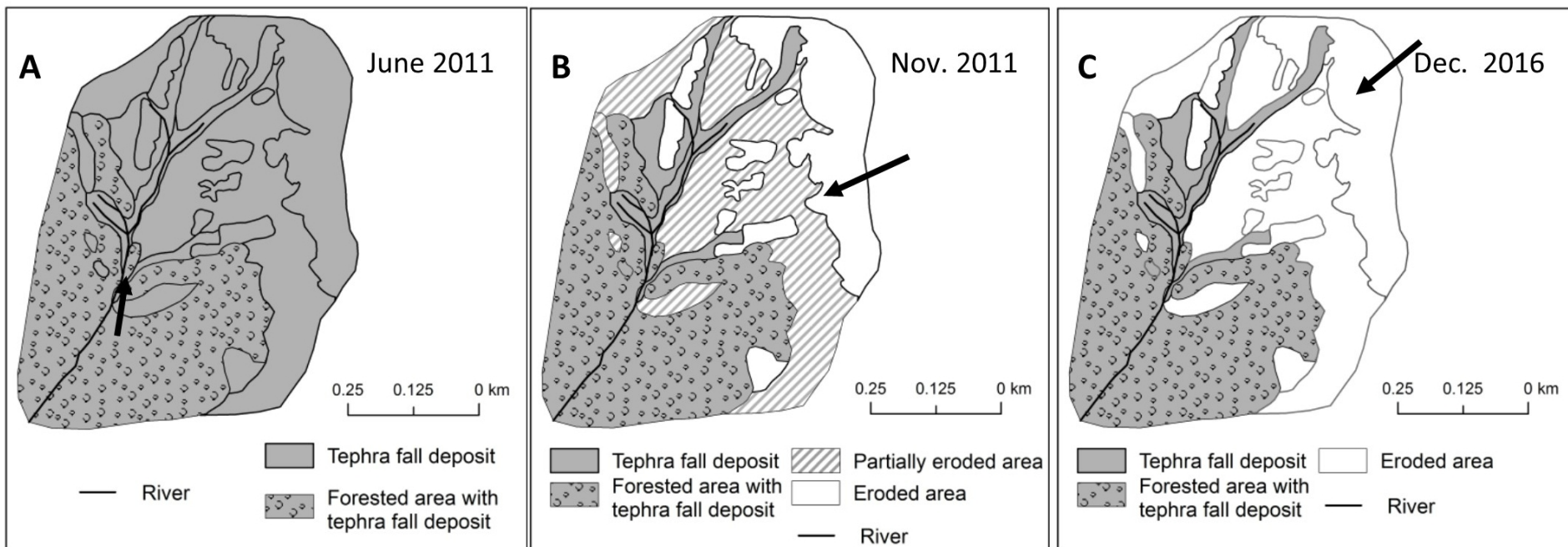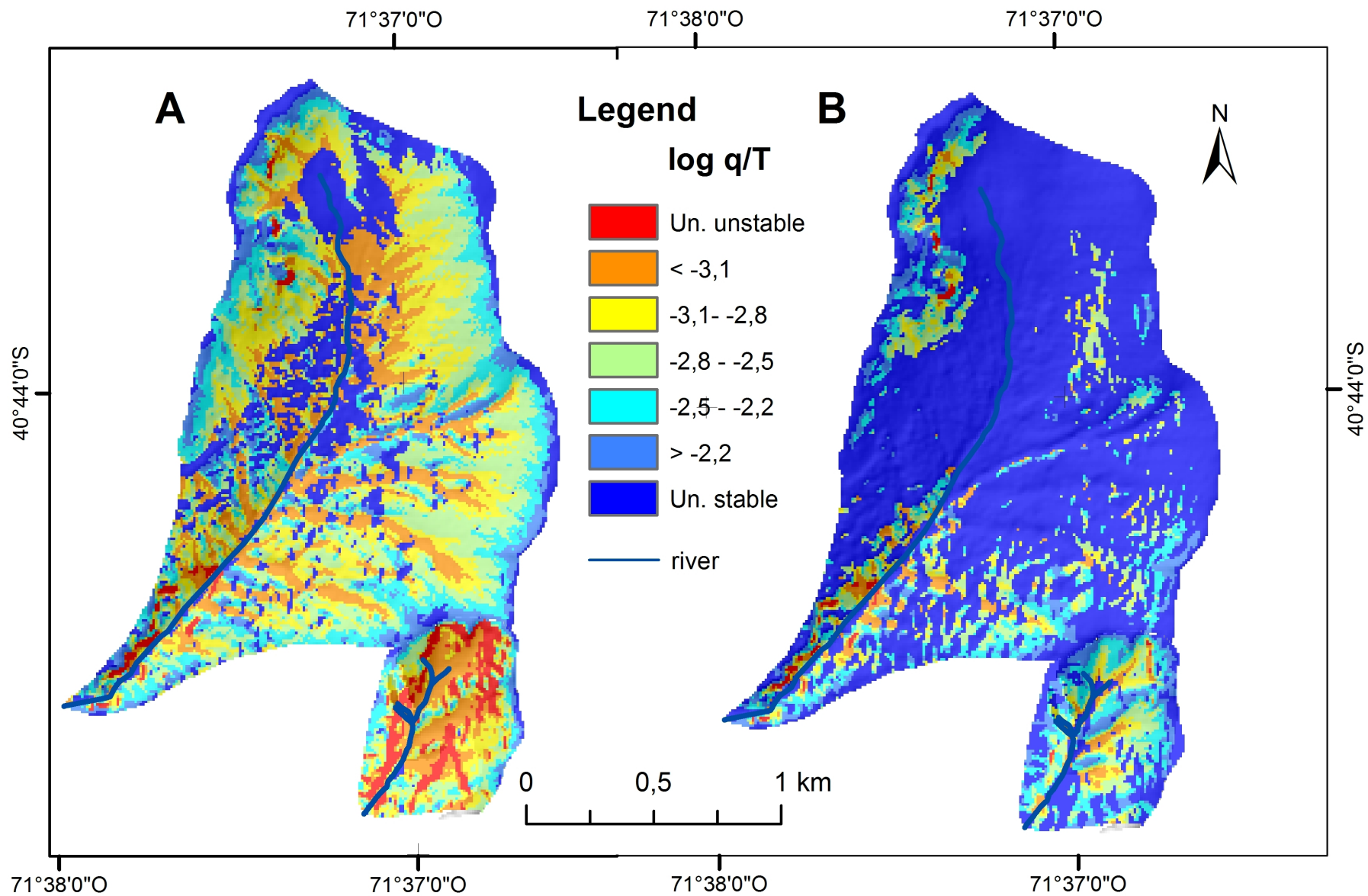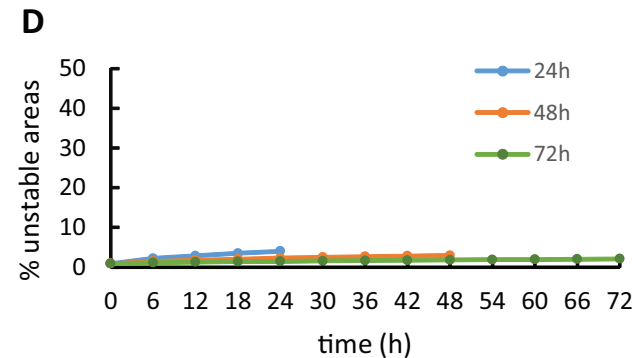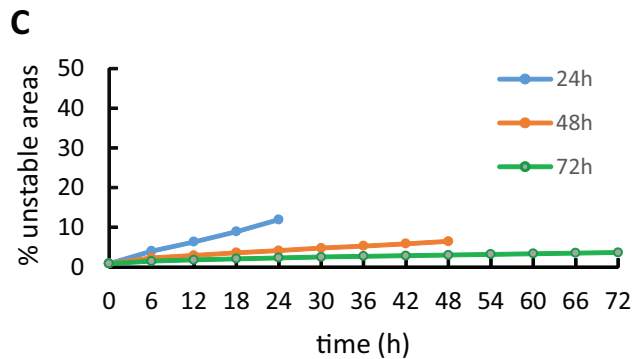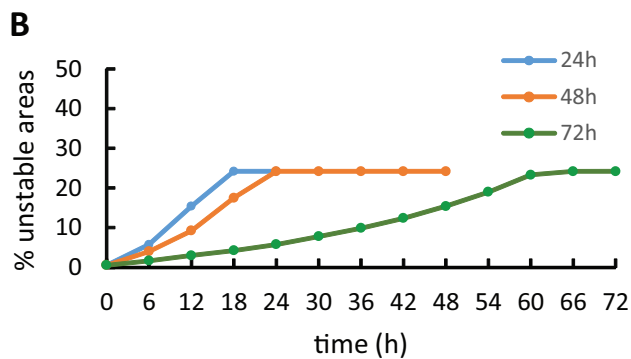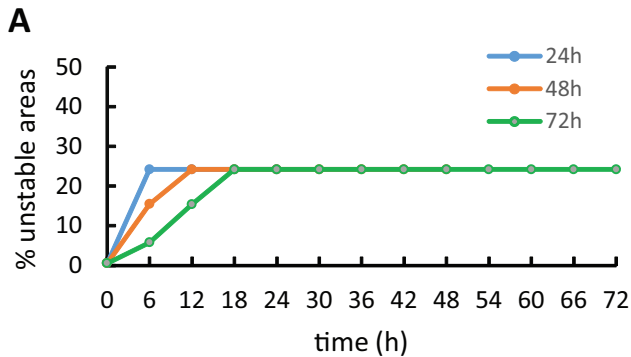
Table 1: Input parameters for SHALSTAB and TRIGRS (scenario 1: ash layer, unit III). **Z:** tephra fallout deposit thickness. K_{sat} was taken from literature. D_0 was computed referring to the procedure reported by Rossi et al. (2013). $T = K_{sat} * Z$. φ´ **and** c´ were obtained from standard direct shear tests in the geotechnical laboratory of University of Salerno. γ_{sat} was investigated in the laboratory. **α, θr and θs** were estimated through analytical procedure.

Z	Ks	Do (1)	T	φ´	Ɣs	c´	θs	θr	α
(m)	(m/s)	(m2/s)	(m2/s)	(deg)	kN/m3	(kPa)	ths	thr	alp
0.15	5 x 10-5	5 x 10-5	7.5 x 10-6	38.4	15.6	0.5	0.62	0.06	0.08

Table 2: Input parameters for SHALSTAB and TRIGRS (scenario 2: lapilli layer, units I and II) Ksat was measured in the field. D0 was computed referring to the procedure reported by Rossi et al. (2013). T. φ´ and c´ were obtained from standard direct shear tests in the geotechnical laboratory of University of Salerno. γsat was investigated in the laboratory.

Z	Ks	Do (1)	T	φ´	Ɣs	c´
(m)	(m/s)	(m2/s)	(m2/s)	(deg)	kN/m3	(kPa)
0.1	3.9 x 10-2	1.8 x 10-2	5.8 x 10-3	53	13	0

Table 3: Rainfall intensities and duration used in TRIGRS simulations. We assumed that a constant rainfall flux of 3.3 mm/h for 24 hours represent a high intensity rainfall. As well, a medium intensity of 1.6 mm/h and a low intensity of 0.83 mm/day were considered for 48 hours and 72 hours, respectively. For forested area we considered a rainfall interception of 25% and calculated the intensities according to this.

Duration	Rainfall intensity	
	bare soil	forested
h	mm/h	mm/h
24	3.30	2.50
48	1.60	1.25
72	0.83	0.63

Table 4: Percent of catchment area in each critical rainfall range for scenario 1 with SHALSTAB for the upper catchments Colorado and Florencia River

basin				Colorado	Florencia	Colorado	Florencia
cohesion				0kPa	0kPa	0.5kPa	0.5kPa
class	**log (q/T)**	**T**	**q**	**area**	**area**	**area**	**area**
		(m²/day)	(mm/day)	%	%	%	%
Un. unstable	-10	0.648	0.3	7.3	21.6	0.5	0.3
unstable	<-3.1	0.648	0.5	24.8	28.7	2.3	6.1
unstable	-2.8	0.648	1.0	19.6	14.9	3.2	8.4
unstable	-2.5	0.648	2.0	17.4	16.1	6.9	16.8
unstable	-2.2	0.648	4.1	8.7	10.3	6.5	21.3
stable	>1.9	0.648	8.2	4.3	5.7	4.6	14.4
Un. Stable	10	0.648		18.0	2.9	76.0	32.6

T: Transmissivity; q: critical rainfall; Un.: unconditionally

Table 5: Summary of unstable percentage area for Colorado River and Florencia River upper catchment calculated with TRIGRS model for scenario 1 and 2, for 3 rainfall intensities and duration (3.3, 24h; 1.6, 48h; 0.83 mm/h, 72h) and two different tephra deposit thicknesses for each scenario.

model	percentage of unstable area			
TRIGRS	C (15cm)	F (15cm)	C (30 cm)	F (30cm)
S1, 24h	24.2%	68.0%	24.2%	68.0%
S1, 48h	24.2%	68.0%	24.2%	68.0%
S1, 72h	24.2%	68.0%	24.2%	68.0%
TRIGRS	C (10cm)	F (10 cm)	C (20cm)	F (20cm)
S2, 24h	11.9%	44.6%	4.0%	11.0%
S2, 48h	6.5%	23.6%	2.9%	6.3%
S2, 72h	3.6%	9.2%	2.1%	3.7%

C: Colorado River upper catchment; F: Florencia River upper catchment; (15 cm): tephra deposit thickness; S1: scenario 1; s2: scenario 2.

Table 6: Performance evaluation for results from two scenario and three rainfall intensity and duration for TRIGS model and one scenario with and without cohesion for SHALSTAB model, based on Jaccard fit, model precision and model sensitivity.

Model and scenario	Validation metrics	Bayesan metrics	
	Jaccard fit (R_J)	Precision (R_{MP})	Sensitivity (R_{MS})
TRIGRS S1, 24h	47.1%	56.9%	73.2%
TRIGRS S1, 48h	47.1%	56.9%	73.2%
TRIGRS S1, 72h	47.1%	56.9%	73.2%
TRIGRS S2, 24h	40.5%	62.9%	53.1%
TRIGRS S2, 48h	23.4%	61.3%	27.4%
TRIGRS S2, 72h	9.0%	55.7%	9.7%
SHALSTAB S1	48.0%	51.2%	88.6%
SHALSTAB S1 c	35.7%	52.5%	52.7%

S1: scenario 1; S2: scenario 2; c: cohesion; (R_{MP}): Model Precision; (R_{MS}): Model sensitivity

Table 7: Summary of unstable areas percent and volume for open land (bare soil) located in the Florencia and Colorado upper basin, for 2 scenarios with the models SHALSTAB and TRIGRS.

	% of instable area					
	Florencia UC			Colorado UC		
	SHALSTAB	TRIGRS		SHALSTAB	TRIGRS	
scenario	S1	S1	S2	S1	S1	S2
Bare soil Hillslope	91.0	73.6	55.3	77.9	27.2	12.9
volume m^3	33180.0	26970.0	13770.0	160140.0	44010.0	15760.0

S1: scenario 1; S2: scenario 2; UC: upper catchment;