Effects of source faulting and fringing reefs on the 2009 South Pacific tsunami inundation in southeast Upolu, Samoa

Cyprien Bosserelle^{1,1}, Shaun Williams^{1,1}, Kwok Fai Cheung^{2,2}, Titimanu Simi^{3,3}, Yoshiki Yamazaki^{2,2}, Volker Roeber^{4,4}, Thorne Lay^{5,5}, Emily M. Lane^{6,6}, Ryan Paulik^{1,1}, and Lameko Simanu^{7,7}

¹National Institute of Water and Atmospheric Research
²University of Hawaii at Manoa
³Ministry of Finance
⁴Univ Pau & Pays Adour
⁵University of California Santa Cruz
⁶National Institute for Water and Atmospheric Research
⁷Ministry of Natural Resources and Environment

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Abstract

The subduction zone along the northern Tonga Trench has the highest plate convergence rate in the world, but limited records of its seismic and tsunamigenic activities. In 2009, a tsunami generated by an MW 8.1 earthquake doublet caused severe impacts in the region including damage and loss of life on the south shores of Upolu and Savaii Islands, Samoa. Here we use numerical modeling aided by recorded data and eyewitness accounts to evaluate which of the published source models in the Tonga Trench region most suitably represents the 2009 event for use in hazard assessment around Samoa. We show that only a few of the published sources are suitable to reproduce large inundation observed in Samoa and none reproduces runup as high as observed in areas that were most severely impacted on the southeast Upolu coast. The distribution and intensity of inundation is dependent on local topographic and bathymetric features, configuration of coastal geomorphology, and trapping of short-period waves over the reef flats. For one of the sources, comparison of the relative contributions of the normal and thrust faulting components of the doublet to the southeast Upolu inundation indicates that the initial intraplate normal faulting dominated the east-northeastward tsunami propagation and inundation compared with the subsequent interplate thrust faulting. Overall, two key source models are discussed and identified for future refinement.

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2	inundation in southeast Upolu, Samoa
3	Cyprien Bosserelle ¹ , Shaun Williams ¹ , Kwok Fai Cheung ² , Thorne Lay ³ , Yoshiki
4	Yamazaki ² , Titimanu Simi ⁴ †, Volker Roeber ⁵ , Emily Lane ¹ , Ryan Paulik ¹ , Lameko
5	Simanu ⁴
6	¹ National Institute of Water and Atmospheric Research (NIWA), New Zealand. ² Department of
7	Ocean and Resources Engineering, University of Hawaii at Manoa, Hawaii, USA. ³ University of
8	California Santa Cruz, California, USA. ⁴ Management Office, Ministry of Natural Resources and
9	Environment, Apia, Samoa. ⁵ de Pau et des Pays de l'Adour (UPPA), HPC-Waves Laboratoire
10	SIAME, Pau, France.
11	Corresponding author: Cyprien.Bosserelle@niwa.co.nz
12	†Current affiliation: Management Unit, Green Climate Fund, Samoa Ministry of Finance.
13	Key Points:
14	• The tsunami water level produced by the 2009 earthquake doublet is simulated for Samoa
15	and compared to observed runup on southeast Upolu.
16	• Geometry of the intraplate normal faulting near the northern Tonga Trench terminus
17	influences the extreme runup on southeast Upolu.
18	• The runup on southeast Upolu is also dependent on fine-scale topographic features and
19	wave excitation over the reef flats.

20 Abstract

The subduction zone along the northern Tonga Trench has the highest plate convergence rate in 21 22 the world, but limited records of its seismic and tsunamigenic activities. In 2009, a tsunami generated by an $M_W 8.1$ earthquake doublet caused severe impacts in the region including 23 damage and loss of life on the south shores of Upolu and Savaii Islands, Samoa. Here we use 24 numerical modeling aided by recorded data and eyewitness accounts to evaluate which of the 25 published source models in the Tonga Trench region most suitably represents the 2009 event for 26 27 use in hazard assessment around Samoa. We show that only a few of the published sources are suitable to reproduce large inundation observed in Samoa and none reproduces runup as high as 28 observed in areas that were most severely impacted on the southeast Upolu coast. The 29 distribution and intensity of inundation is dependent on local topographic and bathymetric 30 features, configuration of coastal geomorphology, and trapping of short-period waves over the 31 32 reef flats. For one of the sources, comparison of the relative contributions of the normal and thrust faulting components of the doublet to the southeast Upolu inundation indicates that the 33 initial intraplate normal faulting dominated the east-northeastward tsunami propagation and 34 inundation compared with the subsequent interplate thrust faulting. Overall, two key source 35 models are discussed and identified for future refinement. 36

37 Plain Language Summary

This study models the September 29, 2009 tsunami in Samoa and compares model outputs with tsunami water level observations, field data of runup and inundation as well as eye-witness accounts collected after the event occured. Whilst several studies have reconstructed the runup and inundation of this event in American Samoa, this is the first time such work has been undertaken in Samoa which lies to the west (formerly known as Western Samoa). We discuss the effects of the complex earthquake sequence that triggered the tsunami event, as well as influence

of nearshore coastal characteristics on the pattern and behavior of tsunami impacts observedonshore.

46 1 Introduction

Tsunami inundation in reef-fringed environments is highly influenced by the characteristics of the source event [e.g., *Li et al.*, 2016] as well as the nearshore geomorphology [e.g., *Tang et al.*, 2009; *Roeber et al.*, 2010; *Gelfenbaum et al.*, 2011; *Dilmen et al.*, 2018]. Local observations of nearshore and runup processes provide the evidence-base for refining and improving source parameters through forward tsunami modeling [e.g., *Yamazaki et al.*, 2018; *Davies*, 2019]. Such improvements can lead to more realistic tsunami hazard representations and

a better understanding of local processes for use in coastal management planning.

54 The 29 September 2009 South Pacific Tsunami (SPT) initiated from an M_W 8.1 earthquake

sequence involving nearly simultaneous intraplate normal faulting in the outer rise oceanic plate

and major interplate reverse faulting ($M_W 8.0$) occurring within two minutes of the initial rupture

57 [Beavan et al., 2010; Lay et al., 2010; Kiser and Ishii, 2012; Fan et al., 2016; Hossen at al.,

58 2017]. The USGS-NEIC epicenter (15.489°S, 172.095°W) for this M_W 8.1 event is located 150

59 km east of the 1917 $M_W \sim 8.0$ event (Figure 1). The combined faulting generated a tsunami across

60 the South Pacific that caused widespread inundation, damage and loss of life on the southern

61 coasts of Upolu and Savaii Islands in the Independent State of Samoa (herein Samoa), Tutuila

Island in American Samoa, and Niuatoputapu Island in northern Tonga [Okal et al., 2010; Goff

63 and Dominey-Howes, 2011].

In the aftermath of the event, extensive investigation of the tsunami source and wave

characteristics were undertaken based on seismic and GPS records, field surveys and numerical

66 modeling [e.g., Beavan et al., 2010; Lay et al., 2010; Okal et al., 2010; Roeber et al., 2010; Goff

and Dominey-Howes, 2011; Yamazaki et al., 2011; Zhou et al., 2012; Fan et al., 2016; Hossen et

al., 2017; *Dilmen et al.*, 2018; *Wood et al.*, 2018; *Han et al.*, 2019; *Williams*, 2019]. Previous

efforts to reconstruct tsunami characteristics in Samoa were limited to simulating wave initiation
and dispersion up to the reef edge [e.g., *Fritz et al.*, 2011; *Okal et al.*, 2011; *Franchello and Annunziato*, 2012; *Zhou et al.*, 2012]. Shoreward wave propagation and inundation was not
previously addressed due to lack of high-resolution topography and nearshore bathymetry for
this region. Recent acquisition of high-resolution LiDAR topography and nearshore bathymetry
[*FUGRO*, 2016] now enables re-evaluation of the event; modeling the tsunami initiation,
propagation, and inundation along southeast Upolu, Samoa.

76 This paper reconstructs the 2009 SPT to study the characteristics of wave initiation, propagation and inundation in Samoa, with emphasis on the observed runup particularly on southeast Upolu 77 Island. This coastline experienced the highest observed runup, flow depths, damage and loss of 78 79 life [Okal et al., 2010]. We assess whether current understanding of the rupture mechanism and propagation explains the extreme runup and flow depths observed in this region. We test rupture 80 81 source mechanisms compiled by Fan et al. [2016], Hossen et al. [2017] and Han et al. [2019] against the Deep Ocean Assessment and Reporting of Tsunami (DART) data closest to the 82 83 source, Apia tide gauge measurements, and runup measurements from post-disaster surveys [e.g. Okal et al., 2010, 2011; Dudley et al. 2011; Reese et al., 2011]. 84

We consider 14 source representations of the 2009 SPT that are obtained directly or adapted from published models. The overall objective is to identify which of the 14 source representations produce the most realistic inundation and runup for future refinement (e.g. in finite-fault models) and use in hazard exposure and impacts assessment. The importance of validating models with surveyed runup, as well as quantifying the effects of coastal topography on wave behavior, are discussed.

91 2 Regional context and setting

92 The Samoan archipelago comprised of Samoa and American Samoa is situated along a
93 volcanic hotspot island chain [*Koppers et al.*, 2008], located approximately 100 km north of the

94 terminus of the Tonga Trench. The tectonics and seismicity of this region are illustrated in Figure 1. The Pacific Plate undergoes strong deformation as it subducts steeply beneath the Tonga Arc, 95 while north of the hook-like terminus the plate continues westward. The bending and tearing of 96 the plate at the terminus [e.g., Millen and Hamburger, 1998; Govers and Wortel, 2005] coupled 97 with mantle plume upwelling [e.g., Koppers et al., 2011; Konter and Jackson, 2012; Chang et 98 99 al., 2016], produces complex strain and faulting that results in significant seismic hazard for the Samoan archipelago [e.g., Gurnis et al., 2000; Fan et al., 2016]. The average convergence rate of 100 101 25 cm/yr between the Pacific plate and the overriding Tonga block is the highest among all subduction zones [Bevis et al., 1995; Basset et al., 2016]. The rate is enhanced by rapid back-arc 102 spreading of the Lau Basin combined with roll-back of the subduction zone [Chang et al., 2016]. 103 104 Despite the high convergence and seismicity rate [Basset et al., 2016], recorded interplate thrust 105 faulting activity only involves events with magnitudes less than ~8.0 [e.g., Meng et al., 2015]. 106 Intraplate earthquakes within the deformed Pacific Plate, such as the 26 June 1917 Samoa (M_m 8; $M_{\rm S}$ 8.7) and 2009 Samoa-Tonga earthquakes located along the bend in the arc (Figure 1), are the 107 largest recorded seismic events in the region [e.g., Okal, 1992; Meng et al., 2015] and both have 108 triggered major tsunamis [Okal et al., 2011]. 109

There is general agreement that the overall source of the 2009 SPT involved both interplate 110 thrust-faulting and intraplate normal-faulting. Beavan et al. [2010] used campaign geodetic 111 observations and approximate tsunami arrival time reports to infer that a slow tsunami-112 earthquake on the plate boundary interface triggered trench-slope intraplate normal-faulting. Lay 113 et al. [2010] used regional and teleseismic waveform modeling to develop a model where 114 intraplate normal-faulting occurred first, triggering interplate thrusting events 49 and 90 s after 115 the initial rupture, with the timing of all large moment release well constrained by the seismic 116 signals. Analysis of two-event centroid-moment tensor solutions by Duputel et al. [2012], Nealy 117 118 and Hayes, [2015], and Fan et al. [2016] support the latter interpretation, but the possibility of a precursory weak, very long-duration initial slip event on the thrust plane has not been ruled out. 119

120 Modeling of seismic waves for the intraplate normal faulting has invoked either a southwestward dipping plane [e.g., Lay et al., 2010; Nealy and Hayes, 2015] or an east-northeastward dipping 121 plane [e.g., Li et al., 2009; Beavan et al., 2010; Nealy and Hayes, 2015; Fan et al. 2016]. The 122 latter orientation is compatible with modeling of the reconstructed tsunami source area [e.g., 123 Hossen et al., 2017] and with models of geodetic and gravity changes including viscoelastic 124 relaxation [e.g., Han et al., 2019]. However, the precise strike and dip of either candidate fault 125 plane is ambiguous because of frequency dependence of the long-period point-source moment 126 127 tensors and their associated best-double couple fault plane candidates [Lay et al., 2010]. Modeling an eastward dipping normal fault implies an increased slab dip and dominance of the 128 oceanic plate bending moment relative to decreased in-plane force, and greater extensional in-129 shear stress causing a negative dynamic topography response seaward of the trench axis [e.g., 130 *Craig et al.*, 2014]. 131

132 While detailed orientation and timing of the complex faulting is not fully resolved, there is agreement that the resulting tsunami was triggered by both normal and reverse faulting. Roeber 133 et al. [2010] attributed the impacts of the tsunami along the shores of Tutuila Island (American 134 Samoa), which lies approximately 97 km east of Upolu, to shelf resonance and tsunami 135 amplification at the natural periods of bathymetric features. The near-shore bathymetry, 136 137 generally characterized by steep slopes and narrow reef morphology on the insular shelf, was suggested to modulate the incident tsunami waves resulting in varying flow conditions around 138 the island. Wood et al. [2018] later showed that multiple waves with similar amplitudes impacted 139 most locations in American Samoa rather than a single large wave. Nevertheless, fringing reef 140 morphology and narrow embayments are important factors in explaining the variations in runup 141 and flow depth on Tutuila [Roeber et al., 2010; Gelfenbaum et al., 2011; Dilmen et al., 2015, 142

143 2018]. In Samoa, such analysis of wave characteristics relative to the magnitude of inundation
144 observed has not previously been carried out and is the focus of this paper.

145 **3 Methods and Data**

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3.1 Model Configuration and Grid Scheme

The shock-capturing hydrodynamics model BG-Flood (Block-adaptive on Graphics 147 processing unit Flood model) is used to reconstruct the 2009 tsunami from the seismic source to 148 149 inundation (Figure 2). BG-Flood simulates shallow water hydrodynamics on Graphics Processing Units (GPU) using an Adaptive Mesh Refinement type grid. The model governing 150 equations are from the Saint-Venant, Shallow Water Equation (SWE) solver of Basilisk 151 [*Popinet*, 2011], but uses a Block Uniform Quadtree mesh that runs efficiently on the GPU [e.g., 152 Vacondio et al., 2017]. The SWE engine is well established and tested in Basilisk through 153 154 tsunami benchmark evaluation, with the solver and its ability to define propagation and inundation described by *Popinet* [2012]. The model, however, does not include dispersive terms 155 (present in Boussinesq-type or non-hydrostatic models). While the model is well suited for 156 testing the range of source scenarios to identify which one/s provide the most representative 157 modelled to observed inundation and runup comparisons, it may not be capable of reproducing 158 159 some of the high frequency tsunami waves and their nearshore behavior.

The BG-Flood model is configured using three nested uniform Cartesian grids to simulate the 160 161 2009 tsunami propagation from initiation to inundation (Figure 2). The level-1 grid extends across the northern region of the Tonga Subduction Zone at a resolution of 250 m to capture 162 DART 51425, DART 51426 and DART 54401, as well as wave propagation around the Samoan 163 Islands. The level-2 grid resolves wave inundation on Upolu, Manono (immediately west of 164 Upolu) and Savaii Islands at 50 m resolution for comparison with surveyed runup in these 165 166 islands [Okal et al. 2010] and includes the Apia tide gauge. The level-3 grid solves for inundation at 10 m resolution on southeast Upolu for more detailed comparison with surveyed 167

runup in this area [*Okal et al.* 2010; *Reese et al.* 2011; *Dudley et al.* 2011]. The digital elevation
models (DEM) used in these simulations were derived from: 1) the Generic Bathymetric Chart of
the Oceans (GEBCO) at ~900 m resolution; 2) multi-beam bathymetry survey over the shelf
slopes at a resolution of ~60 m [*Kruger and Kramer* 2008]; and 3) Light Detection and Ranging
(LiDAR) topography and nearshore bathymetry (down to 30 m depth) at a resolution of 5 m
[*FUGRO*, 2016].

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3.2 Source Models, Observations Data and Comparison

Instrumental records (DART buoys and tide gauge data) and runup measurements were used to validate the tsunami simulations. Clear near-field signals of the 2009 SPT were recorded at the three DART stations used in this study as well as the Apia tide gauge. Raw DART data were processed using a Gaussian filter with a 216 s window; tsunami spectral density was calculated from the filtered data resampled to a 10 s interval using the Welch's method with a 512 sample window.

For the Apia tide gauge, the tsunami signal was extracted by removing the tide signal from
recorded water level. The tide signal was first predicted using the T-tide software [*Pawlowicz et al.* 2002], and using the full tide gauge record from 1993—2019.

Surveyed runup and flow depth were first converted to maximum water level for comparisons.
For the whole country analysis, 30 maximum water level points were used in Savai'i, 127 points
in Upolu and 7 points in Manono. For the southeast Upolu comparison, 65 maximum water level
points were available.

Fourteen source scenarios with different rupture parameters, timing, locations, angles, dips and slip were simulated and compared with the open-ocean DART observations, Apia tide gauge record and surveyed runup (Table 1). This includes a modified source, which applies the rupture dimensions and slip of *Lay et al.* [2010] but with an east-northeast dipping normal fault [e.g.,

Fan et al., 2016; *Hossen et al.*, 2016]. Fault 1 denotes the normal intraplate rupture in the outer
rise and faults 2 and 3 correspond to the later interplate thrusting on the plate boundary (Figure
2) (Table 1).

Source parameters include the rupture initiation and dislocation rise time to reconstruct the time 195 history of seafloor deformation through the half-space solutions of Okada [1985; 1992] for 196 modeling of tsunami generation. The intraplate rupture occurred below the 6000 m deep abyssal 197 seafloor, while the water depth above the thrust faulting extends from 2000 to 7000 m over the 198 199 margin. The 8300 m deep Tonga Trench lies between the locations of the two sub-events and played a significant role in modifying the near-field tsunami through wave refraction associated 200 with rapid variation of the propagation speed. This suggests that the timing of rupture sequencing 201 202 and rise time (or rupture duration) are critical in accurately simulating the inundation.

To evaluate how each source configuration performs for wave propagation and inundation, wewe 203 204 compare the index of agreement between the modelled and observed tsunami wave signal at the DART buoys and the Apia tide gauge. We also compare modelled and observed runup, first for 205 206 the entire country and then for southeast Upolu only. Each scenario was simulated from initiation to propagation to the DART locations for a total of 4.2 hours for our level-1 grid. The country 207 inundation grid (level-2) and southeast Upolu grid (level-3) were also run for 4.2 hours. The 208 goodness of the fit between computed and observed maximum inundation (i.e., combined runup 209 and maximum flow elevation) is estimated using the Willmot index of agreement(Willmot 2012) 210 211 (Eq. 1):

212
$$IoA = \begin{cases} 1 - \frac{\sum_{i=0}^{n} |P_i - O_i|}{c \sum_{i=0}^{n} |O_i - \overline{O}|}, when \sum_{i=0}^{n} |P_i - O_i| \le c \sum_{i=0}^{n} |O_i - \overline{O}| \\ \frac{c \sum_{i=0}^{n} |O_i - \overline{O}|}{\sum_{i=0}^{n} |P_i - O_i|} - 1, when \sum_{i=0}^{n} |P_i - O_i| > c \sum_{i=0}^{n} |O_i - \overline{O}| \end{cases}$$
(Eq. 1)

where $O_i O_i$ is the measured data, P_i is the simulated data, the over bar denotes the average, and c = 2. As general guidance when comparing the complex nature of tsunami with idealized simulation, a negative IoA is poor agreement, IoA<0.25 is weak agreement, IoA>0.5 is good
agreement and IoA >0.8 is excellent agreement.

3.3 Simulation and Computation

The simulation applies the Apia tide level at the time of initiation but does not include the tide variation throughout the simulation. With the largest waves occurring within 15-20 min after the earthquake the tidal variation is not expected to influence the tsunami significantly. Each grid is nested one-way at a 1s interval. Outputs at the DART locations and Apia tide gauge are extracted at the model computational time step (~0.7 s at the DART buoys and ~0.1 s at the tide gauge). Maximum water level and flow depths are updated at each step and compared with observations data from *Okal et al.* [2010].

A uniform quadratic bottom roughness (cf = 0.0001) was applied to the 250 m and 50 m grids. A 225 226 roughness formulation of Smart [2017] is used which is related to the Manning formulation but is also valid for very small flow depth without requiring a depth-dependent roughness length 227 (z₀). A simple spatially varying roughness map with low roughness length ($z_0 = 0.0001$ m) was 228 applied to elevations lower than 15 m below mean sea level, and a high roughness length ($z_0 =$ 229 0.01 m) used for all other areas including coral reefs, vegetated land, and built environment. This 230 231 is reasonable given the relatively low density of buildings and infrastructure along the southeast Upolu coastline. 232

4 Results and Interpretation

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4.1 Comparison of different sources

The simulation of 14 different sources from tsunami generation to inundation provides a wide range of tsunami scenarios whose characteristics are compared with observations. The index of agreement of simulated tsunami against observation (Table 2) as well as visual

comparison of tsunami wave signal at DART buoys and tide gauge (Supporting Information S1S14) are used to summarize the results.

None of the 14 sources produce excellent agreement (i.e. no IoA exceeds 0.8) and no sources 240 produce consistently good agreement against all datasets (i.e. the DART data, the tide gauge and 241 the runup data) (Table 2). Sources labeled Lay et al. [2010] Version D and E, as well as Nealy & 242 Hayes [2015] Version T produce moderate to good agreements at the two closest DART buoys 243 as well as the Apia tide gauge but do not reproduce observed runup in South East Upolu. On the 244 other hand, Lay et al. [2010] Version C, a modified iteration of Lay et al. [2010] Version B for 245 testing in this study, and *Hossen et al.* [2017] produces good agreement with observed runup but 246 cannot reproduce the DART and tide gauge as well as other sources. With the DART buoys 247 located far away from the main impact of the tsunami and recording waves in the order of 0.05 248 m, we deem it more important for a hazard assessment to better match the Apia tide gauge 249 250 recording (~1. 0m tsunami) and runup (up to 15 m above mean sea level).

Overall *Hossen et al.* [2017] and *Lay et al.* [2010] Version C, while imperfect, are most consistent with the observed tsunami data. These two sources are discussed further in the sections below.

254 **4.2 Propagation**

BG-Flood simulates the full complexity of the rupture sequence in the tsunami generation 255 to reconstruct the near-field wave pattern. The computed water levels for *Lay et al.* [2010] 256 Version C are shown in Figure 3 at key moments after the rupture initiation time (see also 257 Supporting Information Animation S15). The normal faulting generates a sea surface trough of 3 258 m heading to the northeast, while the uplift of the thrust faults adds to the southwest traveling 259 positive wave from the normal fault to produce a surface pulse of 3.8 m amplitude. The relative 260 261 position of the faults determines the radiated wave pattern across the ocean. The trough generated by the normal faulting reaches Upolu 11 min after the earthquake. It is immediately 262

followed by a positive wave from upswing of the initial surface depression and a larger and
broader wave generated by the thrust faulting (Figure 3).

The second wave crest reaching the shore of southeast Upolu is augmented by reflection of the first to produce the highest water level 19 min after the earthquake, while standing waves gradually develop over the wider shelf to the west producing a large peak 9 min later. The wave pattern is reversed with a leading crest followed by a deep trough toward the west and south. The maximum surface elevation in Figure 3 shows a complex pattern with focusing of the waves behind islands and seamounts in the general propagation direction.

Tsunami simulations using both *Lay et al.* [2010] Version C and *Hossen et al.* [2017] adequately reproduces the arrival time at both DART locations and the tide gauge. Computed and recorded waveforms and spectra at the three DART buoys and the Apia tide gauge for these scenarios are shown in Figure 4, and compared in Figure 5. The amplitude of the DART observations are overpredicted, but because they are low amplitude, this is considered secondary. The waveform at Apia tide gauge is well modeled, but the amplitude is also overpredicted for this model. Comparisons for other models are shown in the Supporting Information S1-S14.

DART 51425 (northwest of the epicenter) and the Apia tide gauge (northeast of the epicenter) both recorded a leading trough from the normal faulting in the outer rise. At DART 51426 (southeast of the epicenter), a leading crest followed by a deep trough from the combined thrust and normal faulting was recorded. The validation of the DART and Apia tide gauge records corroborates the work of *Dilmen et al.* [2018] and *Wood et al.* [2018] in which a leading trough was observed at the Pago Pago Tide Gauge in Tutuila and at DART 51425 in the north, and a leading crest at DART buoys 51426 and 54401 south of the epicenter.

4.3 Computed and Recorded Inundation

Comparison between simulated and observed runup (Figure 6) shows the variable response of the tsunami along the shore. Some scenarios indeed produce high runup where observed. However, none of the 14 scenarios tested reproduce the extreme runup of 10-15 m on southeast Upolu. These extreme runups occurred East of Lepa, at the apex of Saleapaga bay and on the Lalomanu cliffs. Simulated maximum water levels show localized "spikes" in the runup located in the same area of the observed extreme runups suggesting that these were caused by highly localized amplification of the incoming tsunami wave..

Trapping and slow drainage of the initial wave within the narrow lagoon appear to provide a 293 294 higher baseline for subsequent incoming waves to ride in on causing higher flow depths along this coast compared with other locations which were inundated. While southeast Upolu exhibits 295 the highest overall runup in each scenario, the inability of any of the models to reproduce the 296 297 extreme runup peaks suggests: 1) potential limitations of the DEM resolution in representing 2009 topography and bathymetry in the near- and on-shore; 2) complexity in the source 298 299 mechanism which is not represented using uniform slip models (e.g., heterogenous slip); and/or 3) dispersive effect may be propagating waves with a period that better match the resonance 300 mode of the location of extreme runup. Nevertheless). Nevertheless, the computed maximum 301 302 runup for Lay et al. [2010] Version C and Hossen et al. [2017] scenarios are generally consistent with recorded observations on all three islands, and consistent with eye-witness accounts in 303 304 southeast Upolu whereby the second and third waves were generally observed to be the largest [Okal et al., 2011; Dudley et al., 2011]. 305

The simulation of *Lay et al.* [2010] Version C and *Hossen et al.* [2017] scenarios confirms the large variation observed in runups and flow depths along the shore as shown for Lepa and Saleapaga. At these locations, waves at the shoreline transform from a single peak reaching 6 m elevation at Saleapaga to a single peak of 4 m in Lepa 700 m to the west (Figure 7). A double

310 peak exceeding 5 m is observed 1000 m further west of Lepa. The waves around Saleapaga appear to have excited the local resonant frequencies over the reef flats [e.g., Abe, 2011], and 311 showed the strongest onshore wave energy at these frequencies which is consistent with 312 relatively high flow depths up to 6 m observed in this area (Figure 6b) [Okal et al., 2010]. The 313 strongest offshore energy signal observed at lower frequency from the initial pulse and higher 314 315 onshore energy peaks at lower frequencies are probably due to interference and excitation over the reef flats and onshore [e.g. Gelfenbaum et al., 2011; Cheung et al., 2013; Gawehn et al., 316 2016; Buckley et al., 2018]. 317

318 **5 Discussion**

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5.1 Complex source in complex bathymetry

The 2009 Samoa tsunami was unusual because it was triggered by an intraplate-interplate sequence of earthquake ruptures. The subsequent tsunami waves were transformed and trapped over complex bathymetric features significantly amplifying them as they reached and inundated the Samoan shoreline. The high resolution of the simulation presented here reveals the tsunami waves transformation as they reach the complex reef bathymetry.

While our modelling did not reproduce the extreme runup of 10-15 m observed on southeast 325 Upolu, the highest modelled runups of ~ 9 m were exhibited on this coast. Extreme wave 326 amplification and runup between Lalomanu and Salani on southeast Upolu is caused by very 327 328 localized reef features (e.g. incisions and gullies) at the edge and inside of the narrow fringing reef morphology. These focus tsunami energy towards adjacent shorelines and exacerbated runup 329 by trapping energy between the reef and the onshore cliff (Figure 8). Interference and 330 convergence between reflected and incoming waves explain the local amplification and runup 331 heights along the southeast coast. Roeber et al. [2010] and Dilmen et al. [2018] modelled similar 332 333 amplifications due to nearshore reefs in Tutuila exciting resonant modes of 2-4 min on the insular shelf surrounding American Samoa. Our results show that the relative small change in the 334

incoming tsunami waves and interaction with the narrow reef morphology played a key role in
 amplifying the 2009 SPT on southeast Upolu.

In general these observations support suggestions that high-frequency tsunami waves on reef flats [e.g., *Gawehn et al.*, 2016], as well as geomorphologic configuration and fine-scale topographic features within the reef system [e.g., *Gelfenbaum et al.*, 2011], are crucial local drivers of extreme runup and coastal flooding. Furthermore, they support suggestions by *Baba et al.* [2009] that narrow fringing reefs do not have the width to dissipate wave energy compared with wider reefs. Although, in some circumstances, wider reefs may increase the wave height through resonant interactions.

Interestingly, in the case of Lay et al. (2010) version C scenario, waves triggered by the normal 344 fault rupture alone can explain most of the large runup observed in most of southeast Upolu 345 (Figure 9). However, the normal fault alone can only explain 40 to 70 % of the maximum water 346 347 level offshore, and does not adequately account for the observed runup in Lepa or in the area between Poutasi and Satalo, or the east-facing region of the coast. Longer wave periods 348 349 associated with the thrust rupture are amplified more significantly by the local bathymetry at these locations compared with the shorter period waves produced by the normal fault. This 350 highlights how shorter tsunami waves get more easily trapped on narrow reefs and longer 351 352 tsunami waves are more easily trapped on wider reefs. Nevertheless, our observations corroborate that tsunamis are highly sensitive to source parameters, with significant spatial 353 354 inundation variability for a given source. In addition, the observations suggest that 1-dimensional tsunami attenuation modeling of offshore wave heights at coast, which is sometimes used to 355

estimate tsunami inundation hazard, is not adequate for use in reef fringed islands [e.g. *Smart et al.*, 2015].

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5.2 Refinement of source mechanism

Previous investigations of the 2009 tsunami source were unable to simulate inundation in 359 360 Samoa due to the lack of high resolution bathymetric and LiDAR topographic data. The analysis presented in this study shows that each source generates a significantly different tsunami. The 361 Lay et al.[(2010] Version C and Hossen et al. [2017] sources produce the most consistent 362 agreement with tsunami records (observed runup in Upolu, Manono and Savaii and the recorded 363 at the Apia tide gauge and DARTs). However, the inability of any of the scenarios tested to 364 adequately capture the extreme runup (~14 m) observed at Lepa on southeast Upolu implies a 365 consistent deficiency in the sources tested. 366

367 This deficiency is unlikely to be resolved by testing additional uniform slip scenarios. For example, decreasing the strike of the normal fault, could focus more energy and send larger 368 369 waves towards southeast Upolu, but this would also lead to larger waves reaching the Apia tide 370 gauge, worsening the agreement there. Doing so would also decrease the tsunami wave reaching American Samoa. This would be inconsistent with previous findings that showed that even 371 372 simple sources involving only the normal fault could reproduce runup and tide gauge 373 measurement in Pago Pago [Roeber et al. 2010; Dilmen et al. 2018]. It is also likely that a Boussinesq-type model will improve simulation of extreme runup. This will allow weakly 374 dispersive (high-frequency) waves to propagate from the source to Samoa and allow 375 amplification at these frequency to be captured at the coast.. These extreme runup could also be 376 377 due to swashing of the incoming wave or channeling of waves along gullies that are not 378 adequately represented in the DEM [e.g., Smart et al., 2018] and 10 m model resolution. It is also possible the LiDAR is not representative of pre-event landscape prior to post-disaster 379 reconstruction and modification of local topography. In this case, wave transformation over the 380

reef and trapping within the reef flats, which is dependent on local topography and a key driverof the observed runup, would not be adequately represented.

383 Heterogeneous rupture with variable slip distribution could also reconcile different issues in the scenarios used here. For example, a finite-fault model of the normal fault presented in Lay et al. 384 [2010] could indicates larger slip patches west and east of the epicenter. This could provide 385 additional tsunami energy towards Upolu while maintaining high tsunami energy towards 386 Tutuila and limiting the amount of wave energy traveling to the Apia tide gauge in between the 387 two islands. The higher frequency waves generated from a variable slip rupture may excite the 388 resonance mode in Lepa and Lalomanu and might reproduce extreme runup. Hence, future 389 investigation using a dispersive model and considering heterogeneous rupture with variable slip 390 391 distribution, as well as stochastic source modeling and higher resolution simulation grid [e.g. Li et al., 2016; De Rissi and Goda, 2017; Yamazaki et al., 2018; Davies et al., 2019], might 392 393 improve the inundation and runup presented in this study.

Notwithstanding, the modelled inundation shows high spatial variability of runup and flow
depths similar to surveyed observations, with areas of highest runup observed in southeast
Upolu. Our results suggest variability is driven by the rupture source configuration and localized
tsunami amplifications due to complex bathymetry within the fringing reef and along coastal
topography.

This study highlights that good fits to the DART buoy data does not always guarantee accurate representation of runup, particularly where the buoys are situated away from the main tsunami energy beams. Nevertheless, the study demonstrates the use of tsunami runup simulation to model surveyed and instrumentally recorded coastal and deep-ocean observations to better constrain the event source. This approach, however, is highly reliant on available high-resolution topography and nearshore bathymetry data for tsunami-affected shorelines. Future LiDAR topography and bathymetry acquisition for the northern Tonga Islands could provide a basis to

406 re-evaluate the 2009 tsunami event following the above recommendation and and expand

407 reconstruction of propagation and inundation to all three impacted island groups: Samoa,

408 American Samoa, and northern Tonga. Using the runup information from all three countries

409 would provide better constraints for validation of the finite-fault model of the tsunami source.

410 6 Conclusions

This study reconstructed the 2009 tsunami in Samoa to observe wave characteristics and 411 evaluate the physical factors affecting the observed inundation in southeast Upolu. We applied 412 BG-Flood to simulate the tsunami from initiation through propagation to inundation for 14 413 possible sources for the event and evaluated agreement against instrumental records and 414 415 observed runup. Two source models performed better overall with moderate agreement with the DART and Apia tide gauge records, and good agreement with observed runup on Upolu, 416 Manono and Savaii islands. While our model reproduces the runup along most of southeast 417 Upolu (9 m), it does not capture the observed extreme value (14 m). We speculate that this is due 418 to source complexities which are not represented using a uniform slip model, and limitations in 419 420 the selected mode as well as the representativeness of near- and on-shore DEM. Analysis of the relative contributions of the normal and thrust faulting for this doublet event on tsunami behavior 421 indicates that the normal fault alone can explain approximately 70% of the observed runup and 422 423 flow depths up to 5 m on southeast Upolu.

Our findings indicate that inundation observed in Samoa is sensitive to the source rupture in addition to localized topographic and bathymetric features. The source details control the frequency contents of the tsunami and the fringing reefs control the resonance modes and periods. If the two constructively interfere, as was the case for the 2009 event along southeast Upolu, the impacts become much more severe than expected. The narrow fringing reef and constricted onshore geomorphology on southeast Upolu also plays a major role in exacerbating wave runup and flow depth along this coast. Future investigation and simulation of runup in all

three island groups impacted by the event (Samoa, American Samoa and Tonga) will enable

iterative refinement of the source mechanism and nearshore effects to regional inundation

433 distribution.

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440 at (https://github.com/CyprienBosserelle/BG_Flood), and the code version used in this research is

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data used in this study were obtained from the U.S. National Oceanic and Atmospheric

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447 https://www.ngdc.noaa.gov/hazard/dart/2009samoa_dart.html. Tide gauge data was obtained

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455

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- 615 **Table 1.** Fault parameters for the main scenarios tested for runup in southeast Upolu. Fault
- 616 parameters from original publications (length, width and slip) have been modified to produce 617 comparable output.

No.*	Length	Width	Strike	Dip	Rake	Slip	Centroid of rupture		Rupture	Rise	
	(km)	(km)	angle	angle	angle	(m)	Depth	Long.	Lat.	initiation	time
Liet al (2009)											
1	130	50	340.7	45.4	-46.3	4.3	12	172.5	15	15.7	46.6
2	50	50	183.7	13.9	69.3	15.8	10	172.5	16.1	91.6	50.8
Beaven et al. (2010)											
1	130	, 50	352	53	-32	3.1	12	172.2	15.5	0.0	0.0
2	50	50	173	16	75	11.2	12	172.8	16	30	20
Lay et	al. (2010) (Versio	on A)								
1	130	50	152.0	67.0	-77.0	6.1	18	172.0	15.5	3.0	60.0
2	50	50	185.0	29.0	90.0	5.6	18	172.4	16.0	49.0	40.0
3	50	50	185.0	29.0	90.0	5.6	18	172.4	16.0	90.0	40.0
Lay et al. (2010) (Version B)											
1	130	50	144.0	65.0	-91.0	6.1	18	172.0	15.5	3.0	60.0
2	50	50	185.0	29.0	83.0	4.7	18	172.3	15.8	49.0	40.0
3	50	50	185.0	29.0	83.0	4.4	18	172.3	16.0	90.0	40.0
Lay et	al. (2010) (Versio	on C)								
1	130	50	324.0	65.0	-85.0	10.2	18	172.0	15.5	3.0	60.0
2	50	50	175.0	29.0	90.0	4.7	18	172.4	16.0	49.0	40.0
3	50	50	180.0	29.0	90.0	4.7	18	172.4	16.0	90.0	40.0
Lay et	al. (2010) (Versio	on D)								
1	110	35	340	35	265	6.0	15.2	172.00	15.50	3.0	60.0
2	50	75	175	20	90	4.62	17.8	172.70	15.78	49.0	40.0
3	50	75	180	20	90	4.71	17.8	172.68	16.27	90.0	40.0
Lay et	al. (2010) (Versio	on E)								
1	110	35	320	35	265	6.0	15.2	172.00	15.50	3.0	60.0
2	50	75	175	20	90	4.62	17.8	172.70	15.78	49.0	40.0
3	50	75	180	20	90	4.71	17.8	172.68	16.27	90.0	40.0
Duput	el et al. (2	2012)									
1	130	50	157.0	64.0	-70.0	4.8	16	172.2	15.1	32.4	53.6
2	50	50	176.9	10.9	79.4	13.3	16	172.7	16.3	105.4	40.0
Nealy & Hayes (2015) (Version F)											
1	130	50	349.9	45.9	-84.6	9.2	16	172.2	15.2	33.0	66.0
2	50	50	158.1	39.6	51.7	19.5	16	172.2	15.2	45.0	80.0
Nealy & Hayes (2015) (Version U)											
1	130	50	356.0	40.1	-78.5	9.6	31	171.8	15.6	33.0	66.0
2	50	50	162.4	47.1	57.9	20.2	24	171.9	15.5	45.0	80.0
Nealy & Hayes (2015) (Version T)											

1	130	50	330.0	44.0	-81.0	4.3	18	172.1	15.5	25.0	50.0
2	50	50	178.5	18.0	59.0	14.3	18	172.9	16.4	91.0	182.0
Nealy & Hayes (2015) (Version C)											
1	130	50	153.0	61.5	-81.6	4.8	18	171.9	15.4	3.0	60.0
2	50	50	171.3	23.5	51.1	12.0	18	172.2	15.5	49.0	40.0
Fan et al. (2016)											
1	130	50	344.0	72.0	-68.0	4.3	12	172.2	15.0	0.0	70.0
2	50	50	183.0	15.0	40.0	15.8	12	172.5	16.1	80.0	80.0
Hossen et al. (2017)											
1	130	50	315.0	25.0	-99.0	6.1	18	172.0	15.5	3.0	60.0
2	50	50	180.0	29.0	90.0	7.9	18	172.4	16.0	49.0	40.0

⁶¹⁸ * Fault 1 normal fault; Fault 2 reverse fault; Fault 3 reverse fault.

619 ** Initial rupture in seconds after the earthquake time (2009-09-29 17:48:10 UTC).

Table 2. Evaluation of the main scenarios against offshore and nearshore tsunami observations. The evaluation is given as a correlation coefficient calculated as in Eq.1

Study	DART- 51425	DART- 51426	DART- 54401	Apia tide Gauge	Samoa Runup	South- east Upolu runup
Li et al. (2009)	0.27	0.33	0.43	0.47	0.52	0.23
Beaven et al. (2010)	0.36	0.4	0.41	0.55	0.4	0.13
Lay et al. (2010) (Version A)	0.33	0.39	0.46	0.16	0.56	0.35
Lay et al. (2010) (Version B)	0.37	0.44	0.47	0.36	0.61	0.40
Lay et al. (2010) (Version C)	0.21	0.14	0.2	0.22	0.67	0.56
Lay et al. (2010) (Version D)	0.48	0.48	0.47	0.49	0.45	0.14
Lay et al. (2010) (Version E)	0.48	0.47	0.46	0.41	0.6	0.37
Duputel et al. (2012)	0.25	0.36	0.40	0.36	0.51	0.19
Nealy & Hayes (2015) (Version F)	-0.21	0.10	0.36	0.26	0.51	0.61
Nealy & Hayes (2015) (Version U)	0.13	0.13	0.42	0.42	0.64	0.55
Nealy & Hayes (2015) (Version C)	0.21	0.28	0.43	0.54	0.5	0.20
Nealy & Hayes (2015) (Version T)	0.33	0.37	0.35	0.30	0.45	0.17
Fan et al. (2016)	0.18	0.17	0.39	0.40	0.56	0.24
Hossen et al. (2017)	0.38	0.41	0.49	0.25	0.63	0.52



Figure 1. The northern Tonga subduction zone and the major Samoan Islands. The Pacific 627 Plate underthrusts the Tonga block of the Australian Plate along the Tonga Trench. Spreading 628 of the Lau Basin back-arc increases the convergence rate across the Tonga Trench to about 629 25 cm/yr in this region. The moment tensor solutions from the Global Centroid Moment 630 Tensor catalog for $M_W \ge 6.0$ from 1976-2020 are shown. The star indicates the estimated 631 location of the June 26, 1917 ($M_W \sim 8.0$) earthquake (ISC-GEM). The September 29, 2009 632 event involved an M_W 8.1 outer-rise normal faulting and two interplate thrust faulting events 633 with a combined $M_W 8.0$ [Lay et al., 2010]. 634 635



440000 425000 435000 445000 -175° -170° -165° 420000 430000 450000 455000 636 175° 180° Figure 2. The model region for the 2009 South Pacific Tsunami. (a) Regional view with 637 level-1 (250 m resolution) and level-2 (50 m resolution) computational domain outlined in 638 black and DART buoy locations (red circles). Red star indicates epicenter. Rupture area of 639 the 3 faults is delineated in white, with locations for faults 2 and 3 overlapping. (b) Level-2 640 computational domain showing the insular shelves of the Samoan Islands, the Apia tide 641 gauge, and outline of the level-3 computational domain. (c) Level-3 computational domain 642 showing the nearshore fringing reef system within the insular shelves on southeast Upolu. 643



Figure 3. Snapshots of computed sea surface elevation after earthquake initiation, regional 645 maximum surface elevation and inundation. The relative position of the initial surface 646 disturbances gives rise to a trough leading the tsunami waves toward the north and east and a 647 leading crest to the west and south. (a) Completion of the 3-fault rupture sequence (fault 2 648 and 3 have overlapping locations) note the wave generated from the normal fault is already 649 bouncing back to a positive wave and the thrust faulting adding to the initial crest of the 650 normal fault. (b) Arrival of the leading trough at coast on southeast Upolu. (c) Arrival of the 651 combined first crest at coast on southeast Upolu. (d) Regional maximum water level. (e) 652 Maximum inundation in the Samoan islands at 50m resolution. 653





Figure 4. Comparison of recorded (black) and computed (red and blue) time series and spectra of surface elevations at DART water-level stations and the Apia tide gauge. The model reproduces the leading trough at DART 51425 to the north and the leading crest at DART 51426 to the south to confirm the two-stage rupture sequence. Simulated and recorded spectral amplitude is reasonably similar in all four gauges.



Figure 5. Comparison of measured and computed runup for the whole of Samoa (blue
symbols) and southeast Upolu (red symbols) for the two scenario *Hossen et al.* [2017]
(triangles) and *Lay et al.* [2010] version C (circles).



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Figure 6. Comparison of computed with observed inundation on southeast Upolu using: (a) 669 Lay et al. [2010] version C, and (b) Hossen et al. [2017]. 670



Figure 7. (a) Timeseries of water level for scenario *Lay et al.* [2010] version C showing the

variability of onshore hydrodynamics response at Lepa and Saleapaga, and offshore of Lepa.(b) Amplitude spectrum showing local variability of onshore and offshore wave energy and

674 (b) Amplitude spectrum675 excitation frequencies.



Figure 8. Sequence of inundation due to the first tsunami wave for scenario *Lay et al.* [2010] version C at Lepa and Saleapaga. The leading trough is represented by the initial draw-down of the first waves drying out the lagoon and reef foreshore, with the crest rising offshore. The model was calculated in UTM.



Figure 9. (a) Spatial distribution of Fault 1 (normal fault) contribution to maximum runup. (b) Spatial distribution of Faults 2 and 3 (reverse faults) contribution to maximum runup for scenario *Lay et al.* [2010] version C. Areas of orange/red are where the normal fault alone contributes to most/all of the inundation. Areas of blue/green are where thrust faults contributed 50/30 % of the runup. The model was calculated in UTM.

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Supporting Information for

Effects of source faulting and fringing reefs on the 2009 South Pacific tsunami inundation in southeast Upolu, Samoa

Cyprien Bosserelle¹, Shaun Williams¹, Kwok Fai Cheung², Thorne Lay³, Yoshiki Yamazaki², Titimanu Simi⁴⁺, Volker Roeber⁵, Emily Lane¹, Ryan Paulik¹, Lameko Simanu⁴

¹National Institute of Water and Atmospheric Research (NIWA), New Zealand. ²Department of Ocean and Resources Engineering, University of Hawaii at Manoa, Hawaii, USA. ³University of California Santa Cruz, California, USA. ⁴Disaster Management Office, Ministry of Natural Resources and Environment, Apia, Samoa. ⁵Université de Pau et des Pays de l'Adour (UPPA), HPC-Waves Laboratoire SIAME, Pau, France. ⁺Current affiliation: Management Unit, Green Climate Fund, Samoa Ministry of Finance.

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Introduction

The supporting information includes Figures S1-S14 showing regional maximum water level, recorded and computed time series and spectra of surface elevations at DART water-level stations and the Apia tide gauge, measured and computed runup for the whole of Samoa and southeast Upolu, and computed vs observed inundation on southeast Upolu for each of the fourteen scenarios tested. Movie S1 provides an animation of the tsunami generation, propagation and inundation for our selected sources discussed in the main text.

Figure S1-S14. Regional maximum water level (top left); recorded (black) and computed (red) time series and spectra of surface elevations at DART water-level stations and the Apia tide gauge (right); measured and computed runup for the whole of Samoa (blue dots) and southeast Upolu (red dots) (middle left); computed vs observed inundation on southeast Upolu (bottom), for each of the 14 scenarios tested. These are consistent with the sources described in Table 1 of the main text.





























Movie S15. Animation of the 2009 tsunami from initiation to inundation in Samoa; (available for download at https://doi.org/10.5281/zenodo.4054271).