Changes in ground deformation prior to and following a large urban landslide in La Paz, Bolivia revealed by advanced InSAR

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Abstract. We characterize and compare creep preceding and following the 2011 Pampahasi landslide (~40 Mm^3 ± 50%) in the city of La Paz, Bolivia, using spaceborne RADAR interferometry (InSAR) that combines displacement records from both distributed and point scatterers. The failure remobilised deposits of an ancient landslide in weakly cemented, predominantly fine-grained sediments and affected ~1.5 km^2 of suburban development. During the 30 months preceding failure, about half of the toe area was creeping at 3-8 cm/a and localized parts of the scarp area showed displacements of up to 14 cm/a. Changes in deformation in the 10 months following the landslide are contrary to the common assumption that stress released during a discrete failure increases stability. During that period, most of the landslide toe and areas near the headscarp accelerated, respectively, to 4-14 and 14 cm/a. The extent of deformation increased to cover most, or probably all, of the 2011 landslide as well as adjacent parts of the slope and plateau above. The InSAR-measured displacement patterns – supplemented by field observations and by optical satellite images – indicate that kinematically complex, steady-state creep along pre-existing sliding surfaces temporarily accelerated in response to heavy rainfall, after which the slope quickly achieved a slightly faster and expanded steadily creeping state. This case study demonstrates that high-quality ground-surface motion fields derived using spaceborne InSAR can help to characterize creep mechanisms, quantify spatial and temporal patterns of slope activity, and identify isolated small-scale instabilities. Characterizing slope instability before, during, and after the 2011 Pampahasi landslide is particularly important for understanding landslide hazard in La Paz, half of which is underlain by similar, large paleolandslides.

1 Introduction

Creep – steady, imperceptibly slow movement under the influence of gravity – precedes many large landslides in both bedrock (e.g. Chigira and Kiho, 1994; Kilburn and Petley, 2003) and soil (e.g. Kalaugher et al., 2000; Petley et al., 2002). Its temporal patterns can improve understanding of slope-deformation processes (Petley et al., 2002) and the possible timing of
5   impending catastrophic failure (Voight, 1989; Crosta and Agliardi, 2003; Federico et al., 2015). Its spatial patterns indicate the style (Gischig et al., 2011; Schlögel et al., 2016) and approximate magnitude of instability. Consequently, characterization of temporal and spatial patterns of slow deformation is a focus of many landslide hazard assessments, monitoring systems, and warning programs (Eberhardt, 2012; Baron and Supper, 2013; Hermanns et al., 2013a). In many cases, however, pre-failure creep is not quantified, either because it is not recognized prior to catastrophic failure or because
10 failure carries a relatively low risk to people and property.

Creep is rarely monitored following a landslide, except at sites where renewed catastrophic failure is expected (e.g. Gischig et al., 2011). Limited post-event monitoring stems in part from the common assumption that stress release during catastrophic failure enhances slope stability. In addition, instrumentation that may have been present on or within the slope is typically destroyed during failure (e.g. Crosta and Agliardi, 2003). Consequently, activity soon after failure is generally
15 unclear and can rarely be compared with pre-failure deformation.

Spaceborne RADAR interferometry (InSAR) has long been recognized for its promise in characterizing the spatiotemporal evolution of slopes (e.g. Petley et al., 2002; Colesanti and Wasowski, 2006). Continued technological advancements and data availability offer the potential for further improvements to the characterization of slope behaviour, including prediction of future failure (Moretto et al., 2017). Improvement of RADAR systems increases ground resolution and phase quality,
20 whereas the establishment of satellite constellations is improving temporal resolution and coherence. Due to such improvements, pre-failure accelerated creep has been measured only recently, both for natural (Intrieri et al., 2018) and cut (Carlà et al., 2018) slopes. Such advancements similarly improve opperunities to analyze post-failure creep.

Our quantification of creep during the 30 months preceding and 10 months following the largest modern landslide in the city of La Paz, Bolivia (2011 Pampahasi landslide, ~40 Mm^3) using InSAR demonstrates unexpected slope behaviour. Counter to
25 the conventional view that a major failure event reduces large-scale instability, the affected slope shows enhanced post-failure activity, which suggests a need to revaluate some typical assumptions about slope stability and risk. The InSAR technique we employ is optimized to maximize displacement record density and distinguish abrupt changes in motion, which together improve the characterization of spatially complex deformation patterns typical of landslides. The resulting displacement fields provide nearly spatially continuous quantification of motion over much of the landslide area and
30 surrounding terrain that enable discrimination of localized motion and aid interpretation of failure mechanisms..

2 Setting and material properties

2.1 La Paz region

A tributary of the Amazon River system penetrates the Cordillera Real ~60 km southeast of La Paz and drains the eastern margin of the otherwise internally drained Altiplano Plateau (~4000 m a.s.l.) (Fig. 1A). The city of La Paz (population ~0.8

5 million) is located within the upper part of this watershed where the valley system has incised into up to 800 m of weakly lithified Neogene and Pleistocene sediments shed from the Cordillera Real (Dobrovolny, 1962) (Fig. 1B). South of the city, incision has exposed folded and faulted, Paleozoic to Mesozoic metasedimentary basement rocks forming the root of the range (Fig. 1B). The overlying Cenozoic fill sequence comprises over 600 m of predominantly fine-grained lacustrine and fluvial sediments of the weakly lithified Mio-Pliocene La Paz Formation. It coarsens upward and toward the cordillera into

10 more competent Late Pliocene though Early Pleistocene glacial and proglacial sediments (Roberts et al., 2018) that are up to 400 m thick. The fills have been offset by numerous faults trending northwest and west (Dobrovlny 1962; Lavenu, 1978; Roberts et al., 2018) (Fig. 1B).

The slopes and floors of the La Paz valley system are mantled by colluvium, including deposits of large (up to 50 km^2) paleolandslides derived from the La Paz Formation (Dobrovolny, 1962; Anzoleaga et al., 1977) (Fig. 1B). Most of the

15 landslides date to the Late Pleistocene and Holocene (Dobrovolny, 1962; Hermanns et al., 2012). Historic instability is concentrated within paleolandslide deposits (e.g. Fig. S1), particularly at their margins, likely reflecting low residual strengths and possibly differential stresses related to suspected ongoing creep. Recent landslides, particularly those <1 Mm^3, have happened mainly during the monsoon season (December-March), and small failures occur most years (O'Hare and Rivas, 2005). Only four of the seven historic landslides larger than 1 Mm^3 have happened during the dry season (April-

20 November; Table S1).

2.2 Pampahasi area

Pampahasi Plateau, measuring 2.2 km long by 0.2-0.7 km wide, forms part of the interfluve between ríos Chuquiaguillo and Irpavi (Fig. 1B). Its margins are generally scalloped (Fig. 2) and in several places terminate in nearly vertical cliffs dropping 20-80 m to irregular, locally steep valley slopes below. The plateau surface dips slightly southward and lies at an elevation

25 several hundred metres below the local southwest-dipping Altiplano and its erosional remnants (Fig. 3A).

The Río Irpavi valley is cut into well stratified fine-grained facies of the La Paz Formation, which comprises horizontal to sub-horizontal silt and sand beds interlensing with some mainly granitic pebble-cobble gravel (Fig. 3B). The variable texture and limited (<1 km) lateral continuity of these beds are typical of the unit, particularly its middle part (cf. Ahlfeld, 1945a; Dobrovolny, 1962; Bles et al., 1977). The age of the La Paz Formation beneath Pampahasi Plateau (Fig. 3B) is constrained

30 by the 5.4-Ma Cotacota Tuff (Lavenu et al., 1989; Servant et al., 1989), which outcrops at ~3520 m a.s.l. in the southern part

of the city (Lavenu et al., 1989), and the 2.74-Ma Chijini Tuff (Roberts et al., 2017), which crops out along the west margin of the plateau (Dobrovolny, 1962) between 3800 and 3850 m a.s.l. The local Altiplano surface formed atop the thick sequence of Plio-Pleistocene glacial sediments by 1.8 Ma or possibly as recently as 1.0 Ma (Roberts et al., 2018), attaining an elevation of approximately 4200 m a.s.l. in the vicinity of Pampahasi (Fig. 3A).

5   About 400 m of incision of the Altiplano surface by ancestral Río Irpavi after 1.8-1.0 Ma formed a south-sloping (Dobrovolny, 1962), valley-bottom unconformity onto which the Middle Pleistocene Pampahasi gravel (Fig. 3B) was deposited, either as fluvial channel deposits (Ahlfeld, 1945b) or as an ancient landslide (Bles et al., 1977). This gravel multilithic silty sandy pebble-cobble gravel is commonly more than 20 m thick (Dobrovolny, 1962), and may be up to 50 m thick along the plateau's southern edge where the unit overlies a locally steeply erosional contact (Dobrovolny, 1955).

10 Many slopes descending from Pampahasi Plateau are mantled by ancient landslide deposits derived from the La Paz Formation (Dobrovolny, 1962; Anzoleaga et al., 1977). The deposits consist of largely intact brittle blocks of La Paz Formation within a matrix of softer, ductile and completely disaggregated of La Paz Formation. Historic, small (<1 to 20 ha) failures have occurred within large (~20-200 ha) prehistoric failures in many places. The slope descending ~380 m to Río Irpavi directly east of the southern half of Pampahasi Plateau comprises the undated Pampahasi paleolandslide (Fig. 2A).

15 Details of the Pampahasi paleolandslide must be inferred from its surface expression as no subsurface investigations have been conducted on this slope. Its morphology indicates complex failure, but whether it represents one or multiple events is uncertain. Rotational displacement of the upper quarter of the slope transitions downward to eastward sliding and localized flow. The landslide displaced Río Ipravi eastward from its alignment farther upstream and downstream by at least 300 m. The substantial height of the headscarp (up to 80 m) suggests that the failure zone is many tens of metres below surface in 20 the upper part of the landslide. The failure depth lower on the landslide is likely much shallower.

Surface incision of the Pampahasi paleolandslide by several gullies suggests that it is of substantial antiquity. In the area of the Pampahasi paleolandslide, the Río Irpavi channel is modified with check dams and concrete armoring. Before the 2011 failure, western tributaries of Río Irpavi draining the Pamaphasi and Villa Salomé paleolandslides were little modified and had incised into weak, loose material. The main channels draining the area (ríos Chujilluncani and Jankopampa: Fig. 2) have 25 since been entombed in concrete culverts.

2.3 Geomechanical properties

The different characteristics of the Pampahasi gravel and La Paz Formation give rise to contrasting material properties, which have been documented by Anzoleaga et al. (1977; summarized in Table S2). The Pampahasi gravel (50-75% clasts,

19-50% sand and silt, <6% clay) is permeable (k ~1 * 10-3 m/s), has low to medium plasticity (plasticity index ≤10), and exhibits high internal friction (>30°), particularly when dry. Fine-grained zones are weaker, but uncommon.

The properties of the La Paz Formation are variable due to the unit's heterogeneity. Coarse sand and gravel zones (35-75% clasts, 13-65% sand and silt, <12% clay) are of similar permeability to the Pampahasi gravel, but are commonly cemented by

5   carbonate (≤15%) or iron oxide. Fine-grained zones (≤10% clasts, 40-90% sand and silt, 10-40% clay) are much weaker and, as the main facies, dominate the unit's behaviour. They are weakly cemented by carbonate (<5%) and locally abundant clay. The clay is largely non-expansive, but montmorillonite (smectite) is common in some beds and lenses. Due to low permeability (k <1 * 10-5 m/s), water circulation is mainly through fractures in interstratified gravel lenses or along joints. Plasticity is medium to high (plasticity index of 7-32), and shear strength ranges from 18° to 37°. The unit is generally strong

10   when dry and undisturbed (cohesion 0.2-0.7 MPa) and stands in near-vertical slopes. However, wet silty sediments have much lower cohesion (0.01-0.1 MPa) and shear strength. Smectite-rich clay lenses have greater cohesion (≥0.1 MPa), but high plasticity (plasticity index of 60-80) and low shear strength (≤14°). Parts of the La Paz Formation that have failed in the past are even weaker due to loss of compaction and localized reduction of shear strength to residual levels (as low as 13° in silty/sandy zones and less 8° or less in clay zones).

15   ## 3 Methods

3.1 Failure event characterization

Our characterization of the 2011 landslide is based on field visits in the years before and after the event and on comparison of high-resolution optical satellite images acquired shortly before and after the event. Field observations by municipal staff and eyewitness accounts they gathered from residents provided details on the location, type, and magnitude of damage

20   shortly before, during, and after the 2011 failure. We mapped features of the 2011 landslide from the first cloud-free, post-failure imagery (WorldView-2 acquired on 23 March 2011), and quantified horizontal displacement vectors during the failure by comparing it with a pre-failure (05 January 2011) image from the same sensor that was coregistered using points outside the landslide-affected area. Due to errors resulting from the lack of orthorectification and from elevation differences cause by the landslide, the estimated horizontal vectors are approximate (rounded to the nearest 5-m increment).

25   ### 3.2 HDS-InSAR

We applied Homogeneous Distributed Scatterer InSAR (HDS-InSAR; Eppler and Rabus, 2011; Rabus et al., 2012) to a stack of 44 Fine Beam mode (5.1 m azimuth x 8.0 m ground-range pixel spacing) RADARSAT-2 ascending scenes (look direction: 36.3° incidence angle from vertical at scene center, 76.0° aspect angle counter clockwise from north) acquired between September 2008 and December 2011 (Supplementary Material, Tables S4 and S5). We processed the scenes

separately as a pre-failure stack (32 scenes) and a post-failure stack (12 scenes) to compare deformation patterns before and after the landslide and to reduce decorrelation caused by landslide-induced terrain changes. Although a range of displacement directions is possible, ground-surface displacement vectors due to mass movements are most probable in the downslope direction; we thus estimated true displacement rates from the angular deviation between down-slope vectors and

5 the satellite's line-of-sight (LOS).

The HDS approach combines the strengths of Persistent Scatterer InSAR (PS-InSAR; Ferretti et al., 2001) and Small Baseline Subset InSAR (SBAS-InSAR; Berardino et al., 2002) for considering, respectively, point and distributed targets. A continuously weighted, spatially adaptive filter defines spectrally similar pixel clusters (HDS neighbourhoods) within a rectangular search area based on similarity of their backscatter amplitude time series. The filter is applied to interferograms

10 to generate differential phase and coherence for each HDS neighbourhood (Eppler and Rabus, 2011), which comprises either a single-pixel cluster (a persistent point target) or the weighted mean of a multi-pixel cluster (interpreted as the area covered by a homogenous distributed target). To more efficiently represent the high density of displacement records, we interpolated linear deformation maps from HDS to provide base images of long-term, average line-of-sight deformation over the entire area.

15 Like PS-InSAR, HDS-InSAR generates high-quality displacement time histories attributable to single pixels, each containing a dominating point target with near-linear displacement characteristics (Colesanti and Wasowski, 2006). Additionally, it provides time histories for multi-pixel clusters representing homogeneous distributed scatterers. Like SBAS-InSAR, HDS-InSAR characterizes strongly non-linear motion (Necsoiu et al., 2014) using the higher continuity coverage (Lauknes et al., 2010) of distributed targets. Unlike SBAS-InSAR, however, it does not significantly diminish phase quality

20 or reduce spatial resolution. Two key differences set HDS-InSAR and some other advanced InSAR algorithms (e.g. Ferretti et al., 2011) apart from the PS-InSAR and SBAS-InSAR techniques: the use of adaptive filtering to preserve, as much as possible, spatial resolution while suppressing the noise from surface decorrelation of nearby incoherent pixels; and optimization to characterize spatially uncorrelated ground motion. These features make HDS-InSAR particularly well suited for characterizing landslides, which commonly display ground motion variability over short distances or across abrupt

25 transitions. Although HDS-InSAR and SqueezSAR (Ferretti et al., 2011) share similar adaptive filtering methods, the latter uses wrapped phase triangulation to invert the interferometric network. In contrast, HDS-InSAR relies on prior unwrapping of the network interferograms, with potential unwrapping errors being corrected iteratively using the network redundancy a-posteriori. Despite the differences in their algorithms, the final accuracy and spatial detail of both advanced InSAR methods are similar.

4 2011 Pampahasi landslide

The Pampahasi landslide[1] occurred at the end of February 2011 following a week of variable precipitation (0-39.2 mm daily: Fig. S2), which included one of the wettest days on record (25 February; Table S3). The ~1.5-km^2 failure remobilized a large part of the Pampahasi paleolandslide deposit between the east side of Pampahasi Plateau and Río Irpavi (Fig. 2). It included small volumes of in situ, weakly cemented La Paz Formation and, in the headscarp area, the overlying uncemented Pampahasi gravel (Fig. 2A). None of the sediments involved were substantially weathered.

Several locations in the upper part of the slope showed field evidence of slow ground deformation, largely visible as offset walls and road surfaces (Fig. 4; locations in Fig. 2A) in the years prior to catastrophic failure (Quenta and Calle, 2005; Hermanns et al., 2012). These features were recognized as evidence of an impending landslide (Quenta and Calle, 2005), although not of the magnitude of the 2011 failure. In response, the Municipality of La Paz increased risk communication and conducted localized remedial engineering works (Hermanns et al., 2012). However, due to their localized and shallow nature, the stabilization efforts probably had little, if any influence on the stability of the slope or behaviour of the subsequent failure.

The 2011 Pampahasi landslide occurred in several phases from the evening of 26 February to 1 March (Fig. 5; locations in Fig. 2B). Eyewitness accounts suggest that failure initiated in an area of earlier small slumps along Río Chujilluncani (Hermanns et al., 2012; Fig. 4A) ~400 m downslope of the eastern margin of Pampahasi Plateau. Motion was fastest (up to several metres per second) and largely vertical in the head region, lasting one to two hours and forming a ~60°, 80-m-high headscarp (Fig. 5A). Movement in the central zone and at the toe, which began soon after the initial formation of the headscarp, was slower (metres per minute to metres per day) and largely horizontal. The degree of ground disturbance and the magnitude of displacement generally decreased to the east and north (Fig. 2B). The northern part of the landslide's toe experienced relatively limited, but economically costly, ground disturbance (e.g. Fig. 5B-D) and moved as a more-or-less coherent mass ~15 m southeastward (Fig. 2B). Deformation of the lower part of the landslide continued for days. A bridge across Río Irpavi (Fig. 5E) was shifted 10 m and uplifted on the morning of 27 February; uplift of several metres at the east end of the bridge suggests a rotational failure zone passing, at least locally, under Río Irpavi. Buildings near the south toe of the landslide (Fig. 5F) collapsed on 28 February and 1 March due to delayed, relatively minor translation.

The surface morphology of the landslide likely reflects a complex failure zone with both rotational and translational components. Assuming a minimum average thickness of the landslide deposit of 15 m, its volume is at least 20 Mm^3.

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[1] The 2011 failure is also locally called the Callapa landslide after one of the neighbourhoods it destroyed.

However, the volume is more likely about 40 Mm^3, given deeper failure in both the upper rotational part (40 m average depth over 32 ha) and the main predominantly translational body (25 m average depth over 110 ha).

Due to the its moderate to low velocities and the immediate evacuation of the area, the landslide did not claim any lives. However, about 1000 homes were destroyed and 6000 people were displaced. Rupture of a water line crossing the upper part of the landslide (Fig. 5G) left between least 200,000 and 300,000 people in southern La Paz without potable water for several months (Hermanns et al., 2012; Aguilar, 2013). Continued episodic collapse of the steep headscarp in the years since the landslide has resulted in additional loss of homes and repeated expansion of an evacuation zone on Pampahasi Plateau.

5 Deformation before and after failure

Due to the high spatial density of displacement records, particularly in high-coherence areas (Fig. 6B), interpolated linear deformation maps are preferable for interpreting the spatial variably of slope creep (cf. Fig. 6C). InSAR-measured ground deformation in the Pampahasi area is almost entirely restricted to mapped paleolandslide deposits and the terrain immediately behind their headscarps (Fig. 7). Their characteristic displacement histories are shown in Fig. 8 (locations 'i'-'x'). Data gaps in each stack are locations of low coherence that mainly result from decorrelation related to remedial earthworks performed after a 10-ha landslide in 2009 (Fig. 7A, B) and the 2011 landslide (Fig. 7C, D), or from aliasing where displacement rates are greater than the detection threshold of RADARSAT-2 (~2.6 cm LOS over 24 days). Motion toward the satellite suggested by the data in the east-facing slopes (Fig. 7) is improbable, given that it implies uplift or motion upslope. The issues are almost entirely restricted to the post-failure records (Fig. 7C), suggesting that the small size and limited redundancy of the interferometric network in the thinner stack has resulted in incomplete network correction of either phase unwrapping errors or atmospheric effects.

5.1 Pre-failure creep

Prior to the end of February 2011, about 180 ha of the lower part of the slope within the limits of the 2011 landslide was moving at rates as high as 14 cm/a (9 cm/a LOS; Fig. 7A, B). Motion was most widespread over the northern part of the toe of the Pampahasi paleolandslide (2-5 cm/a LOS; 'i'), where it extend into Río Jankopampa alluvium mantling the paleolandslide deposit. There it terminated abruptly ~50 m north of the river (Fig. 7A) at a locally stable slope comprising undisturbed La Paz Formation (Fig. 2). A 1-ha zone directly west of Río Irpavi halfway along the toe of the paleolandslide ('ii') crept even faster (5-8 cm/a LOS: Fig. 7B). In contrast, activity on the southern part of the paleolandslide toe ('iii') was limited to a few small areas creeping at less than 0.5 cm/a LOS.

Pre-failure movement in the upper half of the Pampahasi paleolandslide was localized (Fig. 7A, B). Minor isolated movement (<1 cm/a LOS) is apparent along what became the 2011 landslide's upper lateral margins (e.g. 'iv'). Creep within

the Pampahasi paleolandslide deposit beyond the limits of the 2011 failure also was localized and generally slower than 1 cm/a LOS (Fig. 7A). The most extensive such area involved 12 ha of paleolandslide material between the 2009 and 2011 landslides ('v') directly upslope (northwest) of the most active part of the slope toe, and crept at 0.5 cm/a LOS. The most rapid pre-failure motion (9 cm/a LOS) was restricted to the upper portion of the future landslide where infrastructure damage had been documented in prior years (Fig. 4B, C). Aside from a small, slow-moving (0.5 cm/a LOS; 'vi') area, no motion is evident along the future landslide headscarp, including sites of localized deformation observed in 2005 (Fig. 4D, E) or on the plateau behind it ('vii'-'x').

5.2 Post-failure creep

During the 10 months following the landslide, the area of movement on Pampahasi Plateau and the valley slope between it and Río Irpavi increased by about two-thirds to nearly 300 ha and included the entire area of the 2011 landslide where coherence was maintained (Fig. 7C). The maximum inferred downslope displacement rates following the landslide (14 cm/a; 9 cm/a LOS; Fig. 7D) were similar to those before it, but occurred in a region of the toe that was previously stable ('iii') and along the new headscarp ('vi') where creep had previously been slow (0.5 cm/a LOS). Displacement rates also increased in the northern part of the toe ('ii', from 2-5 to 3-6 cm/a LOS) and were more spatially variable than in the pre-failure period (Fig. 7).

Detection of movement is not possible over much of the middle and upper parts of the 2011 landslide due to decorrelation resulting from earthworks that continued for years after the event. However, because this zone is bordered on all sides by moving terrain (>1 cm/a LOS; Fig. 7C), it also was likely creeping throughout the period of the post-failure stack. The rate and extent of creep increased in a zone measuring nearly 500 by 500 m (including 'v') in the paleolandslide deposit between the 2009 and 2011 landslides.

The most striking change in ground motion following the 2011 failure is the development of a new area of creep several hundred metres wide fringing the head region of the landslide (Fig. 7C). This zone extends up to nearly 400 m beyond the headscarp on Pampahasi Plateau, in one place reaching the opposite side of the plateau surface. The magnitude of creep decreases steadily away from the headscarp ('vi'-'x' in Figs. 7C, D, and 8B). This zone of new deformation extends hundreds of metres down both lateral margins.

All portions within the 2011 landslide area where coherence was maintained in both InSAR stacks and most areas within several hundred metres of the landslide experienced post-failure changes in slow ground motion (Fig. 9). The changes largely involved increased motion away from the satellite of between 2 cm/a and at least 5 cm/a LOS. Only a few small areas (totaling < 4 ha) in the northeast part of the toe, which were creeping at 4 cm/a LOS or more prior to February 2011, slowed following failure.

6 Conceptual failure model

Combining details of the 2011 failure event with patterns of long-term pre-failure and post-failure creep, we identify several types of instability that inform the geomechanical evolution of the Pampahasi slope.

6.12 Components of failure

5   The failing slope comprises multiple interconnected components (Fig. 10A). Slope morphology before and after the 2011 landslide indicates that long-term failure is predominantly by complex rotational-translational sliding (Fig. 10B). Differing displacement behaviour of several zones within the slope indicate separate, although are likely interacting, kinematic components. The northern part of the Pampahasi slope toe consists of an apparently competent mass that creeped uniformly at rates of 2-5 cm/a LOS prior to late February 2011 (Fig. 7B) and moved more-or-less coherently during the 2011 landslide

10   (Fig. 2B). This block-like mass accelerated slightly to 3-6 cm/a LOS following the 2011 failure and displayed a more spatially variable pattern of deformation (Fig. 7D) suggesting partial break-up during the failure event. The resulting removal of toe support – rapidly during the 2011 landslide, but otherwise gradually – likely drove creep of the 12-ha zone of paleolandslide material directly upslope (to the northwest). Although not part of a historic landslide event (Fig. 2), this zone was creeping slowly (~0.5 cm/a LOS; Fig. 7A) prior to the 2011 failure and accelerated thereafter (>1 cm/a LOS; Fig. 7C).

15   A 1-ha zone near the midline of the landslide toe ('ii') is one of the few locations of decreased post-failure activity (Fig. 9). Its high, localized activity prior to the 2011 landslide (5-8 cm/a LOS; Fig. 7C) suggests a shallow failure driven by incision at a location where Río Irpavi impinges on the toe of the slope (Fig. 2B). This localized activity is probably similar to the east-bank landslide mapped by Anzoleaga et al. (1977) ~ 400 m farther upstream, opposite the Río Jankopampa confluence (Fig. 7A), which was creeping throughout the entire period of RADAR monitoring (Fig. 7). Decreased post-failure activity

20   of the small west-bank instability may reflect local modification of the Río Irpavi channel by the 2011 landslide and subsequent channel remediation efforts, or it may result from post-failure stress reorganization. Several similar, linear areas of decreased post-failure creep along the toe of the 2011 landslide as far north as Río Jankopampa (Fig. 9) may represent similar conditions.

The southern part of the toe of the paleolandslide, which was stable prior to the 2011 Pampahasi landslide (Fig. 7A),

25   reactivated following the failure (Fig. 7C). Although the exact cause of the reactivation is uncertain, it is clearly a result of forces imposed on the toe of the paleolandslide by the 2011 event. Creep in this area was greatest immediately upslope of where several buildings collapsed in the days following the main phase of the 2011 landslide, suggesting that delayed infrastructure damage there relates to the transition from the failure event to the new post-failure instability regime.

The height and steepness of the 2011 landslide headscarp indicates a deep-seated basal failure zone in the upper slope (Fig. 10B). Neither the paleolandslide nor the 2011 failure align with documented faults (Dobrovlny 1962; Lavenu, 1978; Fig. 1B) that might indicate a structural control. A lack of displacement records for the head region following the 2011 landslide limit inferences about post-failure behaviour. However pre-failure records, although incomplete, provide some insight into localized activity. Displacements within the head region prior the landslide and the lack of movement in adjacent steep paleolandslide scarps suggest that pre-failure creep was part of the ongoing slope failure, rather than localized instability resulting only from steep slopes. Localized, small-scale slumping occurred lower in the head region along Río Chujilluncani ('a' in Fig. 7A, B) adjacent to, or potentially within, a region of pre-failure creep (~1 cm/a LOS), supporting eyewitness accounts suggesting that the 2011 landslide initiated there.

The new displacements behind the headscarp and along the upper lateral margins of the 2011 landslide (Fig. 7B) gradually expanded over time into, respectively, the Pampahasi gravel (Fig. 8B) and paleolandslide deposits ('iv' and 'v' in Fig. 8A). Movement in the latter suggests post-failure reactivation of landslide material as a result of the 2011 landslide. Movements on the plateau are likely the result of dilation of the Pampahasi gravel in response to removal of material to a depth of up to 80 m along the 2011 headscarp (Fig. 5A).

6.2 Mechanisms of failure

Progressive failure is increasing being recognized as a precursor of large, rapid landslides (e.g. Petley et al., 2002, 2005; Kilburn and Petley, 2003; Intrieri et al., 2018). In first-time brittle failures, creep reflects damage accumulation during the reduction of intact material strength, discontinuity strength, or both (Brideau and Roberts, 2015; and references therein) and consequent stress concentration (Cornelius and Scott, 1993). In such slopes, acceleration increases over time (Petley et al., 2002; Kilburn and Petley, 2003). Eventual development of a continuous failure surface (Petley et al., 2005) enables catastrophic release, commonly resulting in debris fragmentation, high runout, and velocities exceeding 5 m/s (Hermanns and Longva, 2012). Sudden exceedance of resisting forces – for instance by seismic loading, increased pore pressure, or slope undercutting – may prematurely complete the failure surface, rapidly terminating the displacement trend (Hermanns and Longva, 2012). In slopes failing along a ductile shear zone or an existing plane of weakness, creep acceleration decreases overtime, trending toward zero (Petley et al., 2002). Because displacement typically accelerates in each of the aforementioned failure types (Fig. 10C inset), it provides an opportunity for failure forecasting (Voight, 1989; Crosta and Agliardi, 2003; Eberhardt, 2012; Baron and Supper, 2013; Hermanns et al., 2013a; Federico et al., 2015; Moretto et al., 2017).

In contrast, over the 40-month period we monitored the Pampahasi slope (Fig. 10C), long-term steady-state creep was punctuated by temporary acceleration. Pre-2011 displacement rates are for the most part extremely slow (after Cruden and

Varnes, 1996), but constant. No progressive acceleration was evident at the temporal resolution of RADARSAT-2 or from eyewitness accounts during the days leading up to failure. Displacement rates during the four-day landslide event ranged from moderate to very rapid, but did not reach the extremely rapid range (>5 m/s: Cruden and Varnes, 1996) typical of catastrophic failures. Creep rates following the 2011 landslide were also constant, but were greater than those prior to failure,

5   indicating a new state of instability following the landslide. Constant creep rates suggest that the slope had, prior to failure and quickly afterward, reached steady-state behaviour, consistent with the presence of an existing failure surface (cf. Petley et al., 2002). Consequentially, prediction of large-scale failure events on the Pampahasi slope or at similar sites in La Paz should not rely exclusively on long-term creep time series.

6.3 Controlling factors

10   Geomechanical behaviour of the Pampahasi slope indicates failure nearer to that of sediment than to rock. Both the La Paz Formation and the Pampahasi gravel have compressive strengths well below 1 MPa, the threshold typically used to differentiate between rock and sediment (Brideau and Roberts, 2015). Parts of the main failure zone, as well as various minor rupture zones within the landslide body, likely follow weak, fine-grained seams sheared during the Pampahasi paleolandslide or other past failures. Residual shear strength will be especially low in clay-rich zones. The lower half of the

15   valley slope and the inferred failure zone beneath it dip at a similar angle (~8°: Fig. 10B) to the residual shear strength of the weakest (clay) zones within the failed sequence (Table S2), further suggesting that failure occurred along an existing failure surface.

Geotechnical properties of the slope, particularly its reduced strength resulting from extensive previous failure, enable its continual activity. However, external factors contribute to the long-term instability of this slope and many others in the La

20   Paz basin. On-going fluvial down-cutting and toe erosion by Río Irpavi helps to maintain the slope's meta-stable condition. This driver is a response to the 4-km drop in base level since the Early Pleistocene – from the time when the Altiplano plateau was internally draining to the present drainage to the Atlantic Ocean (Roberts et al., 2018) – and operates along other trunk streams in be basin.

Some, but not all, of the largest historic landslide events in the La Paz area follow intense precipitation events

25   (Supplementary Material), suggesting that enhanced pore-water pressure plays a role in initiating large discrete failures. The sudden onset of the 2011 Pampahasi landslide (Fig. 10C) is typical of triggered landslides (Fig. 10C inset; Hermanns and Longva, 2012). Heavy precipitation on the previous day – the tenth wettest on record – is the only apparent trigger for the failure. Slope wetting was likely enhanced by effluent from homes and concentrated overland flow from impervious surfaces. We thus postulate that rainwater infiltrating the slope and plateau above it on 25 February reached the basal failure

30   zone by the following evening; consequent rapid pore-pressure increase and shear-strength reduction appear to have

triggered the landslide. Rupture of the water line crossing the slope added much more water and likely contributed to mobility of the 2011 landslide directly downslope\ where greatest displacement rates were greatest.

6.4 Sources of uncertainty

Possible acceleration in the days or weeks prior to the 2011 landslide cannot be evaluated with the temporal resolution of the displacement histories used here. RADARSAT-2's 24-day revisit period provides low sampling frequency compared to some other RADAR satellites (Moretto et al., 2017) and especially to in-situ sensors or ground-based remote sensors (e.g. Kalaugher et al., 2000; Crosta and Agliardi, 2003; Gischig et al., 2011; Eberhardt, 2012; Federico et al., 2015; Confuorto et al., 2017; Carlà et al., 2018). Additionally, the temporal filter used in processing (Supplementary Material) attenuates records of non-linear displacement trends. The resulting temporal details reported here are sufficient to characterize gradual acceleration over periods of months (Fig. 8) as has been documented for pre-failure creep using InSAR elsewhere (Carlà et al., 2018; Intrieri et al., 2018) or more abrupt acceleration over just a few RADAR acquisitions, but not shorter term changes. Sporadic displacement activity that may signal impending acceleration (cf. Kalaugher et al., 2000) could similarly have gone undocumented.

The assumption of slope-parallel motion is likely incorrect in some parts of the Pampahasi area. The complex nature of the 2011 failure, particularly the variability in the displacement directions (Fig. 2B), suggests that separate masses may have moved obliquely relative to one another during pre-failure and post-failure creep. Localized compression or extension within the landslide would additionally impart vertical components of movement. Provided that the sense of movement at any given location on the slope has not substantially changed following the 2011 failure, line-of-sight motion vectors record similarly scaled changes in the true rate of motion and thus the degree of slope activity before and after that event. Additionally, the close alignment between the RADAR LOS and the fall-line of the slope ensures that the InSAR-measured displacement record is particularly sensitive to expected gravity-driven slope motion between Pampahasi Plateau and Río Irpavi.

Because RADARSAT-2 acquisitions over the site ended 10 months after the failure, the current activity of the slope is poorly constrained and it is possible that increased post-failure slope creep was short-lived. Additional InSAR acquisitions planned for this and other sites in La Paz, including both ascending and descending orbits, will provide additional insight on the temporal and special complexity slope activity.

7 Advances in InSAR monitoring of landslides

7.1 Comparison with other studies

Our detailed documentation of slope activity preceding and following a large landslide using spaceborne InSAR is, to our knowledge, unique. Numerous studies show evidence of long-term, generally sustained creep of slow-moving landslides
5 (e.g. Colesanti and Wasowski, 2006; Lauknes et al., 2010; Hermanns et al., 2013b; Necsoiu et al., 2014; Schlögel et al., 2016), and in many cases document temporal variability in displacements. A much smaller number of studies identify short-term deformation between successive scenes (Moro et al., 2011) or creep leading up to catastrophic failure (Carlà et al., 2018; Intrieri et al., 2018). However, we are aware of only one other spaceborne InSAR investigation that compares pre-failure and post-failure activity. Confuorto et al. (2017) report displacement rates during a one-year period starting ~20
10 months following the 2012 Via Piave landslide in southern Italy that were marginally faster than rates during a ca. two-year period ending ~16 months before failure (See Supplementary Material for comparison with the dataset presented here). In that investigion, the limited number of point reflectors, their concentration in a narrow band of the headscarp, and the three-year data gap bracketing the landslide limit insight into potential post-failure changes of the slope.

7.2 Expanding utility

15 Advanced InSAR approaches such as HDS-InSAR increase ground-target density and discriminate spatially uncorrelated deformation, improving the characterization of landslides with spaceborne InSAR. Even though applied here to RADAR data of moderate spatial and temporal resolution, HDS-InSAR has provided detailed characterization of a large, dynamic urban slope. Applying the technique to higher spatial resolution datasets will allow even more detailed characterization of landslide creep, and using sensor constellations will help increase the temporal resolution of displacement records. The HDS
20 technique also provides insight into the precise locations of landslide margins and is thus particularly useful in dense urban settings where development and remedial works remove evidence that is useful in landslide investigations.

8 Conclusions

The 2011 Pampahasi landslide is only one event in the long-term, and likely on-going, activity of a meta-stable slope. High-spatial-density InSAR results demonstrate that several parts of the slope were active over the years prior to the landslide, and
25 nearly all regions with coherent ground-motion records within and adjacent to it experienced post-failure increases in the extent and rate of creep. This change in ground deformation is counter to the expectation that slopes commonly stabilize, at least temporarily, following a discrete failure and demonstrates that such an assumption, at least for short-term stability, is imprudent for some slopes.

InSAR-derived deformation patterns provide new insight into failure mechanisms of the Pampahasi slope. The complex failure comprises multiple unstable parts, including many areas of shallow instability and at least one large creeping block. Accelerating creep, a typical feature of first-time and reactivated failures, was not apparent leading up to the disastrous landslide on 26 February 2011. Any increase in displacement during the 13-day window following the last pre-failure

5 RADAR scene (13 February) apparently was not perceived by residents on the slope. The absence of detectable acceleration may instead reflect the achievement of steady-state creep along an existing surface, which is supported by the presence of previously failed sediments and the similarity of the lower slope gradient to the residual strength of the weakest of these units. Increased pore pressures resulting from particularly heavy rainfall triggered a large landslide event that temporarily perturbed the dynamic equilibrium of the slope. The failure reorganized stresses in the slope as well as beneath the plateau

10 above it, leading to a new, although possibly temporary, state of dynamic equilibrium affecting even more of the slope.

Improved understanding of instability of the Pampahasi slope is instructive in evaluating and reducing risk from large-scale slope instability in the city of La Paz, especially given the large extent of paleolandslides in the city and their close association with recent landslides. The Pampahasi slope is similar to other slopes in the city where many of the largest historic landslides have occurred; these slopes comprise heterogeneous, generally weak, and fine-grained sediments of the

15 La Paz Formation. Similar slopes, particularly those descending from the west margin of Pampahasi Plateau, are thus key sites for future stability and risk assessments. Evaluating the possibility of future, large-scale reactivations of these slopes requires consideration of high-rainfall scenarios and should not be based solely on creep acceleration. In light of the evidence presented here of enhanced post-failure activity, creeping slopes in La Paz are generally not appropriate for new settlements or resettlement of residents displaced by disasters.

20 Optimizing InSAR characterisation of landslides requires processing methodologies that afford high ground-target density and discriminate spatially uncorrelated deformation. The improved understanding of the activity of the Pampahasi area reflects the detailed spatial characterization made possible by the HDS-InSAR technique. Applying advanced InSAR processing techniques such as this to modern and next-generation RADAR datasets that, due to continued technological improvement, have higher ground resolution and revisit frequency will improve understanding of complex unstable slopes.

25 **Author contribution**

NJR, JJC, RLH and BTR conceived the investigation. NJR processed RADAR datasets with guidance from BTR. All authors participated in fieldwork. MAG and EM arranged field logistics and provided crucial details on the history of development and instability. NJR, JJC, BTR and RLH interpreted observations and results. NJR prepared the manuscript with contributions from all co-authors.

Competing interests

The authors declare that they have no conflict of interest.

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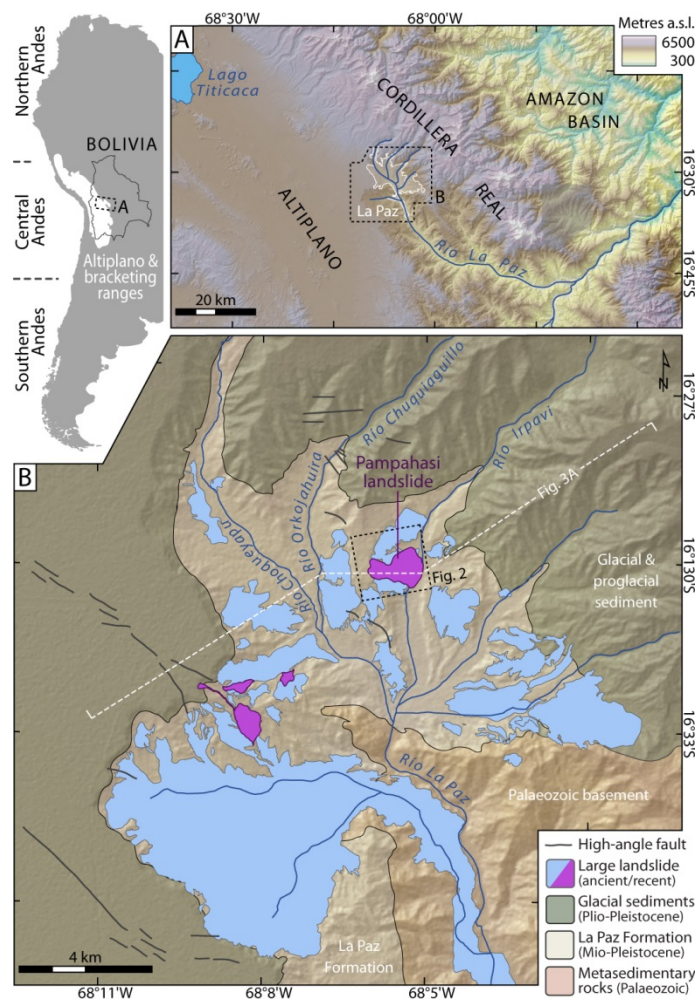
Figure 1. Setting of La Paz, Bolivia. A. Location and physiographic context. **B.** Generalized geology after Dobrovolny (1962) and Anzoleaga et al. (1977), with faults from Dobrovolny (1962) and Lavenu (1978). Terrain is from the ASTER GDEM 2 produced by METI and NASA.
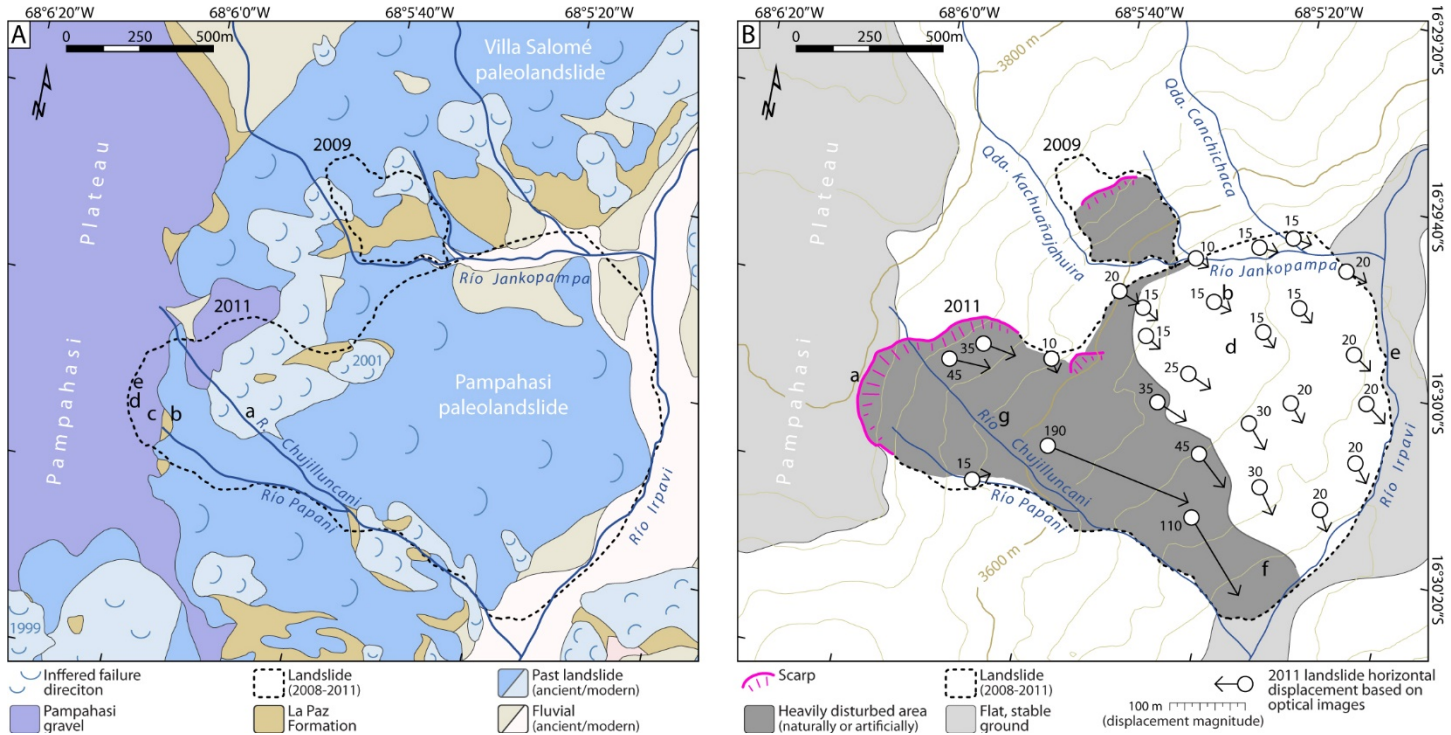
Figure 2. Ground surface conditions in the Pampahasi area. A. Surficial geology prior to the start of the twenty-first century; adapted from Anzoleaga et al. (1977). **B.** Slope morphology and ground disturbance showing the extent and morphologic features of landslides that occurred during the period of RADAR acquisition (September 2008 to December 2011). Horizontal displacements during the 2011 Pampahasi landslide are based on comparison of pre-failure and post-failure WorldView-2 scenes. Contours are from NASA's Shuttle RADAR Topographic Mission V3.0.
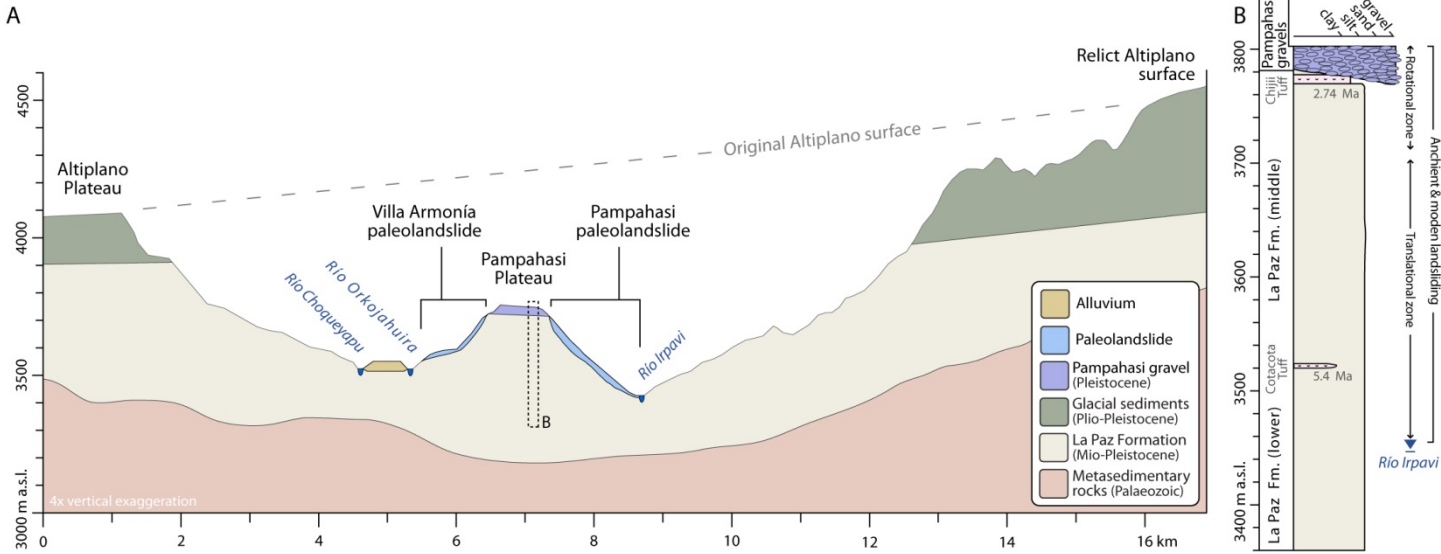
Figure 3. Stratigraphic sequences. A. Cross-section oblique to the trend of the Cordillera Real through fills underlying the Altiplano (location in Fig. 1B). **B.** Lithostratigraphy beneath Pampahasi Plateau, including principle units involved in the Pamphasi paleolandslide and 2011 Pampahasi landslide. Modified from Anzoleaga et al. (1977) based on field mapping throughout the La Paz area.

Figure 4. Field evidence of pre-failure creep in what would become the head region of the 2011 Pampahasi landslide. A. Small recent slumps (arrows) along Río Chujilluncani, ~400 m downslope from the east margin of Pampahasi Plateau. **B.** Minor cracking and centimeter-scale vertical offset of a paved road about 200 m east of, and 75 m below, the east margin of the plateau. **C.** Unpaved road with ~1 m displacement near the 2011 headscarp. **D.** Vertical offset along a gully ~75 m from the future south lateral margin of the 2011 landslide. The side of the gully nearer the middle of the 2011 landslide (right of the photo) has dropped ~0.5 m relative to the adjacent slope, causing deformation of the gabions and curb in the foreground. **E.** Cracks in an adobe wall (right foreground) and brick house (left middle ground) ~50 m downslope of the east margin of the plateau, corresponding to the middle of the 2011 landslide headscarp. See Figures 2A and 7A, B for locations of photographs. Photos taken by R.L. Hermanns in October 2005.

Figure 5. Impacts of the 2011 Pampahasi landslide. A. Headscarp (80 m height) along the east margin of Pampahasi Plateau (note excavator and trucks in lower middle ground for scale). **B.** Ground fissures in the northern part of the lower landslide mass. The wall in the background is ~2 m high. **C.** Back-tilted houses and apartment buildings on the northern part of the lower landslide mass (movement was from left to right). **D.** Community building with collapsed wall on the northern part of the lower landslide mass. **E.** Bridge across Río Irpavi that was deformed and shifted from its footings on 28 February near the centre of the landslide toe. **F.** Buildings on the southern part of the toe of the landslide \that collapsed on 28 February and 1 March (dates of collapse shown; photo by M.A. Guzmán, 2 March 2011). **G.** Water service line ruptured by the landslide (upper pipe) and replacement water service line (lower pipe). See Figures 2B and 7C, D for locations of photographs. Photos taken by N.J. Roberts in May 2011 unless otherwise indicated.
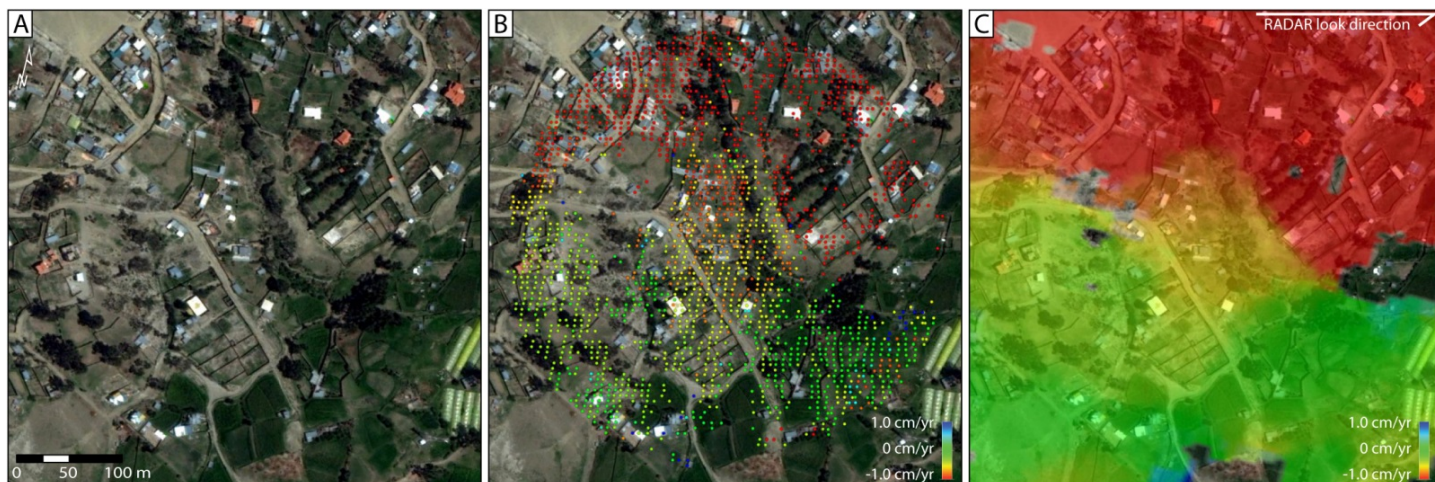
Figure 6. Typical density and spacing of HDS-InSAR results for an area of mixed suburban, agricultural, and commercial development near the toe of the Pampahasi slope from the pre-landslide stack. A. Satellite image showing land cover without InSAR data. **B.** Point data representing HDS neighbourhoods (pixel neighbourhoods), each recording a displacement time history. **C.** Linear displacement map interpolated from HDS points. Displacement rates are in the satellite line of sight. The deforming area is the southern boundary of the active part of the landslide toe. Base image is a 4 January 2011 Quickbird image viewed in GoogleEarth™. See Figure 7A for setting within the Pampahasi area.
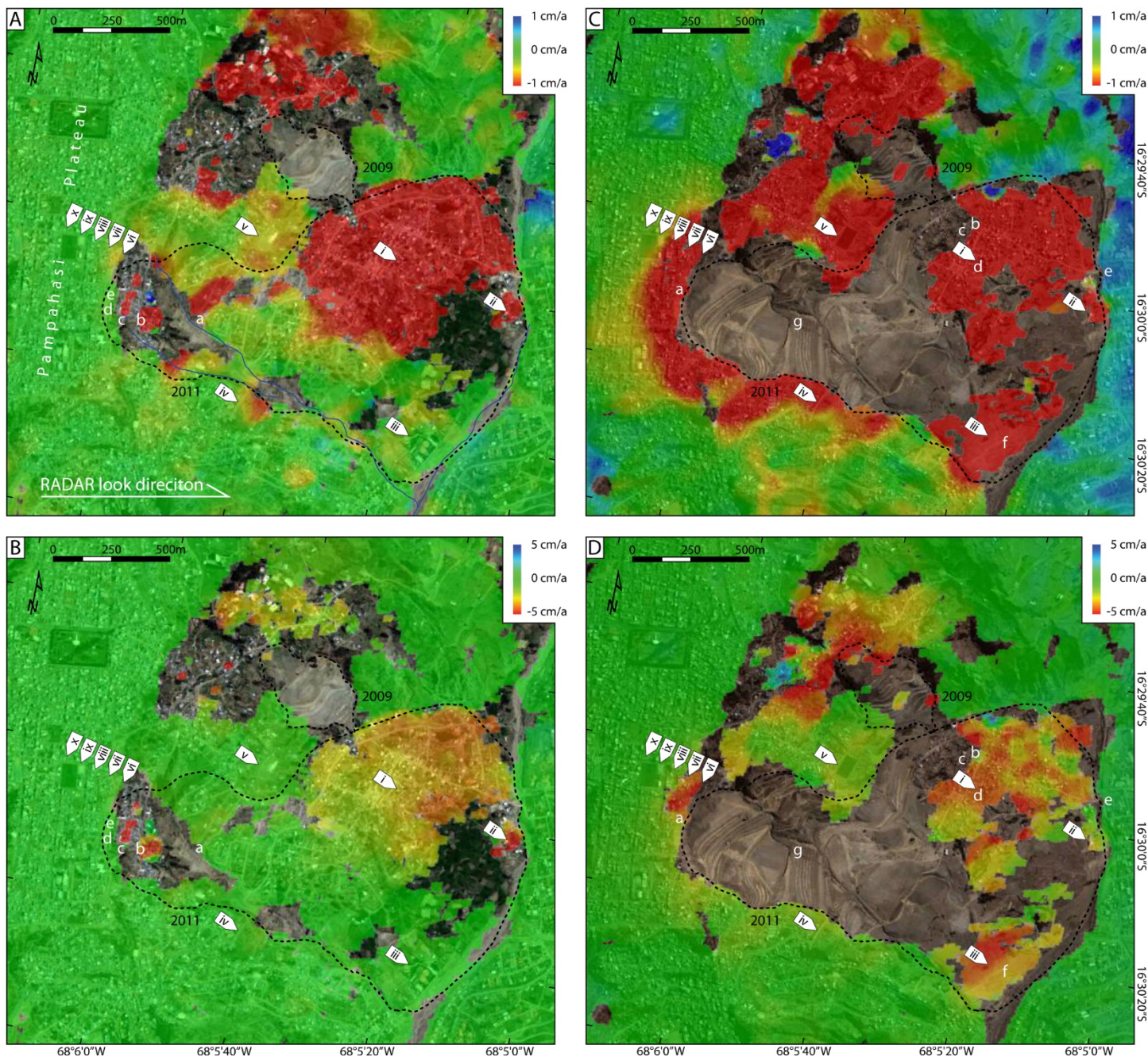
Figure 7. InSAR-measured, line-of-sight ground displacements in the Pampahasi area. Average linear deformation rates during (**A, B**) the 30-month period before and (**C, D**) the 10-month period after the Pampahasi landslide. Compressed (-1 to 1 cm/a; A and C) and expanded (-5 to 5 cm/a; B and D) displacement scales emphasize, respectively, the spatial limits and spatial variability of slope activity. Dashed lines are the limits of landslides that occurred in 2009 and 2011 during the period of RADAR scene acquisition. Roman numerals

show locations of displacement histories in Figure 5. Letters show locations of photos taken, respectively, before (Fig. 4) and after (Fig. 5) the 2011 landslide. Base images are from GoogleEarth ([A, B] 4 January 2011 and [C, D] 25 July 2011).
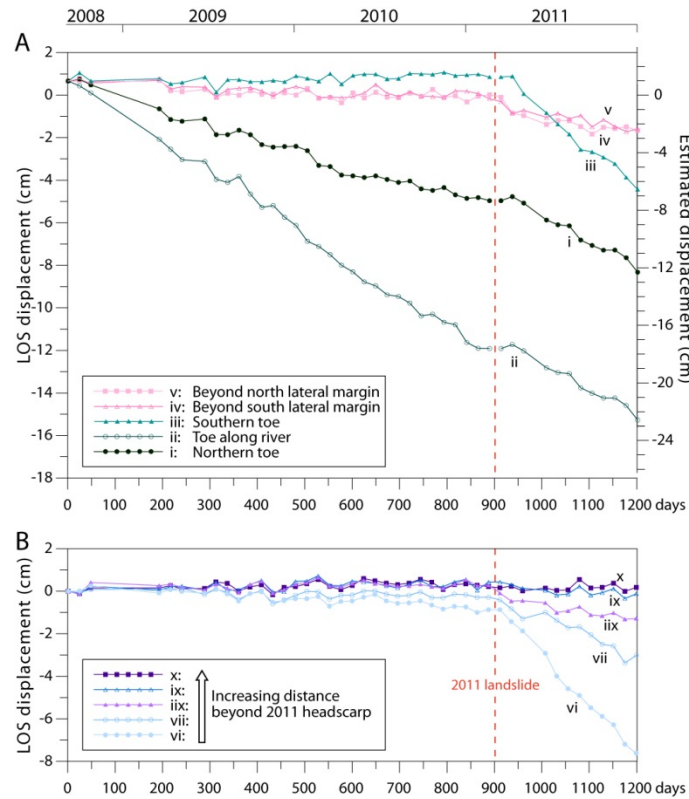
Figure 8. Displacement time series from the Pampahasi area (locations 'i' to 'x' in Fig. 7). A. Movement of the west slope of the Río Irpavi valley within the area affected by the 2011 landslide (i-iii) and adjacent paleolandslide deposits (iv-v). **B.** Movement on Pampahasi Plateau behind the headscarp of the 2011 landslide (vi-x). Variability in the data, which is most pronounced in parts of the time series with low displacement rates, indicates phase noise resulting from factors unrelated to ground motion. Displacements between the end of the pre-failure stack and the start of the post-failure stack are not precisely constrained within the limits of the 2011 landslide (i-iii). For the Pampahasi slope adjacent to the 2011 failure and for the Pampahasi Plateau, displacements between the stacks are based on results of full-stack processing (see Supplementary Material).
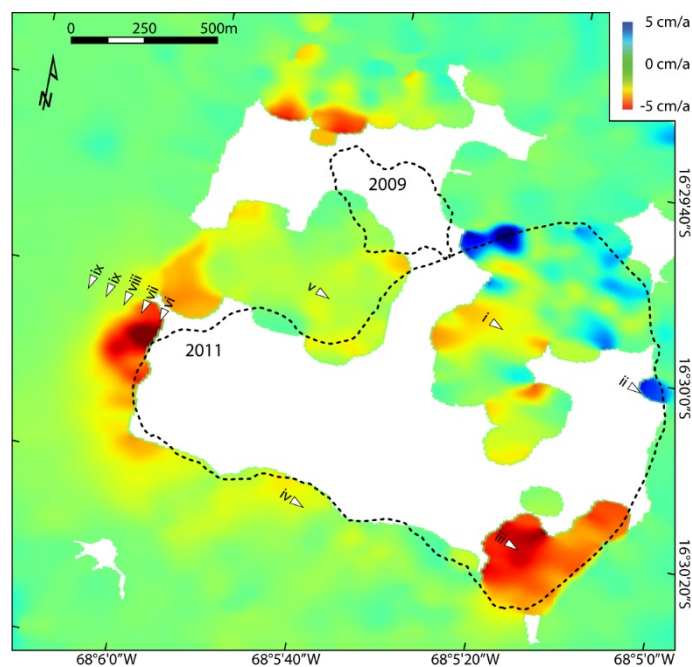
Figure 9. Change in line-of-sight slope creep of the Pampahasi area following the 2011 failure event. Positive and negative values indicate, respectively, decreased and increased post-failure displacement rates away from the satellite. Green areas experienced no change in displacement, and were largely stable both before and after the 2011 landslide. White areas lacked coherence phase in pre-failure scenes, post-failure scenes, or both. Roman numerals show locations of displacement histories in Figure 5.

Natural Hazards
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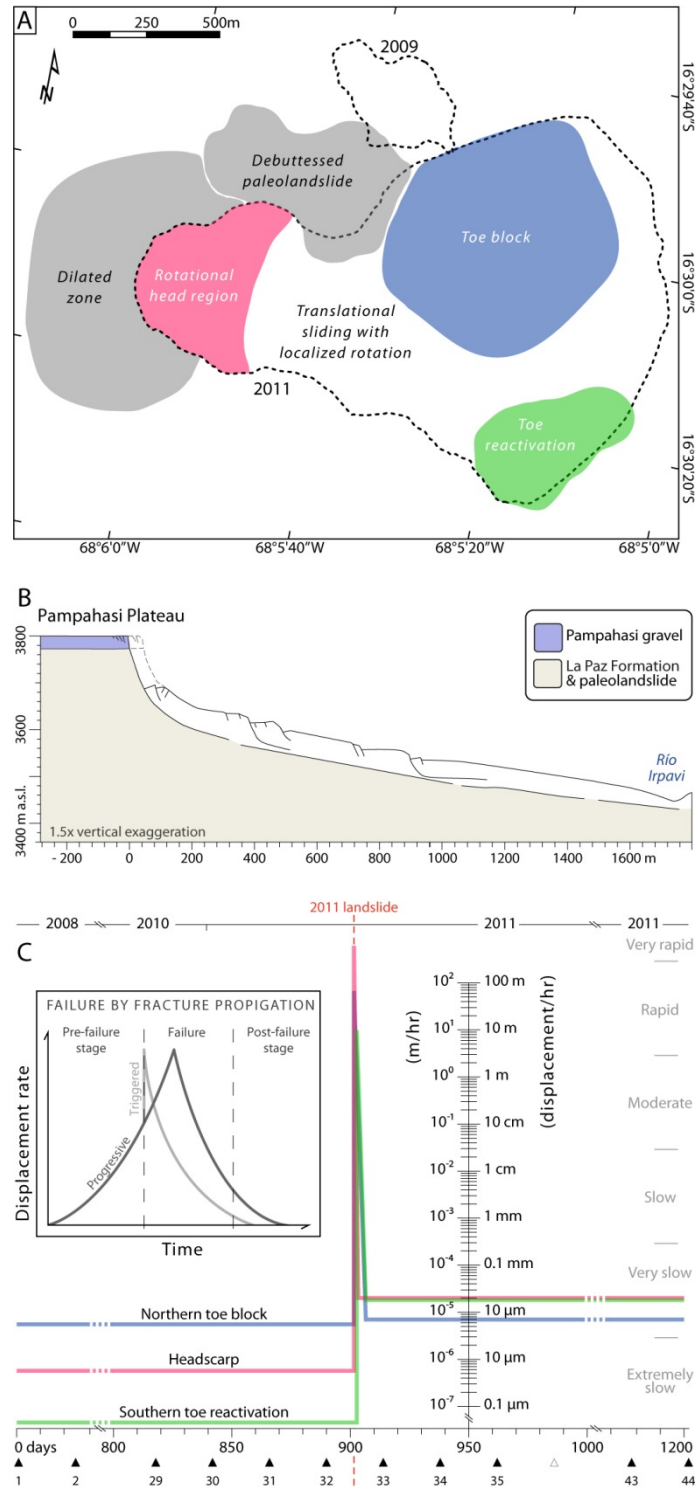
Figure 10. Proposed geomechanical model for the Pampahasi landslide and adjacent terrain. A. Components of failure, which differ in kinematics and activity history. **B.** Generalized section through the failing slope illustrating the approximate depth and attitude of the failure zone. Topographic surface based on NASA's Shuttle RADAR Topographic Mission V3.0. **C.** Approximate temporal displacement history based on InSAR measurements bracketing the 2011 Pampahasi landslide and eyewitness accounts. Due to the revisit frequency of
5 RADARSAT-2 (24 days), the lack of acceleration leading up to the 2011 Pampahasi landslide is based on the lack of such effects reported by the slope's residents. Velocity classes are from Cruden and Varnes (1996), spanning all classes except for 'extremely rapid' (>5 m/s). Triangles denote timing of RADARSAT-2 acquisitions (dates provided in Table S4). Inset shows schematic stages of slope movement during first-time failure by fracture propagation (after Hermanns and Longva, 2012).