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1 **Simplified simulation of rock avalanches and**
2 **subsequent debris flows with a single thin-layer**
3 **model. Application to the Prêcheur river (Martinique,**
4 **Lesser Antilles)**

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19 ABSTRACT

High discharge debris flows in mountainous and volcanic areas are major threats to populations and infrastructures. Modeling such events is challenging because the associated processes are complex, and because we often lack data to constrain rheological parameters. In this work, we show how the extensive use of field data can help model a rock avalanche, and the subsequent remobilization of the deposits as a high discharge debris flow, with a single one-phase thin-layer numerical code, SHALTOP, and up to two rheological parameters. With the Prêcheur river catchment (Martinique, Lesser Antilles) as a case study, we use geological and geomorphological data, topographic surveys, seismic recordings and granulometric analyses to define realistic simulation scenarios and determine the main characteristics of documented events for model calibration. Then, we model a possible $1.9 \times 10^6 \text{ m}^3$ rock avalanche. The resulting deposits are remobilized instantaneously as a high discharge debris flow. We show that, for a given unstable volume, successive collapses allow to better reproduce the dynamics of the rock avalanche, but do not change the geometry of the final deposits, and thus the initial conditions of the subsequent debris flow simulation. The location of the debris flow initiation has also little influence on simulation results. However, progressive remobilization of materials slows down the debris flow and limits overflows, in comparison to an instantaneous release. Nevertheless, high discharge debris flows are well reproduced with an instantaneous initiation. Besides, the range of travel times measured for other significant debris flows in the Prêcheur river is consistent with our simulation results.

Keywords: landslide, lahar, debris flow, modeling, thin-layer

20 Highlights

- 21 • Successful mass flow simulations with up to two rheological parameters
- 22 • Extensive use of field data for model calibration and scenario definition
- 23 • Mapping of areas exposed to high discharge debris flow, for hazard assessment

24 1 INTRODUCTION

25 The remobilization by water of old or recent volcanic materials, during or even long after an eruption, generates sediment-laden
26 flows called lahars that travel in ravines and rivers tens to hundreds of kilometers away from the volcano (Vallance and Iverson,
27 2015; Thouret et al., 2020). Thus, they can be major threats to populations and infrastructures. Non-eruptive lahars can be
28 correlated to landslides that create loose debris reservoirs. Numerical simulations considering both the landslide that creates the
29 reservoir and its remobilization as lahars can improve hazard assessment. However, the modeling process is not straight-forward
30 because the initial landslide and the subsequent lahar are two different phenomena.

31 The initial landslide can take various forms, as water-laden debris avalanches or dry rock avalanches (Hungr et al., 2014).
32 In a first approximation, the physical and rheological properties of materials (such as density or basal friction coefficient)
33 can be considered homogeneous both in space and time, which simplifies the quantification of the propagation (McDougall
34 and Hungr, 2005). In comparison, the subsequent lahars are more complex: they can propagate as hyperconcentrated flows
35 (HFs) or debris flows (DFs). In the following, we will thus talk about lahars to refer to both DFs and HFs. Following (Cousot
36 and Meunier, 1996; Vallance and Iverson, 2015; Thouret et al., 2020), we define DFs as homogeneous mixtures of water and

37 granular rock material with volumetric solid fraction higher than 60%, similar velocities for the solid and fluid phases and
38 densities above 1800 kg m^{-3} . HF's feature solid fractions between 20% and 60%, a vertical separation of the two phases and
39 densities below 1800 kg m^{-3} . We may expect that the remobilization of a small amount of solid materials will produce HF's,
40 while fast remobilization by liquefaction of a large debris reservoir will turn into a DF (Vallance and Iverson, 2015). However,
41 a DF initiated in the upper section of a river may well turn into HF at its tail because of dilution and settling, while its front
42 increases its solid content due to bed erosion. Further dilution downstream can then transform completely the DF into a HF (for
43 a conceptual view of such a process, see Figure 2 in Thouret et al., 2020).

44 The combined effects of particle collision and friction, lubrication, advection and suspension in presence of an interstitial
45 fluid, are difficult to model in a single framework (Andreotti et al., 2013; Delannay et al., 2017). Thus, current solutions where
46 the dynamics of elementary volumes of fluid and/or of each solid particle are considered (in 2 or 3 dimensions) often focus
47 on reproducing some of the physical processes, but never all of them. Discrete element modeling (DEM) is now widely used
48 to model dry and wet granular flows at the laboratory scale (Durán et al., 2012; Lefebvre-Lepot et al., 2015; Windows-Yule
49 et al., 2016, e.g.). Applications to field scale simulations are given for instance by (Zhao and Shan, 2013) and Leonardi et al.
50 (2014) for DF's, and by Yan et al. (2020) and Wu and Hsieh (2021) for rock avalanches. Another approach is to consider a
51 single-phase flow and solve the Navier-Stokes equations (e.g. Hu et al., 2015). However, both DEM and continuous models
52 often require huge computing resources and/or depend on too many user-defined parameters, which is incompatible with the
53 limited knowledge of the flowing material we have in practice.

54 Over the past decades, thin-layer models have been increasingly used to study debris and rock avalanches, as well as lahars
55 (see McDougall (2017) for a general review, and Thouret et al. (2020) for lahar modeling). Their main assumption is that the
56 landslide thickness is negligible in comparison to its length. In turn, flow description is reduced to flow thickness and flow
57 thickness-averaged velocity, which simplifies greatly the governing equations in comparison to 3D models. In their simplest
58 form, thin-layer models describe an homogeneous flow and dissipate energy solely by considering a stress applied at the base
59 of the flow. For instance, with the Coulomb rheology the only rheological parameter is the friction coefficient $\mu_S = \tan(\delta)$,
60 with δ the friction angle. If the topographic slope θ is higher than δ the flow accelerates, and decelerates and stops otherwise
61 (inertial effects and spatial variations in flow thickness may change temporarily this first-order behavior). Such models proved
62 to reproduce well rock and debris avalanches as well as debris flows (Hungri et al., 2007; Pirulli and Mangeney, 2008; Favreau
63 et al., 2010; Lucas et al., 2014; Pastor et al., 2018a). More elaborate numerical codes also model, for instance, two-phase flows
64 (Iverson and George, 2014; Bouchut et al., 2015, 2016; Mergili et al., 2017; Pastor et al., 2018b), three-phase flows (fluid,
65 coarse solid fraction, fine solid fraction, Pudasaini and Mergili, 2019), and erosion along flow path (Iverson, 2012; Pirulli and
66 Pastor, 2012). However, these developments often rely on empirical relations (e.g. for erosion laws McDougall, 2017). Besides,
67 thin-layer equations with complex rheologies are mostly derived on simple topographies (e.g. Pastor et al., 2009; Baker et al.,
68 2016), and the lack of analytical solutions makes it difficult to test the robustness of associated numerical tools. Furthermore,
69 although complex rheologies may model more realistic dynamics, they come at the cost of an increased number of parameters,
70 such as erosion rates, erodible thickness, viscosity, drag coefficient or densities of each phase (e.g. George and Iverson, 2014;
71 Mergili et al., 2017). These parameters can be difficult to calibrate if not enough data are available. Besides, when they are not
72 known, the high number of degrees of freedom may artificially improve back-analysis studies.

73 In practice, experts conducting hazard assessment studies may neither have the time nor the financial resources to carry out
74 a thorough analysis with detailed but complex numerical models. The question is: to what extent can we expect realistic results

75 from simple physically based thin-layer models for rock avalanche and DF simulations? The answer strongly depends on the
76 available field data. In this work, we present a modeling approach with empirical but simple rheologies involving no more
77 than two parameters. To enhance the quality of simulation results, we make an extensive use of field data to define realistic
78 simulation scenarios and characterize past events for model calibration. We will use the thin-layer model SHALTOP (Bouchut
79 et al., 2003; Bouchut and Westdickenberg, 2004; Mangeney-Castelnau et al., 2005; Mangeney et al., 2007b), that proved to
80 reproduce accurately analytical solutions for the dam-break problem (Mangeney et al., 2000; Lucas et al., 2007), and was used
81 successfully to model gravitational flows at the field scale with a simple Coulomb friction law (e.g. Favreau et al., 2010; Lucas
82 et al., 2014; Moretti et al., 2012, 2015; Peruzzetto et al., 2019; Moretti et al., 2020). In comparison to other thin-layer models,
83 SHALTOP also takes into account precisely topography curvature effects that can be significant for fast gravity driven flows
84 (Peruzzetto et al., 2021).

85 Because they have the highest potential impact on infrastructures and populations, we focus on extreme events (avalanches
86 of volumes $> 1 \times 10^6 \text{ m}^3$, and high discharge DFs). We choose the Prêcheur river in Martinique island (Lesser Antilles, French
87 Caribbean) as study site (Figure 1), where such events are documented and where stakes are high, as large DFs threaten the
88 Prêcheur village at the mouth of the river (Figure 2). In a first calibration step, we will use topographic surveys and aerial
89 photographs to construct the initial conditions of (i) a rock avalanche that occurred in 2018 and (ii) a major debris flow that
90 occurred in 2010. Granulometric data help choosing the rheological law, and a range of possible rheological parameters is
91 identified in the literature (see Table 1). By reproducing the travel distance and main dynamic characteristics of the rock
92 avalanche, and the flooded area and travel time of the DF (deduced from aerial photographs and seismic recordings in both
93 cases), we calibrate more precisely rheological parameters. With these fine-tuned parameters, we can then consider the
94 forward prediction of a rock avalanche simulation, whose initial conditions are deduced from geomorphological and geological
95 observations. The resulting deposits are then remobilized instantaneously in another simulation to model the propagation of a
96 high discharge DF. Because in the Prêcheur river rock avalanches do not, in general, transform directly into DFs (Aubaud et al.,
97 2013), we do not consider such a continuous transition in this work.

98 In Section 2 we present in more details our study site, along with the data used to construct simulation scenarios and
99 calibrate our model. Simulation scenarios used for model calibration and forward prediction are presented in Section 3, and the
100 numerical model SHALTOP is detailed in Section 4. Simulation results are then given in Section 5. In Section 6, we investigate
101 the influence of initiation mechanism on simulation results. The latter are discussed in Section 7.

102 **2 DATA**

103 In this section, we present the geological and geomorphological context of our study site, along with the data used to define
104 simulation scenarios. Topographic surveys will be used to define the bed topography and initial volumes. To calibrate the
105 numerical model, we use aerial photographs that give the travel distances and flooded areas of past events. Seismic recordings
106 are used to estimate flow velocity and duration. The granulometry of deposits is also used to choose the rheology in DF
107 simulations. These data are summarized in Table 1.

108 **2.1 Geological context**

109 The Prêcheur river catchment drains part of the western side of Montagne Pelée volcano (Figure 1a). The Samperre cliff is
110 located about 2 km north-west of the volcano summit, at the source of the Samperre river (Figure 2a). Over the past 40 years,

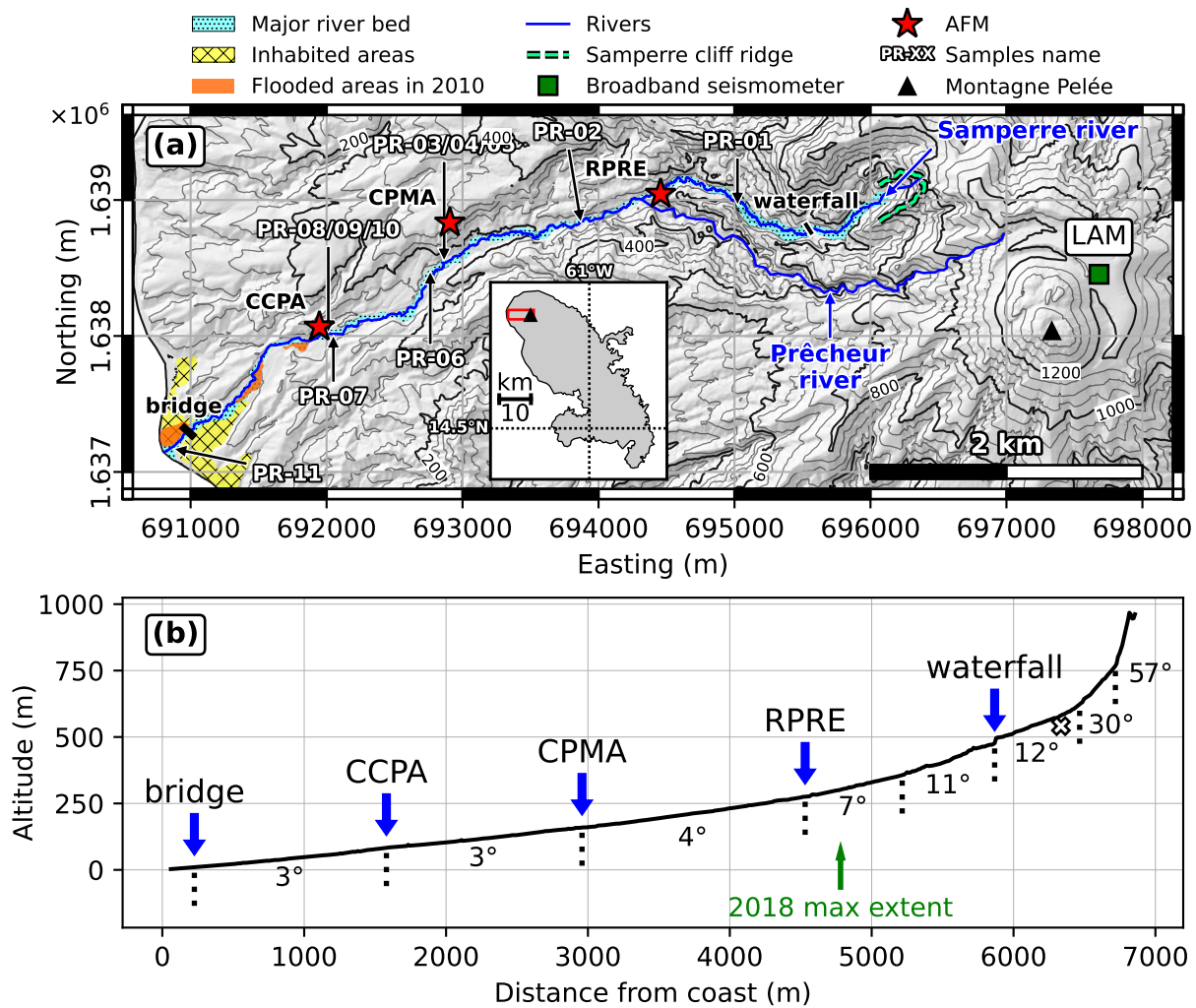


Figure 1. Prêcheur river (Martinique island, Lesser Antilles, French Caribbean) map and section. (a) Map of the Prêcheur river. The insert features the Martinique island, with the red rectangle matching the extent of the map. The 1 m DEM in the river area is from Helimap 08/2018, and from IGN 03/2010 elsewhere. Sampling locations for granulometry analysis are given by black arrows, with corresponding sample names. CCPA, CPMA, RPRE and LAM are the names of AFMs (Acoustic Flow Monitoring) and seismic stations used in this study. Coordinates: WGS84 UTM20N. (b) River cross-section, from the river mouth (left) to the Samperre cliff (right). Green arrow: estimated deposits extents after the 2018 Samperre rock avalanche. White cross: source area for debris flow simulations with imposed discharge (see Sections 6.2 and 6.3). Average slopes are given for each section between dotted vertical black lines. Horizontal and vertical scales differ.



Figure 2. 2018 views of the Samperre cliff and Prêcheur village. (a) Feb. 2, 2018 view of the cliff, after the main rock avalanche of Jan. 4, 2018. The dust cloud generated by a minor collapse is visible on the right side of the cliff. The scree reservoir is highlighted by the black dotted contour. (b) Mar. 30, 2018 helicopter view of the Prêcheur village, constructed on the alluvial fan of the Prêcheur river, with a central view of the bridge.

Table 1. Main characteristics of simulations, derived from literature (citations) and field data (bold).

	Rock avalanche simulation	Debris flow simulation
Topography	08/2018 1-m DEM (modified locally in the Samperre cliff area)	
Initial volume geometry	07/2010, 01/2018, 08/2018 1-m DEMs manually modified following - cliff rim evolution (ORTHO GéoMartinique DEAL February 2007) - geological / geomorphological observations (Nachbaur et al., 2019)	Difference between 01/2018 and 08/2018 1-m DEMs
Rheology choice	Coulomb (e.g. Favreau et al., 2010 ; Lucas et al., 2014 ; Yamada et al., 2018)	Frictional rheology (granulometry of deposits), with Coulomb (Moretti et al., 2015) and Voellmy (McDougall, 2017 ; Zimmermann et al., 2020)
Range of rheological parameters for calibration	$\tan(\delta) \in [\tan(10^\circ), \tan(20^\circ)] = [0.18, 0.36]$ (Lucas et al., 2014 ; Peruzzetto et al., 2019)	$\tan(\delta) \in [\tan(2^\circ), \tan(3^\circ)]$ (riverbed slope at the river mouth) $\xi \in [100 \text{ m s}^{-2}, 500 \text{ m s}^{-2}]$ (Zimmermann et al., 2020)
Calibration data	Travel distance (aerial reconnaissance) Duration and dynamics (seismic signal)	Flooded area (aerial reconnaissance) Travel time (AFMs)

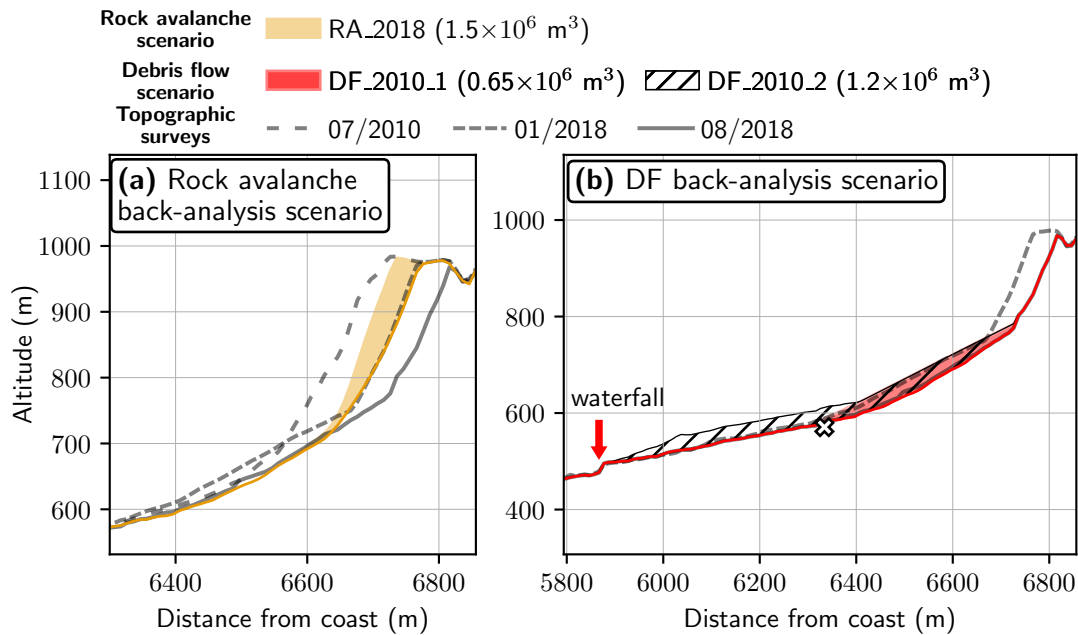


Figure 3. Samperre cliff longitudinal cross-section with topographic surveys and initial mass for calibration scenarios.

(a) Successive topographic surveys (gray lines). Orange patch: collapsing volume reconstructed for the *RA_2018* rock avalanche scenario. Orange line: topography in simulation. (b) Initial reservoirs for the *DF_2010_1* and *DF_2010_2* debris flow simulations. White cross: source area for simulation with imposed discharge (see Sections 6.2 and 6.3). Red line: topography in simulation.

111 the Samperre cliff has produced at least 4 episodes of massive destabilizations in 1980, 1997-1998 (Aubaud et al., 2013),
 112 2009-2011 ($2.1 \times 10^6 \text{ m}^3$, Clouard et al., 2013) and 2018-2019 ($5 \times 10^6 \text{ m}^3$, Queff  l  an, 2018; Nachbaur et al., 2019). However,
 113 another collapse episode is inferred from testimonies in the early 1950s (Aubaud et al., 2013). Thus, the cliff rim retreated
 114 by 250 m between 1988 and 2018 (Nachbaur et al., 2019). Its evolution between March 2010 and August 2018 is given in
 115 Figure 3a (grey lines).

116 A geological interpretation of the cliff main units is given by Nachbaur et al. (2019) and reproduced in Figure 4a. We will
 117 use this interpretation to constrain a potential future cliff collapse (see Table 1 and Table 2). Previous studies (Mathon and
 118 Barras, 2010; Clouard et al., 2013; Nachbaur et al., 2019) identified a stable basal layer progressively exposed by successive
 119 collapses. This basal layer is composed of old indured volcanic deposits emplaced or exposed during a massive flank collapse
 120 216 kyrs ago (D1 event Le Friant et al., 2003; Boudon et al., 2007; Germa et al., 2011; Brunet et al., 2017), and of old pyroclastic
 121 deposits (red and orange patches in Figure 4a respectively). Most of the upper part of the cliff, which collapsed during the 2010
 122 and 2018 destabilization crisis, is constituted of a 100 to 200 meter succession of more recent pyroclastic deposits (Figure 4a,
 123 pink patch). The interface with the basal stable layer is marked by a clear slope break, as well as several water seepages
 124 (Nachbaur et al., 2019).

125 The Samperre river has its source at the cliff toe. About 2.5 km downstream, it joins the Pr  cheur river (Figure 1). In this
 126 upper section, the Samperre river is very narrow (down to 10 m) and steep-walled (the gully is more than 70 m deep at some
 127 locations). Slopes reach up to 30° at the cliff bottom (Figure 1b), which favors the remobilization of rock avalanche deposits.

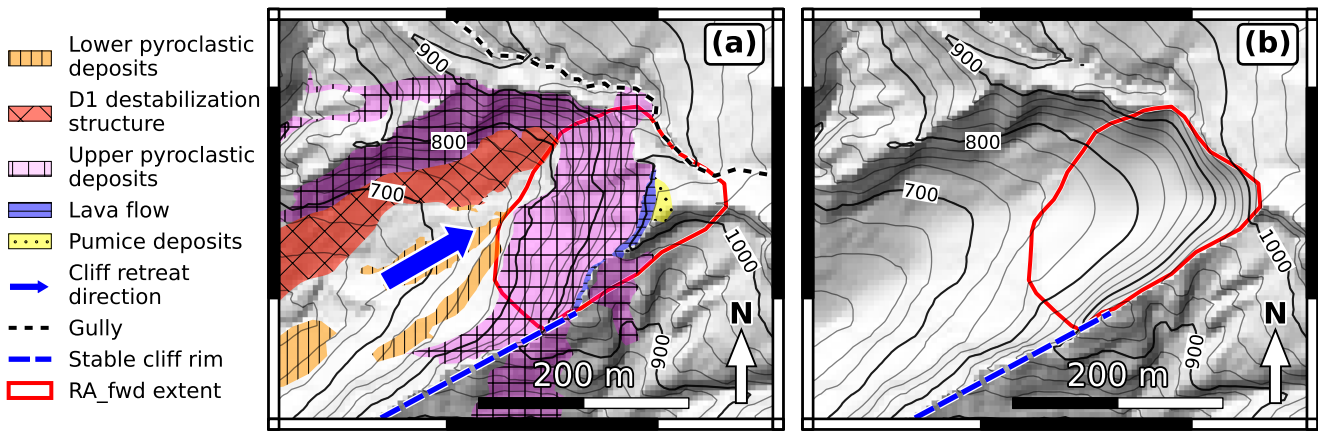


Figure 4. Samperre cliff geology and *RA_fwd* forward prediction scenario. (a) Cliff topography in August 2018 with main geological units (Nachbaur et al., 2019). Vegetation and screes are not displayed. (b) Modified 08/2018 topography with the scar from the potential rock avalanche (*RA_fwd* scenario). The unstable volume is $1.9 \times 10^6 \text{ m}^3$.

128 Supposedly, the most powerful DFs are thus generated in this part of the river. Further downstream, down to RPRE, average
 129 slopes are between 7° and 12° .

130 In the second section of the river, from the Samperre river / Prêcheur river junction down to the river mouth, the river
 131 cuts through relatively poorly resistant materials, such as pumice deposits (Meunier, 1999; Quefféléan, 2018). The river bed
 132 progressively widens (from 30 m to 60 or 70 m) and flattens, with 3° to 4° slopes. Thus, it is mainly a deposition area for DFs,
 133 with meter-sized blocks scattered over the river bed.

134 At the mouth of the river, 7 km downstream the Samperre cliff, the Prêcheur village (Figure 2b) is built on the alluvial fan
 135 and hosted 1300 inhabitants in 2017 (INSEE, 2020). The bridge (Figure 2b) is the only access to the northern part of the village.

136 2.2 Topographic surveys and aerial photographs

137 The main source of quantitative data to constrain initial conditions in simulations are topographic surveys (Table 1 and Table 2).
 138 We use three different Digital Elevation Models (DEMs):

- 139 • 07/2010 DEM: a 1-m DEM derived from a LiDAR acquisition over the whole river after the main rock avalanches and
 140 DFs of 2010. Unfortunately, as the river is rather narrow in its upper section, its quality is rather poor from the cliff
 141 bottom down to RPRE.
- 142 • 01/2018 DEM: A photogrammetric model of the Samperre cliff was constructed from aerial photographs taken by a
 143 drone on Jan. 19, 2018, from which a 1-m DEM of the cliff (which is deprived of vegetation) could be derived.
- 144 • 08/2018 DEM: A 1-m DEM derived from a LiDAR acquisition over the whole river. We only modify it slightly at the
 145 bottom of the cliff to remove patches of screes, that would otherwise lead to incorrect scree reservoir reconstruction for
 146 DF simulation (see Section 3.1.2). This is done in a similar manner to pre-collapse topography and scar reconstruction in
 147 (Guimpier et al., 2021). Screes are identified thanks to slope breaks and slope direction variations at the bottom of the
 148 cliff, and are then removed by modifying manually the 5 m contour lines of the 08/2018 DEM, using contour lines trends
 149 where the cliff is deprived of screes (see Supplementary Figure 1).

150 Simulations are mainly carried out on the 08/2018 DEM, which has the best quality and is deprived of vegetation. Along
151 with topographic surveys, we also use orthophotographs and aerial photographs taken during helicopter overflights: they help
152 quantifying the cliff evolution in between topographic surveys, as well as the travel distance of rock avalanches and flooded
153 areas after DFs.

154 **2.3 River and cliff monitoring**

155 Since 1975, the occurrence and relative magnitude of collapse events is systematically inferred from the seismic network
156 maintained by the Observatoire Volcanologique et Sismologique de Martinique (OVSM) (Aubaud et al., 2013; OVSM-IPGP,
157 2020). In this work, we use the broad-band CMG-40T seismic sensor (60 s - 50 Hz), located on the north-eastern side of
158 the Montagne Pelée, about 1.5 km away from the Samperre cliff (LAM station in Figure 1a). Assuming the duration of rock
159 avalanches can be approximated by the duration of seismic signals (Hibert et al., 2011; Levy et al., 2015), seismic recordings
160 give a first insight on the rock avalanche dynamics.

161 In 1998, 2001 and 2014, three geophones, called Acoustic Flow Monitoring (AFM) sensors (LaHusen, 2005), were installed
162 by the OVSM/IPGP along the river (at CPMA, RPRE and CCPA respectively, see Figure 1a). The AFM system, developed
163 at the Cascades Volcano Observatory (LaHusen, 1998) is the most common system for lahar monitoring, and can be used to
164 trigger alarms. It is currently installed on active volcanoes (e.g. Pinatubo, Marcial et al., 1996, Merapi, Lavigne et al., 2000,
165 Ruapehu, Cole et al., 2009 and Tungurahua, Jones et al., 2015). In this study, we use the so-called FULL channel (signal in
166 10-300 Hz frequency band, low gain) to estimate the DF travel duration between RPRE and CCPA. Values span between 0 and
167 4000 mV, but are usually below 50 mV in normal streamflow conditions. In 2010, sampling interval was 10 min and 5 min in
168 normal conditions for CPMA and RPRE respectively, but was reduced to 1 min when the HILO (high gain, low pass) channel
169 exceeded 500 mV at CPMA and 1000 mV at RPRE.

170 **2.4 Granulometry of lahar deposits**

171 11 samples (PR-01 to PR-11) of lahars deposits were recovered for granulometry analysis, at 5 sites along the river, from its
172 outlet to about 5.5 km upstream (Figure 1a). To our knowledge, it is the first time such a sampling campaign is carried out in the
173 Prêcheur river: Meunier (1999) only analyzed the granulometry of streamflow deposits at the river mouth, and Lalubie (2013)
174 similarly recovered one sample only at 80 m altitude (presumably near the CCPA station). More generally, on-site sampling
175 is rarely carried out to constrain numerical simulations. Although they can hardly be used directly to calibrate simulation
176 parameters, they help understand the physical processes controlling flow dynamics.

177 Granulometric curves as well as an example of a sampling site are presented in Figure 5. All samples contain mainly sand,
178 gravel and boulders, with less than 4% of silts and clays (diameter $d < 0.1$ mm). When compared to granulometric envelopes
179 derived by Bardou et al. (2003) in alpine context, our samples fit neither the "friction-viscous" nor the "viscoplastic" envelopes,
180 whose fine fraction is more important (between 5% and 20% of clay, Figure 5a). Our results are more consistent with grading
181 ranges of lahars deposits on Semeru volcano in Java, Indonesia (Dumaisnil et al., 2010), in particular for hyper-concentrated
182 flow and granular flow deposits (Figure 5b). In their study, granular flows should be understood as DFs with only little silts and
183 clays, such that collision and friction between grains are the main driving forces.

184 The distinction between DF deposits and HF deposits is not easy as each one can evolve into the other one. Following
185 Dumaisnil et al. (2010) we can associate finer grading (mainly sand and gravel) to HFs (as for sampling sites PR-02, PR-07,
186 PR-11 and PR-10) and coarser, unsorted deposits to DFs (as for sampling sites PR-01 and PR-08).

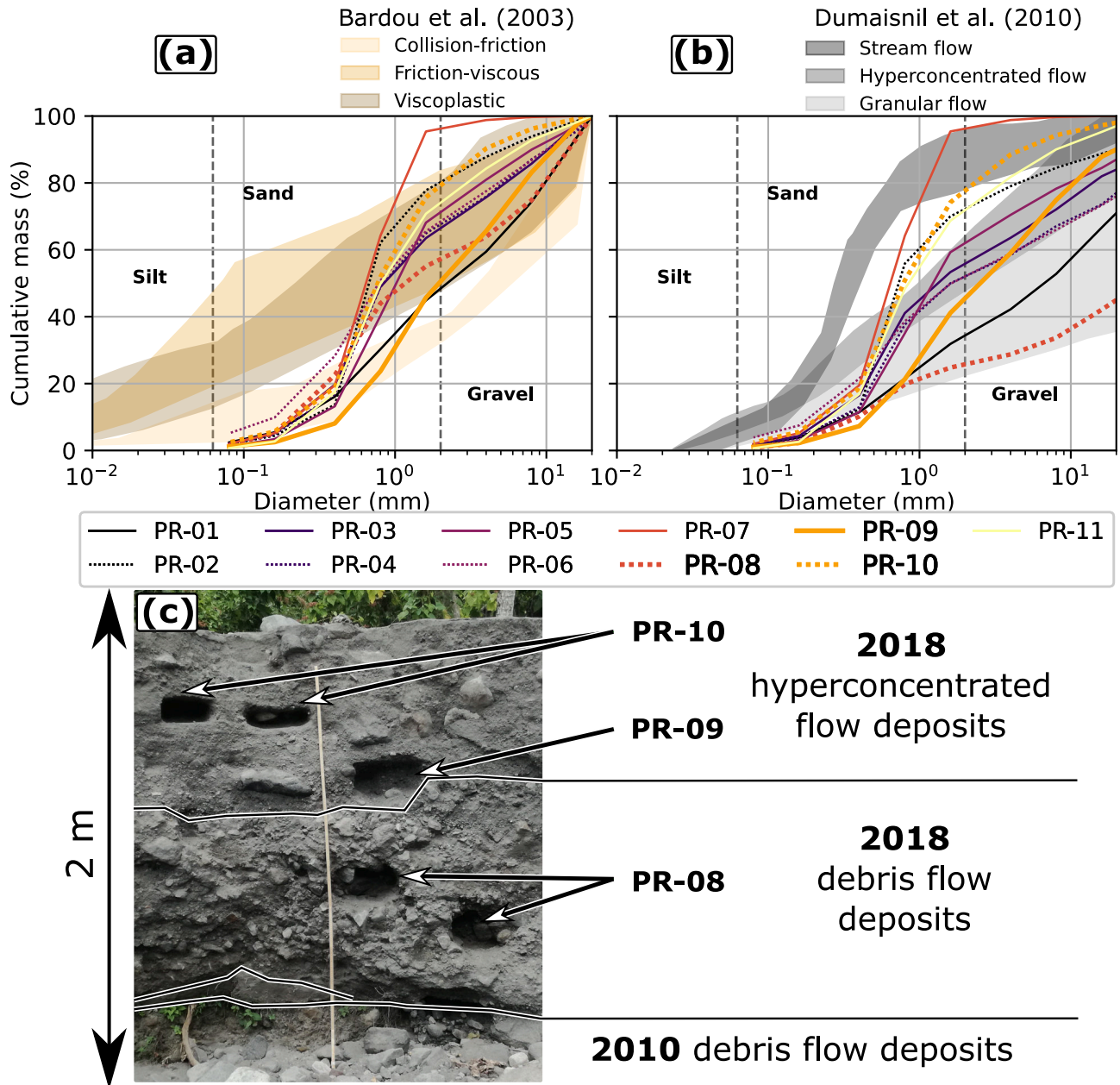


Figure 5. Granulometry of lahar deposits. (a) Lines: granulometry of samples, with boulders larger than 2 cm removed. Colored patches: granulometric envelopes from [Bardou et al. \(2003\)](#) associated to flow rheologies, in alpine context. (b) Lines: granulometry of the whole samples. Grey patches: granulometric envelopes from [Dumaisnil et al. \(2010\)](#), for lahar deposits on the Semeru volcano, Indonesia. (c) Example of sampling site. Granulometric curves of the samples are given in bold in (a) and (b). See [Figure 1a](#) for the location of sampling sites (PR-01 is the most upstream sample, and numbering follows stream direction).

3 SIMULATION SCENARIOS FOR CALIBRATION AND FORWARD PREDICTION

We focus on the modeling of extreme events: rock avalanches with volumes above $1 \times 10^6 \text{ m}^3$ and high discharge DFs. In the following we present two such events and explain how we construct the topography and initial volumes for model calibration. This is summarized in the first three columns of Table 2.

3.1 Model calibration: events description and simulation initial conditions

3.1.1 Jan. 4, 2018 rock avalanche

A major episode of destabilization occurred in 2018-2019 on the Samperre cliff. It started on Jan. 2, 2018, after a particularly rainy wet season. Its main phase lasted about two months, but episodic gravitational readjustments occurred until October 2019. This crisis culminated quickly after it started, on Jan. 4, 2018, with one main rock avalanche at 03:00 UTC. It was recorded widely on the seismic network and lasted about 2 minutes (Figure 6). From helicopter overflight, it is estimated to have reached the river bend just upstream RPRE (Figure 1b, green arrow). The 01/2018 DEM gives the geometry of the cliff after the main destabilizations. However, the previous topographic survey, the 07/2010 DEM, is too old to be used as a pre-collapse topography. Indeed, diachronic analysis of ortho-photographs show that the cliff rim retreated by about 50 m between 2010 and 2017 (Nachbaur et al., 2019).

Thus, in order to define the unstable volume involved in the Jan. 4, 2018 destabilization, we use the cliff rim position observed on February 2017 orthophotographs and reconstruct a synthetic cliff topography, as it may have been just before the 2018 destabilization crisis (Figure 3a). This is done by defining a set of longitudinal and transverse cross-sections on the 07/2010 DEM, changing the corresponding profiles with cubic splines, and interpolating the DEM in between, to finally reconstruct the cliff edge as it was in February 2017 (see Supplementary Figure 2).

The post-collapse topography is given by the 01/2018 DEM for the cliff, and by the 08/2018 DEM for the cliff bottom (as deposits of the Jan. 4, 2018 rock avalanche are included in the 01/DEM but had been washed away by August 2018, see Figure 3a). The $1.5 \times 10^6 \text{ m}^3$ unstable volume is then defined as the difference between these two reconstructed topographies. This is our *RA_2018* scenario (Figure 3a, orange patch).

Though the volumes involved in the rock avalanches in 2018-2019 had been the most important since at least 1980, the scree reservoir at the bottom of the cliff was remobilized progressively. Thus, no DF was powerful enough to leave the river bed. In comparison, the DF that occurred on Jun. 19, 2010 flooded the Prêcheur village. In order to have a risk conservative approach and investigate worst-case scenarios, DF modeling will be calibrated on this latter event.

3.1.2 Jun. 19, 2010 debris flow

In May 2010, a series of destabilizations occurred on the Samperre cliff, involving about $2.1 \times 10^6 \text{ m}^3$ (Clouard et al., 2013). After its main phase on May 11, 2010, the first lahar occurred on May, 14 (Aubaud et al., 2013). On Jun. 19, at 7:30 UTM and after a non exceptional tropical wave, a high discharge DF flooded the Abymes quarter in the Prêcheur village.

AFMs records enable the identification of two initial relatively small amplitude surges, with the main phase (that we try to model) occurring between 08:30 and 09:00 UTM (Figure 7a and 7b). The 3000 mV peak value registered at CPMA is particularly high: in all the other lahars from 2009 and 2010, it exceeded 1000 mV on a few occasions only. The signal amplitude then progressively decreased until 11:00 UTM. A last small surge can be spotted at 11:30 UTM, lasting about 30 min (Figure 7a). As pointed out by Aubaud et al. (2013), the triggering rainfall was not particularly strong (11 mm in 1h40), but

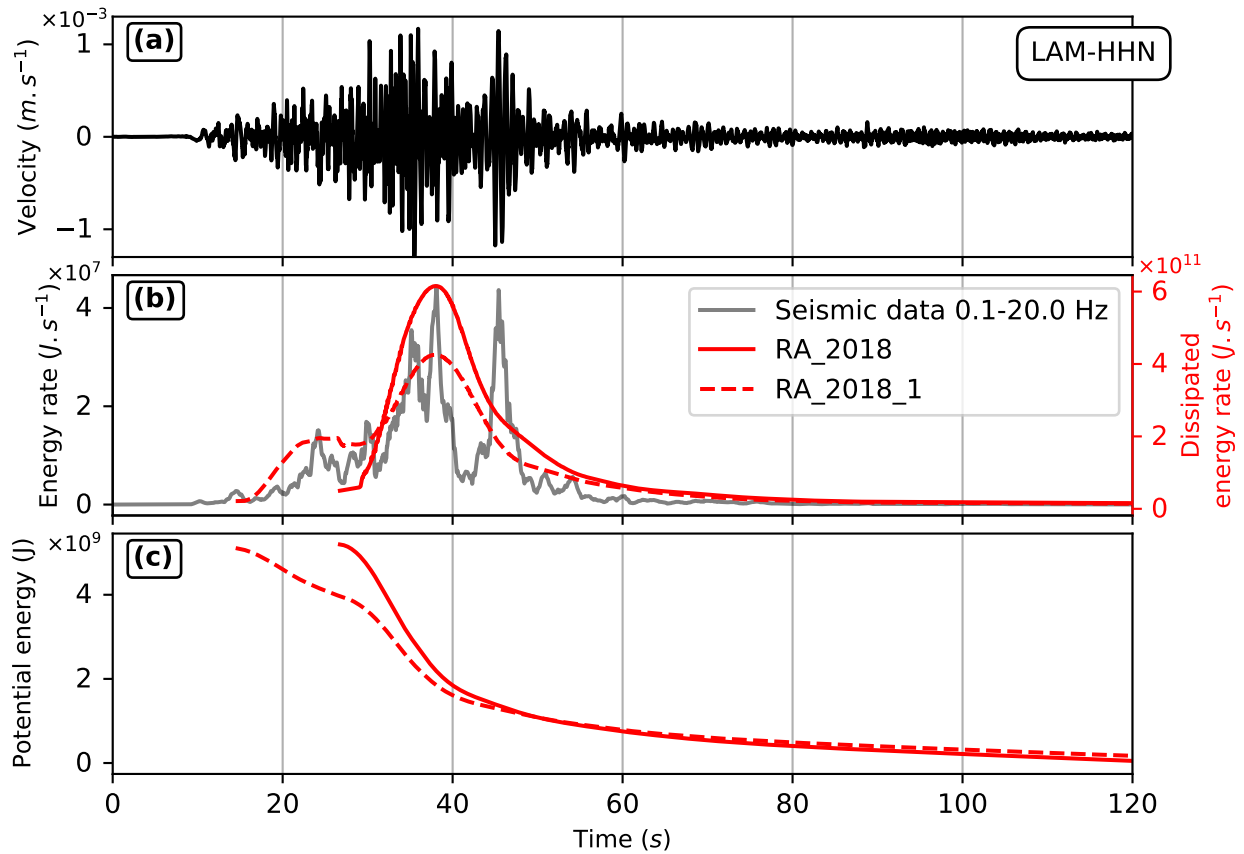


Figure 6. Seismic recordings of the Jan. 4, 2018 Samperre rock avalanche. (a) Signal recorded at station LAM, horizontal northern component. $t = 0$ is 03:00 UTC, Jan. 4, 2018. Signal is filtered between 0.1 and 20 Hz. (b) Grey line: Seismic energy rate at station LAM. Red lines: energy dissipated during the *RA_2018* and *RA_2018_1* scenarios (plain and dashed lines, respectively), with friction coefficient $\mu_s = \tan(14^\circ) = 0.25$. Grey and red lines are aligned for their maximums to match. See Supplementary Note 1 for details on energy computation. (c) Potential energy of the simulated rock avalanche in scenarios *RA_2018* and *RA_2018_1*.

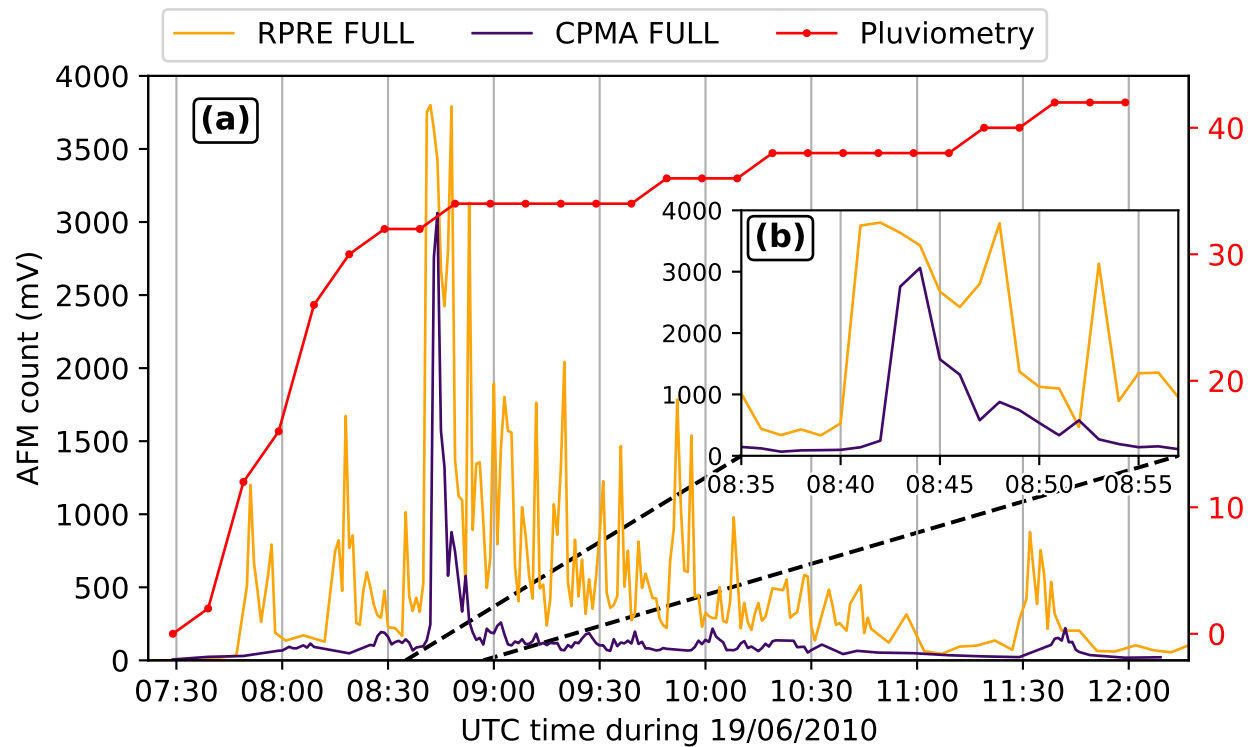


Figure 7. AFMs recordings of the Jun. 19, 2010 lahar from RPRE and CPMA FULL channel. (a) Full event recording, with cumulated pluviometry recorded in CPMA (red line with dots). The black dashed line locates the main event plotted in (b). (b) Main phase of the lahar, with the main DF surge. Time is in hours, UTC.

223 the main surge was preceded by 1 hour long 30 mm precipitations (as recorded in CPMA station, Figure 7a). This surge was
 224 particularly fast: the peak amplitude was recorded with a 2 to 3 min interval between RPRE and CPMA (Figure 7a). Given the
 225 1.5 km distance between the two stations, it yields an average velocity of 30 to 45 km hr⁻¹ (8 to 13 m s⁻¹). The extent and
 226 location of overflows are given in Figure 1a.

227 The Jun. 19 2010 lahar is described as a DF by Mathon and Barras (2010) and Laigle and Macabies (2010). The sample
 228 PR-06 was recovered from deposits that were not present before the 2010 lahars. As the vegetation cover is too important to
 229 have developed after 2018, we associate the sample PR-06 to the deposits of the Jun. 19, 2010 DF. Although it features the
 230 highest fine fraction, it remains low and is similar to other deposits: only 5% of clays and silts within the 20 mm fraction, and
 231 less than 4% of the total flowing sediment. Even if water circulation may have washed away part of the fine fraction since 2010
 232 (Dumaisnil et al., 2010), what must be actually considered is the clay fraction, which will be even less. Following Coussot and
 233 Meunier (1996), we may thus assume that the DF dynamics were controlled by collisional and frictional interactions, and not
 234 viscous forces.

235 The high DF velocity, as well as the screens washout at the cliff toe, suggest it may have been triggered by the instantaneous
 236 or at least very quick remobilization of the scree reservoir, in what Lalubie (2013) called a liquefaction triggered lahar. However,
 237 no topographic data is available to constrain directly the reservoir. On the contrary, the reservoir produced by the first rock
 238 avalanches of the 2018 sequence can be clearly identified on the 01/2018 DEM. Thus, we use the geometry of the 2018 scree
 239 reservoir as a proxy for the reservoir remobilized in 2010.

240 This is done by adjusting a sloping plane on the reservoir surface on the 01/2018 DEM, through a simple Root Mean Square
241 Error (RMSE) minimization between the surface points and a plane, with the CloudCompare software. With a RMSE of 2.1 m,
242 when the reservoir is about 120 m large and 340 m long, the fit is rather good. We assume the 2010 reservoir shared the same
243 characteristics, as the materials involved are similar. The difference between this plane and the 08/2018 DEM provides us with
244 an initial volume of $0.65 \times 10^6 \text{ m}^3$: this is our *DF_2010_1* scenario (Figure 3b, red patch).

245 As the total volume of the rock avalanches in May 2010 is estimated to $2.1 \times 10^6 \text{ m}^3$ (Clouard et al., 2013), we will also
246 consider a larger reservoir (*DF_2010_2* scenario). This is done by filling the main river bed between the bottom of the cliff and
247 the waterfall (600 m downstream, upper estimation of the maximum distance reached by the rock avalanches in 2010) by a
248 30 m thick layer of materials (Figure 3b, black hatches). Such a thickness is indeed consistent with observations made during
249 helicopter flights. We thus create a $1.2 \times 10^6 \text{ m}^3$ reservoir.

250 These simulation scenarios are used to calibrate the model. With the resulting rheological parameters, we will then be able
251 to consider a forward prediction scenario, whose initial conditions are presented in the following section.

252 3.2 Forward-prediction scenario: simulation initial conditions

253 In our forward prediction scenario, we model the propagation of a possible future rock avalanche (*RA_fwd* scenario, see Table 2),
254 and the subsequent instantaneous remobilization of the simulated deposits to produce a DF (*DF_fwd* scenario, see Table 2). We
255 use geological and geomorphological data (see Section 2.1) to constrain the initial unstable mass in the cliff. Following its
256 historical retreat direction (Nachbaur et al., 2019), we infer that the north-west part of the cliff is the most likely candidate for
257 future large collapses (Figure 4a and b, red line). Following Nachbaur et al. (2019), the western limit is constrained by the
258 contact between the unstable upper pyroclastic deposits (Figure 4a, pink patch), and the stable basal units (Figure 4a, orange and
259 red patch). We match the northern extent of the unstable volume with the gully running behind the Samperre cliff (Figure 1a,
260 black dashed line). Finally, for the south-east limit, we extend the actual cliff rim towards the north-east: over the past decades,
261 it has constantly progressed in this direction (Figure 1a and 1b, blue dashed line).

262 Within this extent (Figure 4a and 4b, red line), the topography is modified manually (in the same way as the *2018_1*
263 pre-collapse topography) to get the collapse scar, so that slopes inside and outside the scar are consistent (Figure 4b). The
264 resulting $1.9 \times 10^6 \text{ m}^3$ initial volume of the avalanche is compatible with the volume of previous destabilizations (Clouard
265 et al., 2013; Nachbaur et al., 2019).

266 The deposits of the simulated rock avalanche will then be instantaneously remobilized to model a subsequent high discharge
267 DF. This is done by changing the rheological parameters in simulations. It will be explained in the next section, where we
268 present the SHALTOP numerical code, that we use to model the propagation of rock avalanches and debris flows.

269 4 NUMERICAL MODEL

270 The SHALTOP thin-layer numerical code simulates the dynamics and emplacement of flows on general topographies (Bouchut
271 et al., 2003; Bouchut and Westdickenberg, 2004; Mangeney-Castelnau et al., 2005; Mangeney et al., 2007a). It has been
272 successfully tested to reproduce both real landslide (e.g. Moretti et al., 2015; Brunet et al., 2017; Peruzzetto et al., 2018b) and
273 laboratory experiments (Mangeney-Castelnau et al., 2005; Mangeney et al., 2007a). In SHALTOP, the material layer moving on
274 the topography is considered homogeneous and erosion is not modeled. Energy is dissipated through a force applied at the base
275 of the flow, in the opposite direction to flow velocity. We use the same rheological law in the whole DF, without considering

Table 2. Simulation scenarios for model calibration and forward prediction

	<i>RA_2018</i>	<i>DF_2010_1</i>	<i>DF_2010_2</i>	<i>RA_fwd</i>	<i>DF_fwd</i>
Purpose	Calibration for rock avalanche (Jan. 4, 2018 rock avalanche)	Calibration for debris flow (Jun. 19, 2010 debris flow)		Forward prediction simulation (rock avalanche)	Forward prediction simulation (debris flow)
Bed topography	01/2018 in the cliff sector, 08/2018 DEM elsewhere (modified manually to remove deposits)	08/2018 DEM (modified manually to remove deposits)		08/2018 DEM (modified manually to construct collapse scar)	08/2018 DEM (modified manually to remove deposits)
Initial volume geometry	Difference between 2017 DEM (reconstructed) and 01/2018 DEM	Difference between 01/2018 and 08/2018 DEMs	Difference between 01/2018 and 08/2018 DEMs + 30 m of materials over 600 m downstream	Difference between 08/2018 DEM and synthetic collapse scar	Deposits of the <i>RA_fwd</i> rock avalanche simulation
Volume ($\times 10^6$ m³)	1.5	0.65	1.2	1.9	1.9
Calibrated rheological parameters	$\mu_S = \tan(14^\circ)$	Coulomb: $\mu_S = \tan(2^\circ)$ or $\mu_S = \tan(3^\circ)$ Voellmy: $\mu_S = \tan(2^\circ)$ and $\xi = 500$ m s ⁻²			

276 possible dilution and sediment settling at its tail.

277 We model rock avalanches with the Coulomb rheology, as it proved to reproduce correctly real landslides deposits (e.g.
278 Lucas and Mangeney, 2007; Lucas et al., 2014; Peruzzetto et al., 2019) and dynamics when compared to the force inverted
279 from seismic data (Favreau et al., 2010; Yamada et al., 2018; Moretti et al., 2020). With this rheology, the basal stress T is:

$$280 \quad T = \mu_S \rho h (g \cos(\theta) + \gamma u^2), \quad (1)$$

281 where $\mu_S = \tan(\delta)$ is the friction coefficient and δ the friction angle, ρ is the flow density, h the flow thickness, g the gravity
282 field, θ the local slope angle, γ the topography curvature along flow path and u the velocity norm. Note that in SHALTOP, γ is
283 computed with the topography curvature tensor (see Peruzzetto et al. (2021) for details). In Equation (1), μ_S is used to take into
284 account empirically all dissipative processes occurring within the flow. Other more complex rheologies exist to describe internal
285 friction, e.g. with a soil mechanics approach (Savage and Hutter, 1989), or the $\mu(I)$ -rheology (GDR MiDi, 2004; Jop et al.,
286 2006). Nevertheless, these modeling solutions are either still debated (Gray et al., 2003), or only adapted to flow on simple
287 topographies (e.g., inclined planes in Baker et al., 2016). With the Coulomb rheology, the friction coefficient μ_S needed to
288 model observed deposits decreases as the volume of the avalanche increases (Lucas et al., 2014), at least for dry avalanches.
289 Lucas et al. (2014) suggest the empirical relation between μ_S and the landslide volume V :

$$290 \quad \mu_S = V^{-0.0774} \quad (2)$$

291 Such friction coefficients also proved to reproduce correctly the dynamics of both large (Moretti et al., 2015; Yamada et al.,
292 2018) and small (Levy et al., 2015) landslides. Using this relation with our 1.5×10^6 m³ volume estimation of the Jan. 4, 2018
293 rock avalanche, we get $\mu_S = \tan(18.4^\circ) = 0.33$. However, as shown for instance in Peruzzetto et al. (2019), this estimation may
294 sometimes underestimate the mobility of the rock avalanche, especially when water is present in the avalanche. To model such
295 water-laden avalanches, it is necessary to decrease the friction coefficient μ_S in simulations. Thus, for model calibration, we test
296 friction coefficients between $\mu_S = \tan(10^\circ) = 0.18$ and $\mu_S = \tan(20^\circ) = 0.36$.

297 In order to model the DF, we use frictional rheologies and do not consider visco-plastic rheologies (e.g. Pastor et al., 2004),
 298 as suggested by the granulometry of deposits (see Section 2.4). We test the Coulomb rheology with a friction coefficient
 299 lower than for rock avalanche simulation: their simulated deposits can thus be remobilized. We use $\mu_S = \tan(2^\circ) = 0.03$ and
 300 $\mu_S = \tan(3^\circ) = 0.05$. Such values are low in comparison to other DF simulations carried out with SHALTOP (e.g., $\mu_S = \tan(8^\circ)$
 301 in Moretti et al., 2015). However, with $\mu_S \geq \tan(4^\circ)$, the flow would stop before it reaches the Prêcheur village, which is not
 302 consistent with observations of the Jun. 19, 2010 DF. Besides, such low values are not uncommon in the literature to model
 303 lahars on volcanic slopes (e.g. Pastor et al., 2018a; Frimberger et al., 2021).

304 For snow avalanche and debris flow modeling, the empirical Voellmy rheology is also commonly used (Salm, 1993; Hungry
 305 et al., 2007; Pastor et al., 2018a). It introduces in the basal stress a turbulence term proportional to the square velocity:

$$T = \mu_S \rho h (g \cos(\theta) + \gamma u^2) + \rho g \frac{u^2}{\xi}, \quad (3)$$

307 Following Zimmermann et al. (2020), we choose turbulence coefficients ξ between 100 m s^{-2} and 500 m s^{-2} . Influence of
 308 further increasing ξ is investigated with the Coulomb rheology, as it is equivalent to choosing infinite values for ξ .

309 5 CALIBRATION AND FORWARD PREDICTION SIMULATION RESULTS

310 5.1 Rock avalanche back-analysis

311 The travel distance of the *RA_2018* rock avalanche scenario with various friction coefficients is displayed in Figure 8. The
 312 extent of the Jan. 4, 2018 deposits (dashed green line in Figure 8) is best reproduced with $\mu_S = \tan(14^\circ) = 0.25$. This is
 313 less than $\mu_S = \tan(18.4^\circ) = 0.33$, that is derived from the empirical law of Lucas et al. (2014) (see Section 7.1.1 for further
 314 discussion). With $\mu_S = \tan(14^\circ)$, the flow dissipated energy rate reproduces correctly the main seismic energy increase phase
 315 (Figure 6b, at 30 s). The durations of the seismic signal (60 s) and of the main phase of the simulated energy dissipation (80 s)
 316 are also similar (see Supplementary Note 1 for details on energy computation). However, the flow dissipated energy rate fails to
 317 reproduce the signal complexity, with successive energy peaks (see Section 7.1.2 for a discussion). While most of the energy is
 318 dissipated after 100 s (Figure 6b, red plain line), at that time the flow front is still mobile, about 500 m away from its final
 319 position (see Supplementary Figures 3 and 4). Afterwards, 600 s are still needed for the front to stop. This behavior will be
 320 discussed later on (see Section 7.1.2).

321 For the forward prediction DF simulation, we use the deposits of the rock avalanche simulation as the initial reservoir.
 322 Considering that the extent of deposits observed in 2018 is well reproduced with $\mu_S = \tan(14^\circ)$, we use this parameter to model
 323 a potential future rock avalanche, even though the dynamics of the rock avalanche may not be properly modeled.

324 5.2 Debris flow back-analysis

325 In the *DF_2010_1* scenario, the Voellmy rheology with $\mu_S = \tan(2^\circ)$ and $\xi = 500 \text{ m s}^{-2}$, and the Coulomb rheology with
 326 $\mu_S = \tan(2^\circ)$ and $\mu_S = \tan(3^\circ)$, reproduce relatively well observed flooded areas as well as travel durations.

327 In the village, the thickness of the deposits is mostly below 1 m (Figure 9a-e). On the right bank, the best fit with
 328 observations is obtained with Coulomb and $\mu_S = \tan(2^\circ)$ (Figure 9c). On the left bank, other 2010 overflows are reproduced
 329 by all simulations (Figure 9a-c, green outlines between the bridge and CCPA). However, the flooded area on the left bank is
 330 over-estimated, especially with $\mu_S = \tan(2^\circ)$ (both with the Coulomb and the Voellmy rheologies, Figure 9d).

331 The Jun. 19, 2010 DF travel duration between RPRE and CPMA (1.5 km) is estimated from AFMs recordings between 1
 332 and 4 min. When picking the maximum discharge time at these locations in simulations, only the Coulomb rheology with

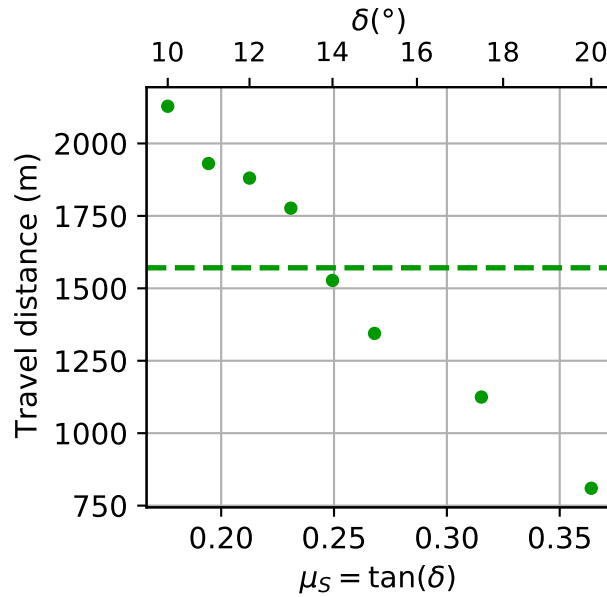


Figure 8. Simulation results for *RA_2018* rock avalanche simulations, with various friction coefficients $\mu_S = \tan(\delta)$. Travel distances are measured from the cliff toe (white cross in Figure 1b and 3b). Error bars (computed by considering 1 to 10 m thickness thresholds when locating the extent of the deposits) are not displayed, but are at most twice the size of the markers. The green dashed line is the observed travel distance of the Jan. 4, 2018 rock avalanche.

333 $\mu_S = \tan(2^\circ)$ could reproduce a 4 min interval (Figure 9f, blue plain line). The second and third smallest interval are 5 min
 334 (Voellmy, $\mu_S = \tan(2^\circ)$ and $\xi = 500 \text{ m}^2$) and 5 min 20 s (Coulomb, $\mu_S = \tan(3^\circ)$). For these 3 simulations, the corresponding
 335 flow durations between RPRE and the Prêcheur bridge (4,3 km) vary between 10 and 24 min (Figure 9g).

336 In comparison, the *DF_2010_2* scenario, that involves a larger volume ($1.2 \times 10^6 \text{ m}^3$), yields travel durations that are more
 337 compatible with observations, both with Coulomb and $\mu_S = \tan(2^\circ)$ or $\mu_S = \tan(3^\circ)$, and with Voellmy and $\mu_S = \tan(2^\circ)$ and
 338 $\xi = 500 \text{ m s}^{-2}$ (Figure 10f). With these parameters, the flow travel time between RPRE and the bridge is less than 20 min
 339 (Figure 10g). However, flooded areas are largely over-estimated, both on the right and left banks (Figure 10a-d). In particular
 340 with Coulomb and $\mu_S = \tan(2^\circ)$, the DF runs over the river right bank about 400 m downstream CCPA, and enters two adjacent
 341 gullies (Figure 10c, black dashed lines on the northern side of the river). This suggests scenario *DF_2010_1* is more realistic
 342 than scenario *DF_2010_2* to reproduce the Jun. 19, 2010 DF.

343 Considering the uncertainty on the calibration parameters for DF simulation, we use Coulomb (with $\mu_S = \tan(2^\circ)$ or
 344 $\mu_S = \tan(3^\circ)$) and Voellmy (with $\mu_S = \tan(2^\circ)$ and $\xi = 500 \text{ m s}^{-2}$) for DF modeling in the forward prediction simulation.

345 5.3 Forward-prediction simulation results

346 In the *RA_fwd* scenario, we model a potential future $1.9 \times 10^6 \text{ m}^3$ rock avalanche from the Samperre cliff with Coulomb and the
 347 calibrated friction coefficient $\mu_S = \tan(14^\circ) = 0.25$. The final deposits are similar to the *RA_2018* simulation ($1.5 \times 10^6 \text{ m}^3$)
 348 with the same friction coefficient, as they extend only a few tens of meters further downstream (Figure 11a). Their maximum
 349 thickness is about 30 m.

350 This reservoir is then used as a source term for the propagation of the DF. Following the calibration results, we test three
 351 rheologies: the Voellmy rheology with $\mu_S = \tan(2^\circ)$ and $\xi = 500 \text{ m s}^{-2}$, and the Coulomb rheology with $\mu_S = \tan(2^\circ)$ or

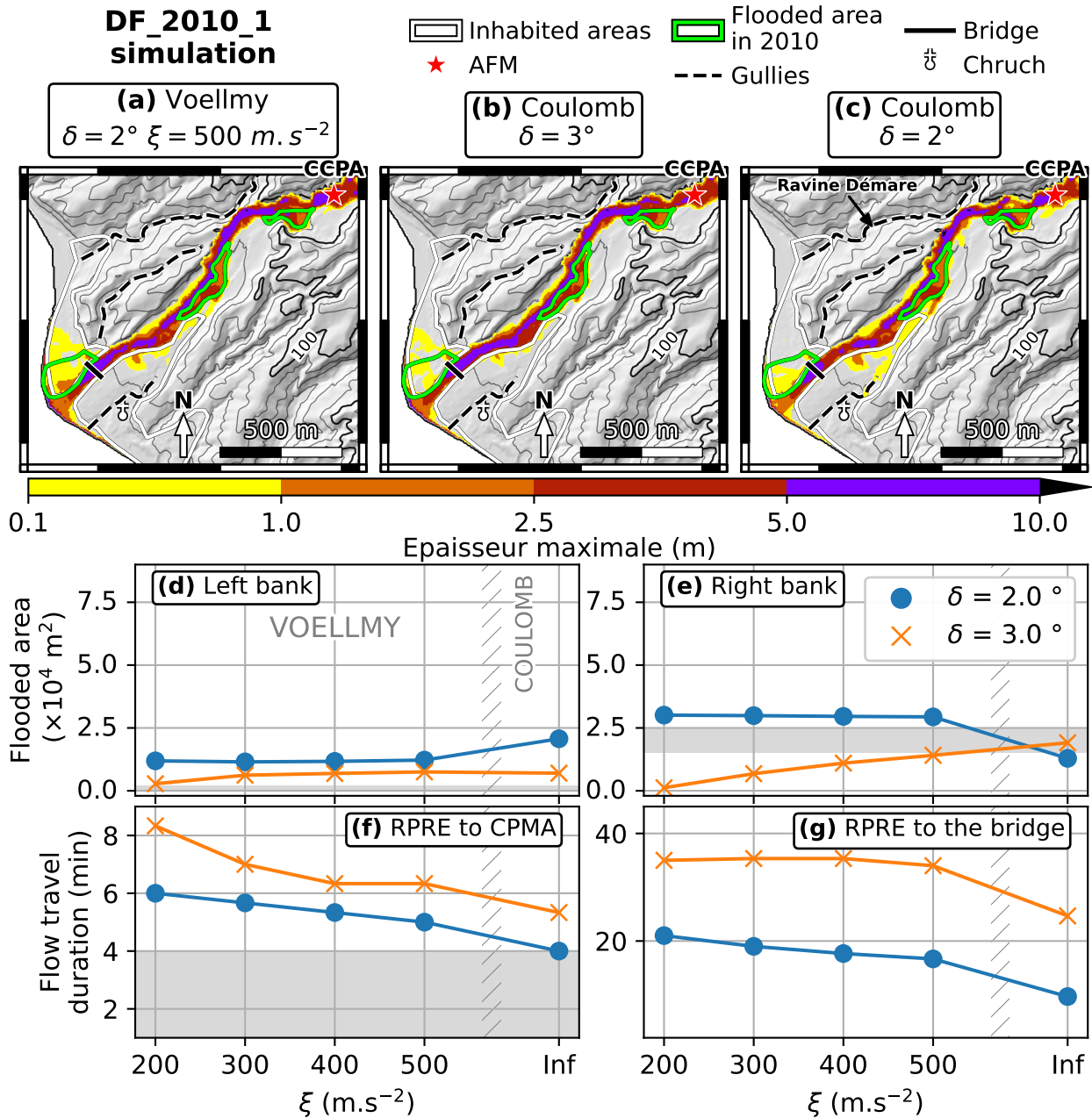


Figure 9. Simulation results for the *DF_2010_1* scenario ($0.65 \times 10^6 \text{ m}^3$). (a) Maximum flow thickness with the Voellmy rheology, $\mu_S = \tan(2^\circ) = 0.03$ and $\xi = 500 \text{ m} \cdot \text{s}^{-2}$, (b) with the Coulomb rheology and $\mu_S = \tan(3^\circ) = 0.05$, and (c) with the Coulomb rheology and $\mu_S = \tan(2^\circ) = 0.03$. Topography is the 08/2018 DEM. Each point in (d), (e), (f) and (g) is a simulation result, with friction coefficient given by line color and turbulence coefficients given by the x-coordinate. Left of hatches is for the Voellmy rheology, right is for the Coulomb rheology (equivalent to infinite turbulence coefficient). (d) and (e): Area flooded on the left (d) and right (e) riverbank, within inhabited areas. (f) and (g): Flow travel duration between RPRE and CPMA ((f), about 1.6 km), and between RPRE and the Prêcheur bridge ((g), about 4.3 km), measured by picking the maximum of the discharge at each location. Grey patches are observations for δ for the Jun. 19, 2010 DF, taking into account uncertainties.

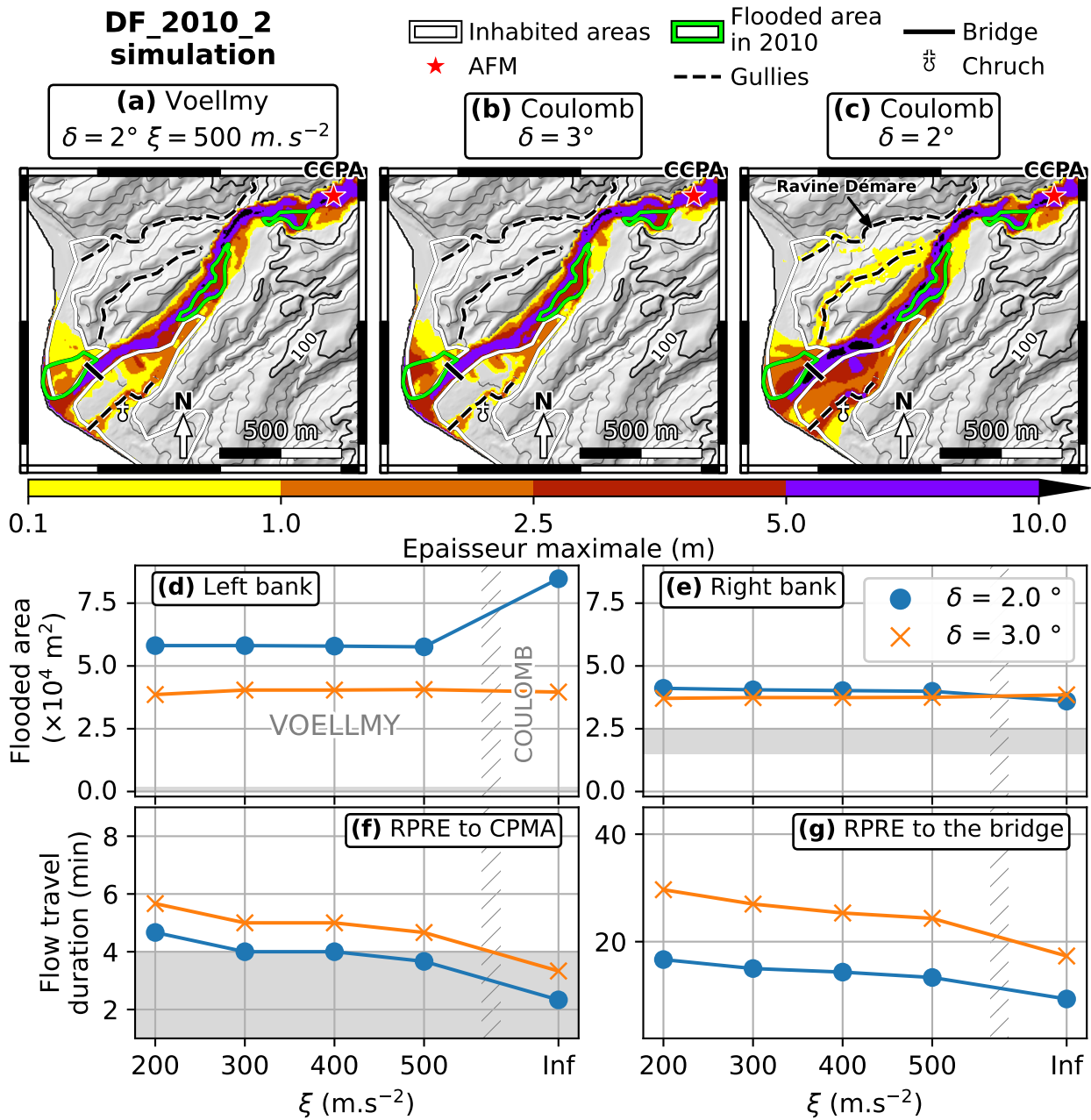


Figure 10. Simulation results for the *DF_2010_2* scenario ($1.2 \times 10^6 \text{ m}^3$). See Figure 9 for legend.

352 $\mu_S = \tan(3^\circ)$. With the Voellmy rheology, travel durations and flooded areas are very similar to results derived in the *DF_2010_2*
353 scenario (Supplementary Figure 5). However, the DF velocity is reduced by about 10% when the Coulomb rheology is used.
354 As a matter of fact, in comparison to the *DF_2010_2* scenario, the initial mass is spread more broadly in the river bed, such
355 that the flow front accelerates on a shorter distance. This effect is not observed with Voellmy because the turbulent term in
356 Equation (3) prevents the flow from accelerating indefinitely. Peak discharges at RPRE vary between 4,000 and 6,000 $\text{m}^3 \text{s}^{-1}$
357 (Figure 11b-d, blue lines): this is coherent with field observations in other contexts, for this range of volumes (see Figure 2 in
358 Rickenmann, 1999). With Coulomb and $\mu_S = \tan(3^\circ)$, some of the flowing material stops before it reaches the sea, such that
359 the peak discharge at the bridge does not exceed 400 $\text{m}^3 \text{s}^{-1}$. To the contrary, $\mu_S = \tan(2^\circ)$ increases mobility, and peak
360 discharges reach almost 1000 $\text{m}^3 \text{s}^{-1}$ with the Voellmy rheology (Figure 11c), and more than 1600 $\text{m}^3 \text{s}^{-1}$ with the Coulomb
361 rheology (before the DF overflows the river bed, Figure 11d).

362 These results provide a first insight on the most exposed areas in the case of a future massive rock avalanche followed by a
363 high discharge DF remobilizing all deposits, provided the rock avalanche and DF have similar behaviours and solid content
364 as the events used for calibration. In the following, we investigate the influence of initiation processes and location on the
365 simulation results.

366 6 SENSITIVITY ANALYSIS: INFLUENCE OF INITIATION MECHANISM

367 6.1 Influence of successive destabilizations on rock avalanches simulations

368 To investigate the influence of retrogressive destabilizations on runout prediction, we release the $1.5 \times 10^6 \text{ m}^3$ of the *RA_2018* in
369 two successive steps, instead of one. In the resulting *RA_2018_2* scenario, $0.8 \times 10^6 \text{ m}^3$ are first released at the cliff bottom (A
370 in Supplementary Figure 6a), and the rest (B in Supplementary Figure 6a) collapses 13 s later. The two volumes are constructed
371 arbitrarily by separating the extent of the initial mass of the *RA_2018* scenario approximately at the middle of the cliff. Thus,
372 the resulting two volumes are similar. The 13 s delay between the two collapses matches the initial duration of the seismic signal
373 before the seismic energy starts increasing sharply (see Figure 6b). Because SHALTOP models one-phase/one-layer flows, it
374 should be noted that in the *RA_2018_2* scenario, the second avalanche is assumed to be mixed with the first one as soon as they
375 join. As a result, we do not model the possible development of a two-layer flow, with the second avalanche propagating above
376 the first one. This could enhance mobility by flattening the topography and favoring erosion (Mangeney et al., 2010; Farin et al.,
377 2014).

378 Successive collapses do help reproduce, to some extent at least, the complexity observed in the Jan. 4, 2018 seismic signal
379 (compare red dashed line and black line in Figure 6b). However, the geometry of final deposits (and thus the geometry of the
380 debris reservoir that will be remobilized later on as a DF) remains the same (compare Supplementary Figure 6b and 6d).

381 6.2 Influence of progressive release on debris flows simulations

382 In our DF simulations, the initial reservoir is remobilized instantaneously. Although we manage to reproduce rather correctly
383 the flooded area and travel times of the Jun. 19, 2010 DF, it is in general difficult to characterize the initiation process of DFs.
384 Besides, for a given debris reservoir, the initiation mechanism may not be independent from the remobilized volume. Such
385 correlations are beyond the scope of this study. In this section, we only explore the influence of the initiation process on debris
386 flow dynamics, for a given debris flow volume. In order to investigate empirically the effect of progressive remobilization, we
387 release $0.65 \times 10^6 \text{ m}^3$ (i.e., the same volume as in the *DF_2010_1* scenario) over a 200 m^2 area at the cliff bottom (white cross

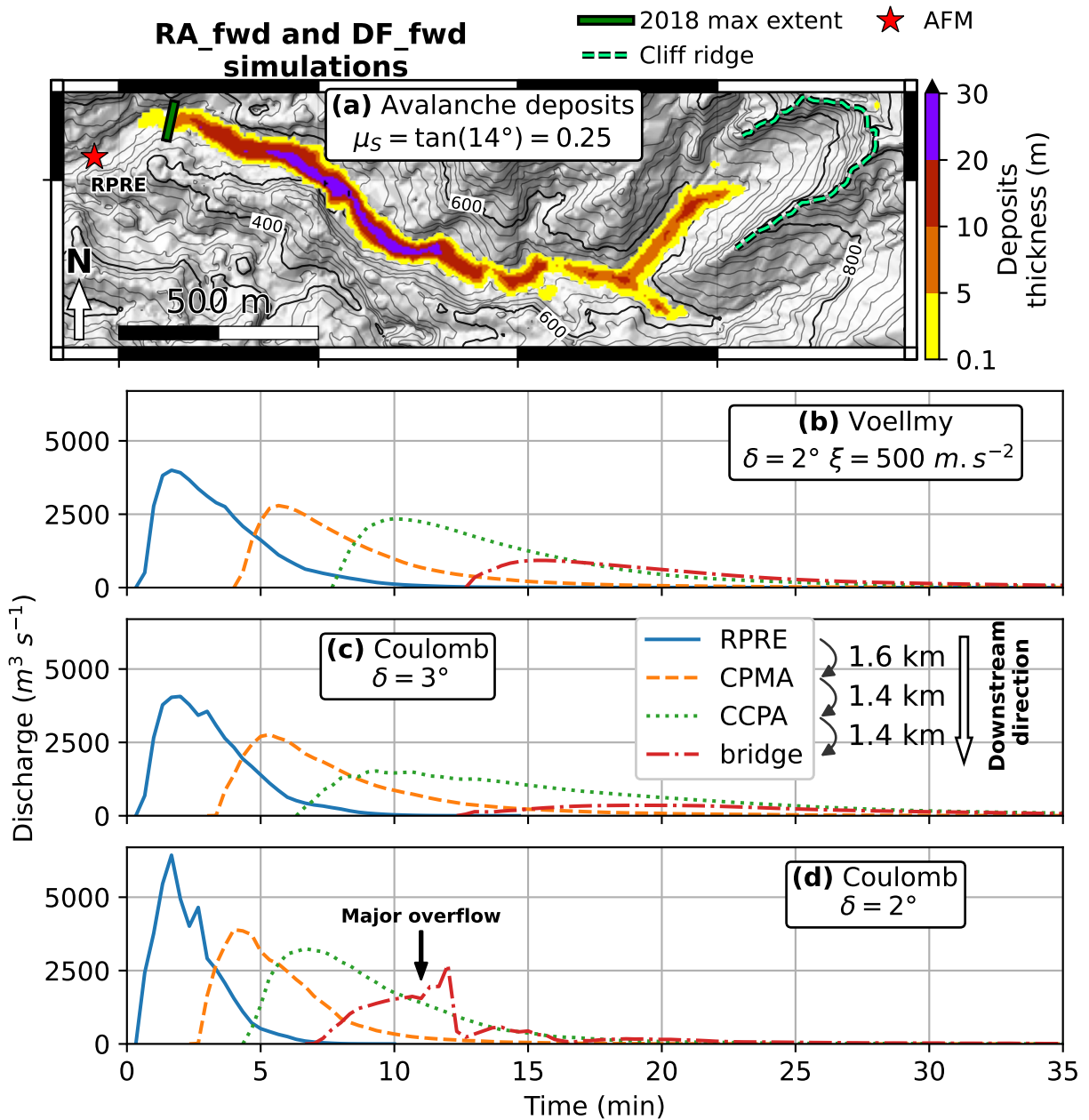


Figure 11. Results of the *RA_fwd* rock avalanche scenario and subsequent *DF_fwd* DF simulation. (a) Final deposits of the rock avalanche, modeled with Coulomb and $\mu_S = \tan(14^\circ) = 0.25$. The green line is the observed runout of the Jan. 4, 2018 rock avalanche. The topography is the 08/2018 DEM. (b), (c) and (d): Simulated discharges at RPRE, CPMA, CCPA and the bridge. (b) Voellmy rheology, $\mu_S = \tan(2^\circ)$ and $\xi = 500 \text{ m} \cdot \text{s}^{-2}$. (c) Coulomb rheology, $\mu_S = \tan(3^\circ)$. (d) Coulomb rheology, $\mu_S = \tan(2^\circ)$. Strong variations in (d) for discharge at the bridge result from major overflows around the bridge (see Supplementary Figure 5c).

388 in Figure 3), through a constant discharge lasting $\Delta t = 10$ or 20 min. Thus, simulations differ solely by the release duration,
 389 allowing for the comparison of results. The initial discharge is thus inversely proportional to Δt . Results are given in Figure 12.
 390 Increasing release duration slows down the DF and reduces flooded area. Using $\Delta t = 10$ min and the Coulomb rheology with
 391 $\mu = \tan(2^\circ)$ enhances the match between observed and simulated flooded areas, but over-estimates slightly the travel duration
 392 between RPRE and CPMA (compare blue circles and shaded area in Figure 12d-f). Note that in Figure 12 travel durations are
 393 measured by picking the onset of discharge increase, because no clear maximum can be identified in RPRE when we impose a
 394 constant discharge in the source area (see Figure 13b).

395 If we assume AFM records are qualitative proxies for flow discharge at nearby locations, the temporal evolution of RPRE's
 396 record (with a sharp increase and a progressive decrease) of the Jun. 19, 2010 DF is better reproduced with an instantaneous
 397 release (compare Figures 13a and 13c). However, the 15 min duration of the flow at RPRE is better reproduced with a
 398 progressive release (compare Figures 13b and 13c). This may indicate that most of the debris involved in the Jun. 19, 2010
 399 DF was released instantaneously, but that part of the initial reservoir was remobilized afterwards. Thus, more realistic initial
 400 set-up would involve a non constant discharge, but such initial conditions are not implemented in SHALTOP. Nevertheless, an
 401 instantaneous release proved to be sufficient to reproduce the main characteristics of the Jun. 19, 2010 DF (travel duration and
 402 flooded area), at least in a first approximation.

403 6.3 Influence of source area on debris flow simulations

404 For a given released volume, the location of the release area has in comparison little influence on the results of DF simulations.
 405 When the release is instantaneous, we saw that the *DF_2010_2* and *DF_fwd* scenarios, that involve similar volumes but different
 406 initial geometries, yield similar results (see Figure 10 and Supplementary Figure 5). The same conclusion is drawn when using
 407 a constant discharge, located either at the cliff bottom, at the waterfall or at RPRE (see Figure 1 for locations): travel durations
 408 and flooded areas are very similar (see Supplementary Figure 7).

409 7 DISCUSSION

410 7.1 Rock avalanche modeling

411 7.1.1 Choice of rheological parameters

412 In this study, the friction coefficient μ_S used in the rock avalanche forward prediction simulation is chosen after a calibration
 413 step, as often done in the literature (e.g. Sosio et al., 2012; Pastor et al., 2018a). To our knowledge, it is difficult to estimate
 414 μ_S directly from physical characteristics of the materials. Indeed, simulations of laboratory experiments involve high friction
 415 coefficient (for instance, $\mu_S = \tan(30^\circ)$ in Gray et al., 1999) that fail to reproduce deposits and dynamics observed at the field
 416 scale.

417 If no calibration data are available, another solution is to use empirical laws derived from field observations. Lucas et al.
 418 (2014) estimate the mobility of landslides through the effective friction coefficient μ_{eff} . μ_{eff} differs from the traditional angle
 419 of reach (or Heim's ration) μ_H : while μ_H only depends on the landslide runout, μ_{eff} also takes the initial mass geometry into
 420 account. We have:

$$421 \mu_{eff} = \tan(\theta) + \frac{H_0}{\Delta L}, \quad (4)$$

$$422 \mu_H = \frac{H}{\Delta L'}, \quad (5)$$

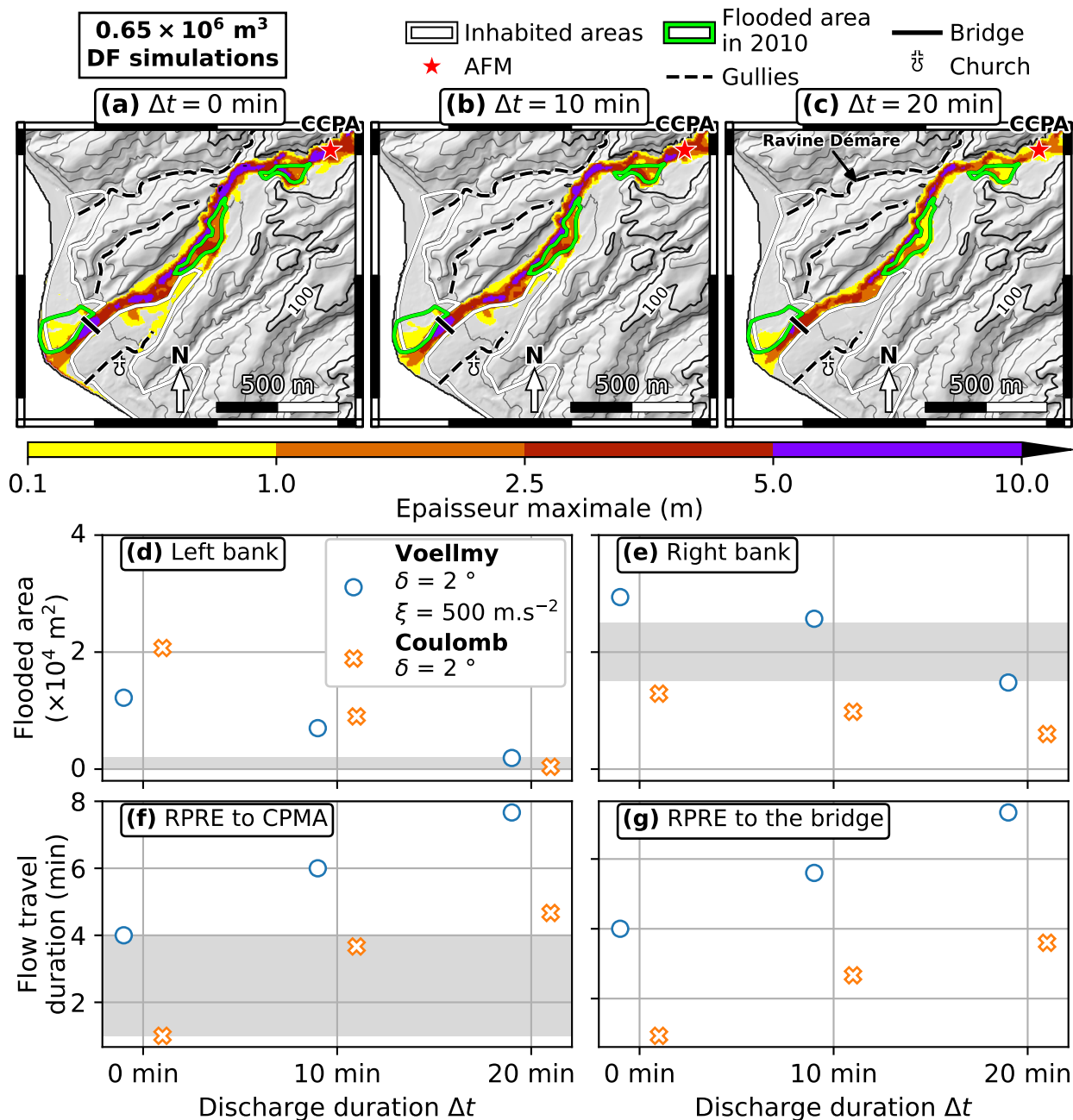


Figure 12. *DF_2010_I* simulation with instantaneous or progressive release (10 or 20 min, see abscissa). The released volume is always $0.65 \times 10^6 \text{ m}^3$. (a), (b), (c) Maximum flow thickness, for different durations Δt of initial discharge. (d) and (e): Area flooded on the left (d) and right (e) riverbank, within inhabited areas. (f) and (g): Flow travel duration between RPRE and CPMA ((f), about 1.6 km), and between RPRE and the Prêcheur bridge ((g), about 4.3 km), measured by peaking the onset of discharge increase. Grey patches are observations for the Jun. 19, 2010 DF, taking into account uncertainties.

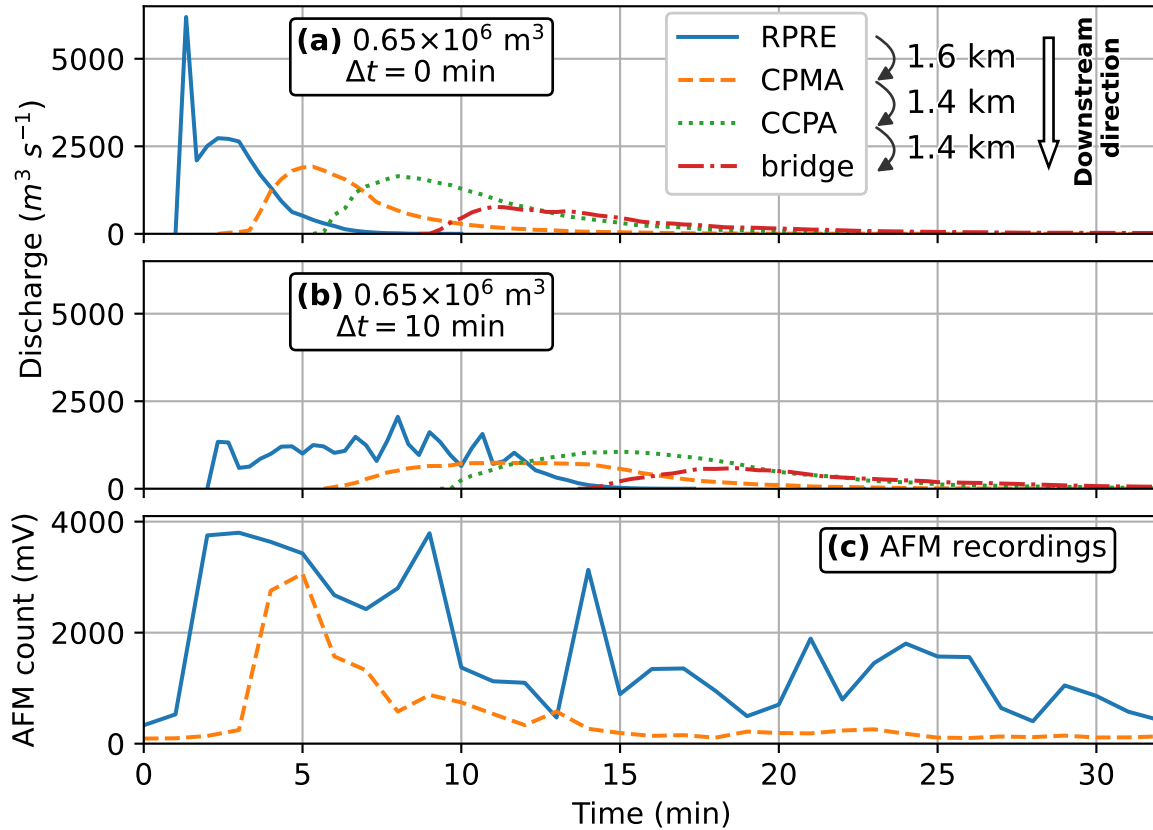


Figure 13. Comparison between simulated discharges at fixed locations and AFMs recordings. (a) Simulated discharges at RPRE, CPMA, CCPA and at the bridge (see Figure 1 in the main body of the article for locations) for the *DF_2010* scenario, with Coulomb and $\mu_S = \tan(2^\circ)$, and an instantaneous release. (b) Same as (a), but with a constant source discharge during 10 min. (c) AFM recordings at RPRE and CPMA for the Jun. 19, 2010 DF. $t = 0$ min is 8:39 UTC, Jun. 19, 2010.

424 with θ the topography average slope along flow path, H_0 the maximum thickness of the initial mass and ΔL the landslide
 425 travel distance along topography from the scar toe. H (drop height) and $\Delta L'$ are respectively the difference in altitude and
 426 horizontal distance between the upper scar and furthest deposits location (see supplementary materials in [Lucas et al., 2014](#)).
 427 The expression (4) of μ_{eff} is derived from the analytical solution of thin-layer dam-break ([Mangeney et al., 2000](#); [Faccanoni](#)
 428 [and Mangeney, 2012](#)). [Lucas et al. \(2014\)](#) use a database of terrestrial and non-terrestrial landslides with a small amount of
 429 water to estimate empirical relations relating μ_H and μ_{eff} to the landslide volume V :

$$430 \mu_{eff} = V^{-0.0774}, \quad (6)$$

$$431 \mu_H = 1.2V^{-0.089}. \quad (7)$$

433 When we apply these relations to the 2018 Samperre rock avalanches, we get values between $\tan(18.5^\circ) = 0.33$ and $\tan(19.5^\circ) =$
 434 0.35 for both μ_H and μ_{eff} . This is in good agreement with values computed directly from observations, using Equations (4)
 435 and (5) (between $\tan(19^\circ) = 0.34$ and $\tan(19.5^\circ) = 0.35$ for both μ_H and μ_{eff}).

436 In comparison, we used $\mu_S = \tan(14^\circ) = 0.25$ to reproduce observed travel distances. It has been shown that μ_H cannot be
 437 used to estimate directly the flow mobility: although it is related to the effective mobility of the landslide, it also includes purely
 438 geometrical descriptors such as topographic slope or initial mass geometry (e.g. [Lucas and Mangeney, 2007](#); [Lucas et al., 2014](#)).
 439 The latter are corrected add by the more complex definition of μ_{eff} , such that it proved to better estimate the friction coefficient
 440 μ_S needed to reproduce real landslides ([Lucas et al., 2014](#)). As shown in Figure 14, the empirical relation (6) is globally in
 441 agreement with values of μ_S calibrated with SHALTOP on other sites. Nevertheless, significant dispersion is observed both
 442 for the empirical relation (see the 95% confidence interval for Equation (6), shaded area in Figure 14) and calibrated values
 443 (e.g., for volumes above 10^8 m^3 , Figure 14). This dispersion may be partly explained by the fact that μ_S does not depend only
 444 on volume. The mobility also depends, for instance, on water content (e.g. [Peruzzetto et al., 2019](#)), path material ([Aaron and](#)
 445 [McDougall, 2019](#)) and erosion processes ([Mangeney et al., 2010](#)). Besides, the expression of μ_{eff} was derived for flows on
 446 constant and laterally uniform slopes. The generalization of Equation (4) to general topographies with, for instance, varying
 447 slopes and bended channels is not straightforward. This may also explain the uncertainty of the empirical relation (6), and the
 448 difference with calibrated values of μ_S .

449 Interestingly, Equation (6) seems to over-estimate μ_S when calibration is done by reproducing deposits (blue circles and
 450 white square in Figure 14), and slightly under-estimate μ_S when calibration uses seismic signal (pink crosses and orange
 451 diamond in Figure 14). This is consistent with results of [Moretti et al. \(2020\)](#): when they use only the force applied on the
 452 ground (inverted from seismic recordings) to calibrate μ_S , the observed travel distance is under-estimated. However, [Lucas](#)
 453 [et al. \(2014\)](#) do not highlight any systematic bias between μ_{eff} and values of μ_S calibrated from deposits (see their Figure 3b).
 454 To investigate more thoroughly these discrepancies, a larger database of back-analyzed landslides would be needed. This is
 455 beyond the scope of this study, but highlights the uncertainty associated to the calibration of simulation parameters.

456 **7.1.2 Influence of initiation mechanism on deposits geometry**

457 The fact that for a given volume, the initiation mechanism has little influence on the travel distance is consistent with results
 458 from [Moretti et al. \(2015\)](#) who model the 2010 Mount Meager landslide, with 1, 2 or 3 successive collapses. It can be
 459 explained by the fact that the initial potential energy is dissipated quickly in the first 30 s to 50 s (see Figure 6c). Indeed,
 460 whatever the initiation mechanism, the rock avalanche is blocked at the inlet of the Samperre river (just upstream the waterfall,
 461 see Figure 1), that is too narrow for the avalanche to enter it at once. Then, the rock avalanche can move further downstream

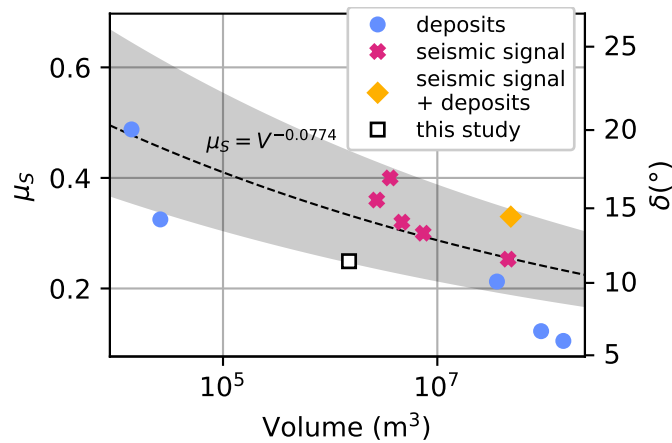


Figure 14. Values of $\mu_S = \tan(\delta)$ calibrated with SHALTOP to reproduce terrestrial landslides with a single constant friction coefficient, using deposits and/or seismic data (Lucas et al., 2007; Kuo et al., 2009; Moretti et al., 2015, 2020; Yamada et al., 2018; Peruzzetto et al., 2019). See Supplementary Table 1 for details. The square is the calibration result of the RA_2018 simulation. The black dashed line gives the empirical relation $\mu_S = V^{-0.0774}$, with the 95% confidence interval (see Supplementary Table 4 in Lucas et al., 2014).

462 only if relatively small friction coefficients (close to or smaller than the topographic slope) are used in the simulations, whatever
 463 the initial dynamics.

464 Nevertheless, the fact that the initial mechanism has little influence on the travel distance may be true for large collapses
 465 only. On May 11, 2010, destabilizations occurred as a succession of 47 successive events (Clouard et al., 2013). Given the
 466 estimated $2.1 \times 10^6 \text{ m}^3$ total volume that collapsed during the whole crisis, this suggests an average volume of less than
 467 $50,000 \text{ m}^3$ per event. Following Lucas et al. (2014), friction coefficients around $\mu_S = \tan(23^\circ) = 0.42$ are needed to model the
 468 propagation of such volumes. In turn, these small granular avalanches stop in the vicinity of the cliff toe, as observed in the
 469 field, and do not enter the river bed. In comparison, larger granular flows are modeled with lower friction coefficients (e.g.
 470 $\mu_S = \tan(14^\circ) = 0.25$ in our simulations) and have longer runouts (for a review of possible mechanisms enhancing the mobility
 471 of large landslides, see e.g. Korup et al., 2013).

472 To investigate into more details the initiation mechanisms, Discrete Element Methods simulations can be carried out to
 473 model explicitly the interactions between blocks (e.g. Chen and Wu, 2018; Do and Wu, 2020; Feng et al., 2021). However, we
 474 believe that such models are not necessarily better suited than thin-layer models to simulate the propagation. As a matter of
 475 fact, the Samperre cliff is composed of indured pyroclastic deposits that disintegrate rather quickly after the destabilization into
 476 sand and boulders. Given the volumes considered (about $1 \times 10^6 \text{ m}^3$ for the large rock avalanches), modeling explicitly each
 477 particle in DEM simulations would demand too much computational resources. The explicit modeling of fracture propagation
 478 and disintegration is also possible but meets the same computational limitations (Stead and Coggan, 2006).

479 7.2 Debris flow modeling

480 7.2.1 Rheology and rheological parameters

481 In this work we have tested only the Coulmb and Voellmy rheologies. Another possible rheology that could have been
 482 investigated (but that is not implemented in SHALTOP) is the combined Darcy-Weisbach and Manning rheology (Chow, 1959;

483 O'Brien et al., 1993; Jakob et al., 2013):

$$484 T = \rho n^2 \frac{u^2}{h^{1/3}}, \quad (8)$$

485 where n is the Manning coefficient. Note that Equation (8) resembles Equation (3) giving the basal stress in the Voellmy
486 rheology. Assuming a Manning coefficient $n = 0.05$ (Jakob et al., 2013), a flow height $h = 5$ m (average flow depth in our
487 simulation), a turbulence coefficient $\xi = 500 \text{ m s}^{-2}$ (as in our simulations), and $g = 9.81 \text{ m s}^{-2}$, we have:

$$488 T = 1.5 \times 10^{-3} \rho u^2 \text{ for the Darcy-Manning rheology,} \quad (9)$$

$$489 T = 2.0 \times 10^{-2} \rho u^2 \text{ for the Voellmy rheology.} \quad (10)$$

490 We may thus expect faster flows with the Darcy-Manning rheology. However, note that Equation (8) is derived empirically for
491 permanent flows in open channels only (Chow, 1959). Thus, we do not believe that the Darcy-Manning rheology is more fitted
492 to debris flow simulations than the Voellmy or Coulomb rheologies.

493 The Voellmy rheology is commonly used to model fast gravity-driven flows such as snow avalanches and debris flows
494 because Coulomb sometimes fails to reproduce observed velocities (Peruzzetto et al., 2018a). It can indeed yield velocities
495 unrealistically high, as the flow accelerates as long as the topographic slope exceeds the friction coefficient (Kelfoun, 2011).
496 Note that this problem could be the result of the shallow approximation (i. e. hydrostatic pressure) that lead to strong
497 overestimation of the velocity (Figure 9b of (Mangeney et al., 2010), Figures 19 and 20 of (Garres-Díaz et al., 2021)). Although,
498 in some cases, the Voellmy rheology allows to better fit observed velocities (e.g. Peruzzetto et al., 2018a), its two parameters
499 can be difficult to constrain. Indeed, several couples (μ_s, ξ) may give similar results.

500 In the case of the Prêcheur river, we showed that the Coulomb rheology could reproduce both the travel duration and flooded
501 area of the Jun. 19, 2010 DF. Thus, with the data available to characterize this event in particular, there is no clear advantage of
502 using the Voellmy rheology, and thus of introducing a second rheological parameter.

503 When no calibration data are available, the choice of rheological parameters is more complex. In the case of visco-plastic
504 DFs, rheometry, slump tests and flume tests can be done at the laboratory scale to estimate, in particular, the flow viscosity
505 and yield stress (e.g. Coussot et al., 1998; Remaître et al., 2005; Bouteiller et al., 2021). However, the resulting rheological
506 parameters do not always allow to reproduce observations in thin-layer simulations, because the samples are generally sieved to
507 include only the fine fraction for experimental constraints, and may thus not be representative of the actual DF (Sosio et al.,
508 2007). Besides, viscosity and yield stress depend on solid concentration (Iverson, 2003).

509 Anyway, in our case, the granulometry of the deposits suggests that the DF dynamics have a frictional mechanical behaviour.
510 To our knowledge, no laboratory experiment allows to estimate the friction coefficient μ_s used to model debris flows in these
511 conditions, with the Coulomb rheology. A basic approach, though, is to consider the slope where the debris flow is expected to
512 stop, and use the corresponding friction coefficient. This rationale helped us define a range of possible values for μ_s before
513 calibration, and was also used for instance by Franco-Ramos et al. (2020) with the Voellmy rheology. It demands, of course, an
514 a priori on the debris flow expected runout, that can be justified by field observations or expert judgment. Thus, in this case, the
515 operational relevance of simulations is not to indicate whether the debris flow will reach a particular location. It is rather to
516 estimate key characteristics of the flow such as travel time or flooded areas, provided the debris flow reaches a given location.
517 Such information are important for hazard assessment.

518 With the Voellmy rheology, the turbulence coefficient is, by definition, empirical (Salm et al., 1990), and thus must be
519 calibrated. A range of possible values may however be given by the literature (e.g. Zimmermann et al., 2020).

520 **7.2.2 Erosion processes**

521 As discussed previously, we have not considered entrainment in our simulations. Apart from the influence such a process could
522 have on the DF initiation, we may expect that erosion influences the DF dynamics further downstream as shown in laboratory
523 experiments of granular flows (Mangeney et al., 2010; Farin et al., 2014; Iverson et al., 2011; Mangeney, 2011). In particular,
524 the upper river section above RPRE is narrow and steep-walled, with slopes between 7° and 12° , such that it is prone to bed
525 (from previous lahar deposits) and lateral erosion. The increase of DF volume is difficult to estimate in our case. However,
526 drastic volume increase is sometimes observed in other contexts (e.g., from 150 to 1620 m³ for the 2000 Tsing Shan debris flow
527 in Hong Kong, Pirulli and Pastor, 2012).

528 Nevertheless, such processes are difficult to model and constrain. Erosion rate is classically assumed to be proportional to
529 the flow momentum (McDougall and Hungr, 2005; Pirulli and Pastor, 2012), but other studies suggest it is actually inversely
530 proportional to the flow velocity (Iverson, 2012; Lusso et al., 2017, 2020). Bouchut et al. (2008), and later on Iverson (2014),
531 highlight the methodological complexity of deriving a physically based model for erosion, in particular to ensure energy is
532 preserved in the momentum equations (see also Iverson and Ouyang, 2015; Pudasaini and Fischer, 2020). Both with empirical
533 and physically-based erosion laws, simulation results strongly depend on an a priori expert knowledge of erosion areas and
534 erodible thicknesses.

535 As shown in Section 6.3, the initial mass geometry or source location of DF simulations have a limited influence on
536 simulation results, at least when the DF is initiated in the upper section of the river, above RPRE. Thus, the DF volume increase
537 due to erosion in this section can be accommodated for empirically, in a first approximation, by directly changing the DF initial
538 volume. In the second section of the river that is wider and flatter, we may expect that deposition will prevail over erosion. It
539 may nevertheless not stand true when DFs occur one after another, entraining loose and unconsolidated deposits of previous
540 DFs. To investigate such situations and model DF bulking, it may be necessary to take into account erosion, even empirically.

541 **7.2.3 Overflow hazard**

542 DF simulations provide a first insight on the areas most exposed to overflow hazard. The possibility that DFs overflow the
543 river banks between the bridge and RPRE, or enter adjacent gullies, is a major concern. In an expert report, Quefféleán (2018)
544 suggests that the rocky edge separating the Prêcheur river from the Ravine Démare, a few hundred meters downstream CCPA,
545 could be overflowed (or even destroyed) by high discharge DFs. Although the over-topping of river banks is a highly non-linear
546 phenomenon, with thresholds effects (Mergili et al., 2018; Peruzzetto et al., 2019) that are not easy to predict precisely, such an
547 overflow is reproduced in our *DF_fwd* simulation with Coulomb and $\mu_S = \tan(2^\circ)$. This simulation also suggests that part of
548 the flow may enter the gully between the Prêcheur river and the Ravine Démare (Figure 10c). This possibility had not been
549 considered by Quefféleán (2018), and should be further investigated in future field works.

550 Analyzing flood hazard in the village is also of prior importance, but its quantification is not easy either. Indeed, flow
551 mobility has competing effects. On the one hand, more material will reach the Prêcheur village when smaller friction coefficients
552 and/or higher turbulence coefficients are used, increasing overflow hazard. On the other hand, low friction coefficients favor the
553 evacuation of debris into the ocean. For instance, in the *DF_2010_1* scenario, when we increase the turbulence coefficient (up
554 to infinite values for the Coulomb rheology), the flooded area on the river right bank expands for $\mu_S = \tan(3^\circ)$ but lessens for
555 $\mu_S = \tan(2^\circ)$ (Figure 9e).

556 At the mouth of the river, overflows are all the more hard to model as they strongly depend on the river bed filling level, that

557 can vary during a DF because of progressive sediment settling. Such a process is not modeled in SHALTOP where we consider
558 a one-phase flow, with the flowing column stopping at once. Multi-phase shallow water models, such as *D-Claw* (George and
559 Iverson, 2014; Iverson and George, 2014), *ravaflow* (Mergili et al., 2017; Pudasaini and Mergili, 2019) or *GeoFlow_SPH*
560 (Pastor et al., 2018b), could help investigate such effects. But, as discussed previously, they are more complex to calibrate and
561 the design of appropriate erosion/deposition laws is still an open issue.

562 Another key physical process that we do not model, but that may be important to assess correctly overflow hazard at the
563 mouth of the river, is the dilution of the DF as it reaches the sea. As we do not have bathymetric data, the altitude in the sea is
564 set to 0 and we let the material flow freely through the grid boundary. Provided bathymetric data is available, the interaction
565 between sea water and the DF can, in theory, be empirically modeled with two-phase or multi-phase models (Pudasaini and
566 Mergili, 2019). We may however expect some process, such as the transformation of the DF into a turbidity current
567 (Elverhøi et al., 2000), not to be properly simulated. To our knowledge, research has mainly focused on understanding the
568 generation of tsunamis by debris flows (e.g. Walder and Watts, 2003; de Lange et al., 2020), rather than on the influence of
569 debris flow dilution in a large water body on the upstream dynamics. As the Prêcheur village is built around the river mouth, it
570 may be worth investigating this aspect.

571 **7.2.4 Comparison between DF simulations and other documented events**

572 We focused on the modeling of high discharge DFs because their velocity favors the mixing of solid and fluid phases and
573 prevents sediment settling. In turn, the assumption of a homogeneous flow is more acceptable for high discharge DFs than
574 for smaller events, and in particular HFs, where the solid and fluid phases are separated. However, we may wonder if our
575 simulations allow to reproduce empirically and in a first approximation the distribution of flow travel durations between RPRE
576 and CPMA of other documented events. In Figure 15, we compare travel times measured on the 8 strongest lahars (without
577 distinguishing between DFs and HFs) between September 2009 and August 2010 (classified as "strong" or "very strong" by
578 Aubaud et al., 2013, without distinguishing between DFs and HFs), to travel times modeled for the *DF_2010_1* and *DF_2010_2*
579 scenarios with various rheological parameters (μ_S between $\tan(2^\circ)$ and $\tan(4^\circ)$, and ξ between 100 m s^{-2} and 500 m s^{-2}).

580 Observed average travel durations decrease for increasing peak FULL values at RPRE (Figure 15a). When the latter are
581 higher than 3000 mV, lahars need no more than 7 min to go from RPRE to CPMA. However, when RPRE FULL records are
582 about 1000 mV, travel durations span from 2 to 15 min. Any further interpretation is difficult because of picking uncertainty:
583 sampling interval is only 1 min and the identification of maximum couples in RPRE and CPMA is sometimes difficult.

584 However, we could reproduce the same range of travel durations by using different initial conditions and rheological
585 parameters (Figure 15b). High discharges at RPRE (more than $5000 \text{ m}^3 \text{ s}^{-1}$) are associated to travel durations between RPRE
586 and CPMA below 5 min, while a discharge of $2500 \text{ m}^3 \text{ s}^{-1}$ yields durations spanning from 5 min to 12 min. With the Voellmy
587 rheology, changing rheological parameters only slightly changes the modeled discharge but entails important variations in travel
588 durations (e.g., triangles in Figure 15b). To the contrary, with the Coulomb rheology, a same simulation scenario will produce
589 different discharges depending on the friction coefficient (e.g., triangles with dashed black circles in Figure 15b).

590 This preliminary analysis is encouraging: even with a simple one-phase thin-layer model and no more than two parameters,
591 we model realistic travel times. However, a more thorough comparison with other recorded DFs and HFs is needed to assess
592 more precisely the capabilities of SHALTOP, and to estimate rheological parameters depending on the lahar (DF or HF)
593 characteristics. This could be done with a catalogue of more recent lahars: their dynamics is better constrained thanks to the

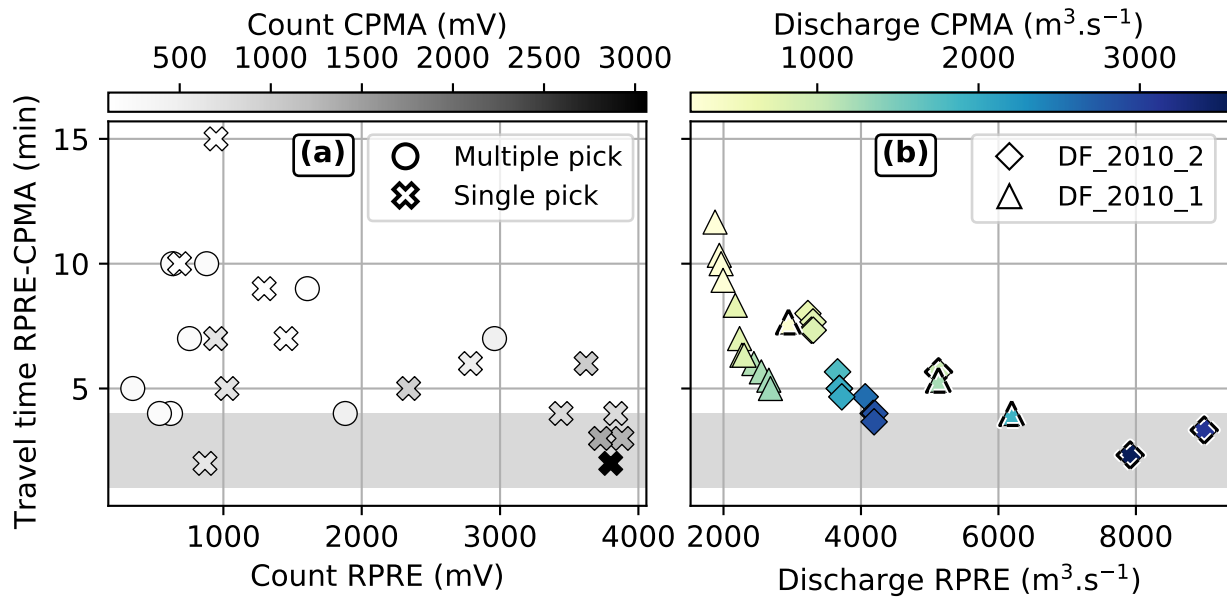


Figure 15. Flow travel durations between RPRE and CPMA deduced from AFMs recordings and simulations. (a) Dephasing between RPRE and CPMA FULL channel maximum, as a function of RPRE FULL channel maximum. Greyscale gives the maximum amplitude recorded on CPMA FULL channel. Picking is done manually for lahars with "strong intensity" between 2009 and 2011, from the database of [Aubaud et al. \(2013\)](#). Crosses: match between RPRE and CPMA FULL maximum is unambiguous. Circles: uncertain pick, with multiple maximums in FULL CPMA possibly matching one maximum in FULL RPRE. (b) Dephasing between maximum discharges at RPRE and CPMA in simulation, as a function of maximum discharge at RPRE. Colorscale gives maximum discharge at CPMA. Symbols give the simulation scenario. Symbols with dashed black contour indicate simulations where the Coulomb rheology is used (Voellmy otherwise). Friction coefficient is $\mu_S = \tan(2^\circ)$, $\mu_S = \tan(3^\circ)$ and $\mu_S = \tan(4^\circ)$. Turbulence coefficients range from 100 to 500 $m s^{-2}$. Grey patches give observation ranges for the Jun. 19, 2010 DF.

594 CCPA AFM that was installed in 2014.

595 8 CONCLUSION

596 In this work, we have modeled a rock avalanche, and the subsequent remobilization of the deposits as a high discharge debris
 597 flow, with a single thin-layer numerical code, SHALTOP. SHALTOP is used empirically, with a maximum of two rheological
 598 parameters (Coulomb or Voellmy rheology). We focus on extreme events, and in particular high discharge DFs, in a risk
 599 conservative approach. The simplicity of the modeling solution is compensated by an extensive use of field data to define
 600 realistic simulation scenarios and calibrate rheological parameters. By doing so, we can reproduce the main characteristics of
 601 extreme events, at least in a first approximation. We argue that more complex models may not necessarily yield better results.
 602 Although they can simulate more complex processes (such as erosion or variations in solid concentrations), they include more
 603 parameters that are difficult to determine, and whose number may improve artificially the quality of back-analysis.

604 Besides, we show that, in our simulations:

- 605 • For similar volumes, successive rock avalanches yield a better match between simulations and seismic signals, but do not

change the geometry of the simulated deposits.

- An instantaneous remobilization of the debris reservoir and a simple Coulomb rheology are sufficient to reproduce the main characteristics of a documented high discharge DF.
- For a DF of a given volume, a progressive remobilization of the debris reservoir slows down the DF, in comparison to an instantaneous release.

Our results pave the way to better quantifying flood hazard in the Prêcheur village, by identifying the areas at risk and potential overflows in adjacent gullies. Although we focused on modeling extreme events, we show that our simulated travel durations are consistent with observations for the main lahars (DFs and HF's alike) of 2009 and 2010. Thus, the construction of a simulation database with SHALTOP could also provide first order scaling laws between DF characteristics in the upper and lower parts of the river, which would be useful for real time monitoring. Further work is however needed to assess SHALTOP performance for smaller DFs and HF's, in comparison to observations and other more complex models. Future research could also investigate the relation between lahar initiation, volume and dynamics, and try modeling the continuous transition from a rock avalanche to a DF.

We considered rock avalanches and DFs in a volcanic context, but the sequence of these two kind of events is also relatively common in all mountainous areas (e.g. [Walter et al., 2020](#)). The methodology presented in this work can be, supposedly, extended to such contexts.

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10 AUTHOR CONTRIBUTION

Conceptualization, Methodology and Validation, M.P., C.L., Y.T., G.G., A.M., A-M.L., A.N., Y.L., J-M.S.; Formal Analysis, Investigation and Visualization M.P.; Resources and Data Curation, M.P, C.L., Y.T., A-M.L., A.N., Y.L., J-M.S., V.C., T.D., S.L.; Writing Original Draft Preparation, M.P.; Writing Review & Editing, M.P., C.L., Y.T., G.G., A.M., A-M.L., A.N., Y.L., B.V., J-M.S., V.C., T.D., F.F., M.M., S.L., J-C.K., A.L.F., A.L.; Supervision, Project Administration, and Funding Acquisition A.M., G.G., C.L., Y.T. A-M.L., B.V., J-C.K., A.L.F., A.L., F.F.

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