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Landslide monitoring using seismic refraction tomography – The importance of incorporating topographic variations

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Abstract

Seismic refraction tomography provides images of the elastic properties of subsurface materials in landslide settings. Seismic velocities are sensitive to changes in moisture content, which is a triggering factor in the initiation of many landslides. However, the application of the method to long-term monitoring of landslides is rarely used, given the challenges in undertaking repeat surveys and in handling and minimizing the errors arising from processing time-lapse surveys. This work presents a simple method and workflow for producing a reliable time-series of inverted seismic velocity models. This method is tested using data acquired during a recent, novel, long-term seismic refraction monitoring campaign at an active landslide in the UK. Potential sources of error include those arising from inaccurate and inconsistent determination of first-arrival times, inaccurate receiver positioning, and selection of inappropriate inversion starting models. At our site, a comparative analysis of variations in seismic velocity to real-world variations in topography over time shows that topographic error alone can account for changes in seismic velocity of greater than ±10% in a significant proportion (23%) of the data acquired. The seismic velocity variations arising from real material property changes at the near-surface of the landslide, linked to other sources of environmental data, are demonstrated to be of a similar magnitude. Over the monitoring period we observe subtle variations in the bulk seismic velocity of the sliding layer that are demonstrably related to variations in moisture content. This highlights the need to incorporate accurate topographic information for each time-step in the monitoring time-series. The goal of the proposed workflow is to minimize the sources of potential errors, and to preserve the changes observed by real variations in the subsurface. Following the workflow produces spatially comparable, time-lapse velocity cross-sections formulated from disparate, discretely-acquired datasets. These practical steps aim to aid the use of the seismic refraction tomography method for the long-term monitoring of landslides prone to hydrological destabilization.

Keywords

seismic refraction, geophysical monitoring, active landslides, topographic change, hydrogeophysics

1. Introduction

The implementation of robust and appropriate monitoring strategies is critical for the ongoing assessment of potentially destabilising processes in landslide systems (Angeli et al., 2000). Near-surface geophysical methods are increasingly used to monitor the subsurface conditions of landslides susceptible to hydrological destabilization (Perrone et al., 2014, Whiteley et al., 2019), most commonly by active-source DC electrical resistivity (ER) (e.g., Lucas et al., 2017, Uhlemann et al., 2017) and passive-source seismic monitoring (e.g., Walter et al., 2012). ER can provide information on the moisture dynamics of an unstable slope, and passive-source seismic can provide information on the kinematics of failure events. One major advantage of active-source geophysical methods, such as ER, is their ability to produce spatially high-resolution, time-lapse tomographic images of the subsurface. However, the majority of seismic landslide monitoring campaigns utilise passive-source methods, which provide superior temporal resolution, but are limited in their spatial resolution due to practical limitations on the number of sensors in an array.

Seismic refraction tomography (SRT), an active-source seismic method, can characterize the spatial heterogeneities in elastic properties of materials in landslide systems (e.g., Uhlemann et al., 2016). SRT determines the travel-time of artificially generated seismic waves, to build up a series of travel-times for waves propagating through the subsurface. These travel-times are inverted to produce subsurface models of seismic velocity. The two types of body waves used in SRT, P-waves and S-waves, propagate through subsurface media at different speeds depending on lithological and physical properties. The P-wave velocity, V_p, is given by

$$V_p = \sqrt{\frac{K + \frac{4}{3}G}{\rho}}, \tag{1}$$

in which K is the bulk modulus (a measure of a material's resistance to uniform compression), G is the shear modulus (a measure of a material's resistance to shear strain) and ρ is material density. The S-wave velocity, V_s, is given by

$$V_s = \sqrt{\frac{G}{\rho}}. \tag{2}$$

In solid rock, the relationship between seismic velocity and saturation has been empirically demonstrated, and is relatively well understood. Considering a fully saturated rock, as liquid (with a higher *K;* Equation 1) in pore spaces initially become replaced by gas, V_p decreases rapidly and V_s decreases with increased saturation due to changes in bulk density and shear modulus (Equation 2) (Wyllie et al., 1956). These seismic attributes and their relationship to the petrophysical properties of rock can be used to determine the effects of saturation on seismic velocity (e.g., Biot, 1956, Gassmann, 1951).

In soils, the effect of variations in saturation on seismic velocity is less well-understood. Existing evidence indicates that both the distribution of moisture throughout the soil structure, as well as the influence that capillary forces have on effective pressure, influence V_p at small scales (Romero-Ruiz et al., 2018). Experiments in artificial, well-mixed, homogenous soils, have demonstrated that V_p decreases with increasing saturation (Lu and Sabatier, 2009) and similar results have been obtained from laboratory measurements on undisturbed samples of Loess soils (Flammer et al., 2001). These decreases in V_p are dominated by changes in the matric potential of the soil (related to capillary forces). The effects of capillary forces are likely to be very different between artificial and natural soils, with the former having no internal structure or little consolidation, both of which reduce the influence of capillary forces.

Despite this lack of understanding on the precise mechanism by which seismic velocities are influenced by moisture content in soils, seismic attributes are still routinely used in larger scale field studies to assess characteristics of near-surface sediments. The ratio between V_p and V_s (V_p/V_s) can be used to assess lithology, strength and quality, structure and saturation of near-surface sediments for geotechnical investigations (Bhowmick, 2017). Seismic surveys to obtain in-situ V_p, V_s and V_p/V_s have been used to image physical properties, including ground saturation, in the field (Dashwood et al., 2019, Pasquet et al., 2016b), and have been used to monitor shallow saturation processes in the laboratory (Pasquet et al., 2016a). Poisson’s ratio, a property closely related to V_p/V_s ratio which measures lateral strain to axial strain, has been shown to relate to porosity in near-surface sediments, and can be used to determine areas of localised saturation (Uhlemann et al., 2016, Uyanık, 2011).

The use of SRT as a tool for long-term landslide monitoring is almost absent from the literature. Examples of active-source seismic landslide monitoring campaigns focus on the characterization of surface fissures (see Bièvre et al., 2012, Grandjean et al., 2009) rather than the monitoring of moisture-induced elastic property variations. The dearth of studies using SRT as a long-term monitoring tool for landslides is likely due to the complexity of managing and minimizing the several sources of error in the individual surveys (i.e., time-steps) that comprise a monitoring time-series. In this study, we present a methodology to acquire, process and invert a long-term SRT time-lapse dataset collected from an active landslide. To our knowledge, the use of SRT in a monitoring campaign at an active landslide site has not previously been implemented. The methodology is applied to time-lapse SRT monitoring at a site of active slope failure in North Yorkshire in the UK. This study aims to develop a practical approach for active-source time-lapse seismic surveying of vulnerable slopes, and to demonstrate the applicability of high spatial resolution subsurface monitoring using SRT. The approach taken is summarised in a workflow, from which a practical walkthrough of how the time-lapse SRT data were acquired, processed and inverted using a novel two-stage inversion procedure is presented. The importance of incorporating the topography of the landslide surface from every survey (i.e., for each time-step) in a monitoring campaign is highlighted. Summary results from the SRT monitoring campaign are presented and discussed, and support the use of SRT to monitor moisture dynamics at active landslide sites. The approach and results of this study should be of interest to researchers studying the evolution of subsurface processes acting to destabilise landslide systems (Jaboyedoff et al., 2019), and to those using geophysical methods in landslide early-warning systems (Intrieri et al., 2012) and monitoring environmental changes.

2. Seismic refraction tomography monitoring at the Hollin Hill Landslide Observatory

The Hollin Hill Landslide Observatory (HHLO) in North Yorkshire, UK (Chambers et al., 2011, Merritt et al., 2013), is operated by the British Geological Survey. The landslide comprises an interbedded series of Lower and Middle Jurassic sandstones and mudstones (Figure 1), namely the Whitby Mudstone Formation (WMF) and Staithes Sandstone Formation (SSF).

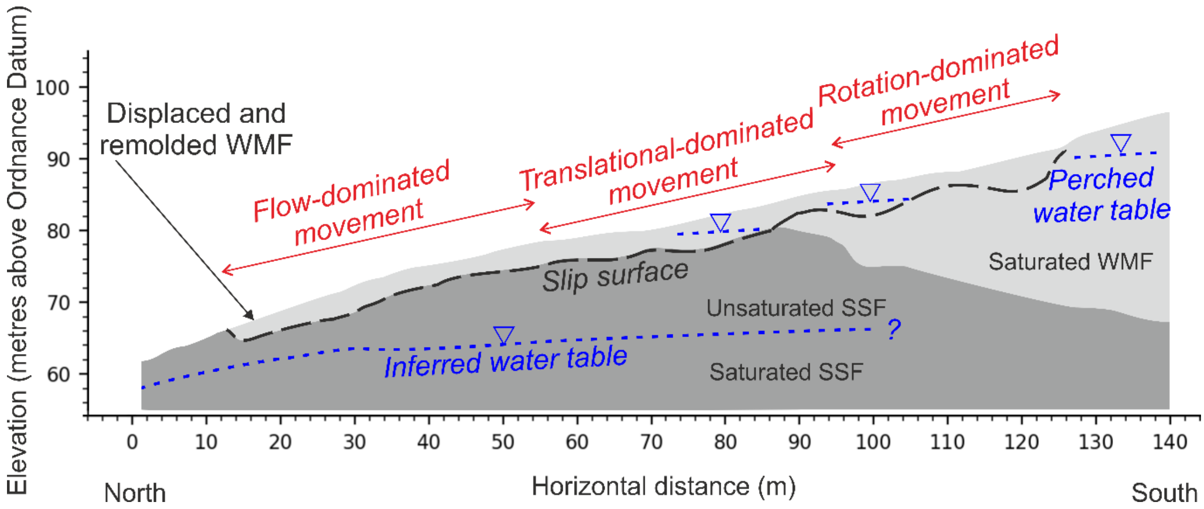

Figure 1: A simplified conceptual model of the HHLO (modified from Uhlemann et al., 2016), indicating movement domains, slip-surface, indicative positon of water tables, and main lithological units comprising the Whitby Mudstone Formation (WMF) and Staithes Sandstone Formation (SSF).

The moisture content of the WMF controls displacement occurrence at the site. The WMF is of low permeability and drains slowly into the underlying SSF. Hence, during periods of increased precipitation, moisture content within the WMF increases quickly (creating localized perched groundwater tables), and decreases slowly during periods of lower precipitation. Slope failure is most likely during these periods of intense and prolonged rainfall.

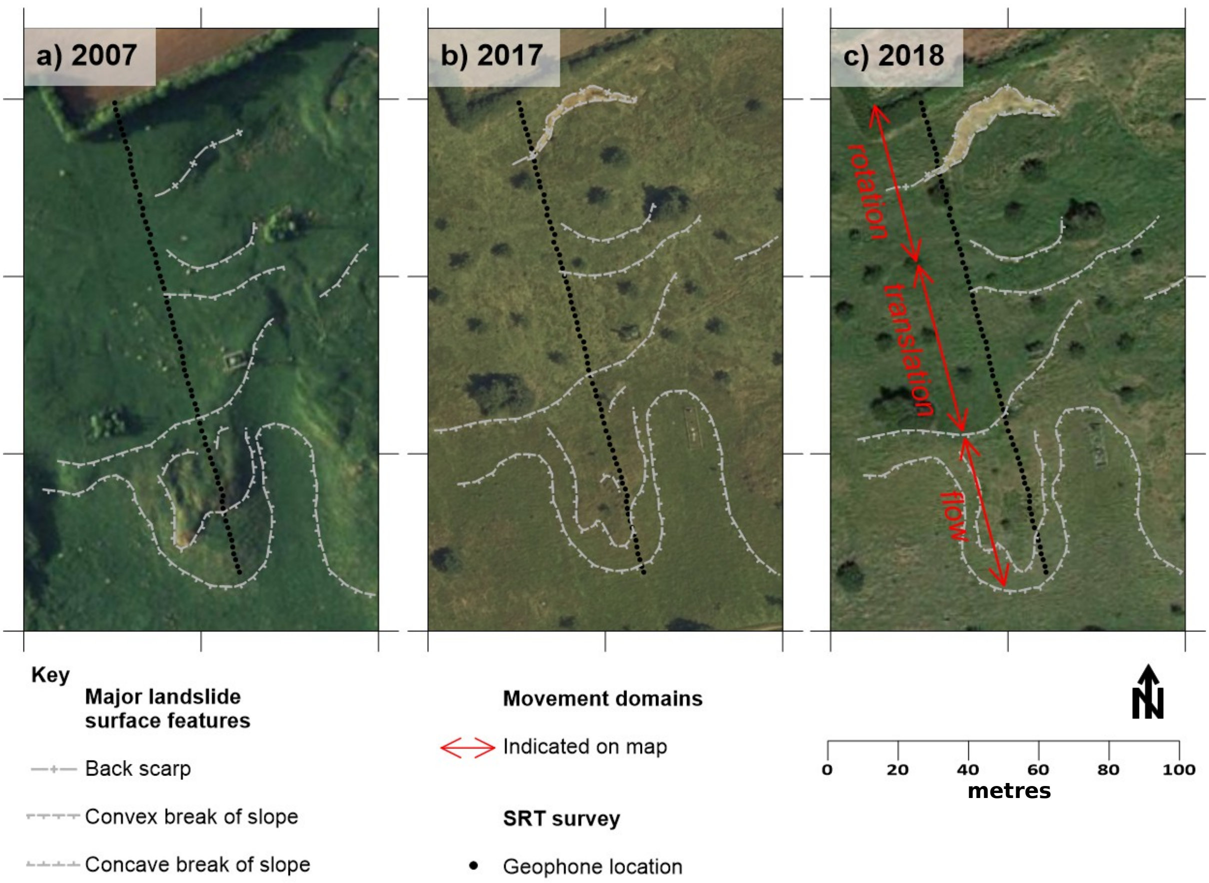

Figure 2: Aerial photographs from the HHLO. a) Image from 2007 showing the main features of the landslide, including backscarps at top of slope (north), and flow-lobes at base of slope (south). Map data: Google, Infoterra Ltd and Bluesky. b) Image from 2017 showing development of new backscarp after movement in 2016. Map data: Digimap. c) Continued backscarp development shows landslide extension, and propagation of the backscarp to the west. Map data: Google. Black dots are the indicative locations of receivers used in the SRT surveys, with the first receiver location (northern most dot) located outside of the active landslide area, acting as a static reference point against which the receiver arrays are deployed. The location of this receiver is marked by a ground peg installed at the site.

Seasonal variations in moisture content, linked to regional groundwater levels and local infiltration of rainwater, decrease restraining soil-suction forces (potentially producing destabilising positive pore-water pressures) initiating movement at the slip-surface mid-slope. This translational displacement propagates uphill as support for overlying material is removed, culminating in the development and widening of rotational backscarps in the saturated WMF at the top of the slope. Downslope, mobilised material is reworked to form flow lobes at the base of the landslide, where movement is eventually arrested through drainage to underlying deposits of well-sorted, aeolian quaternary sands deposited at the top of the SSF. Aerial imagery from 2007, 2017 and 2018 shows the development of geomorphological landslide features at the HHLO (Figure 2).

SRT monitoring at the HHLO aims to identify changes in the elastic properties of the underlying lithological units. These variations in elastic properties are primarily driven by variations in slope moisture dynamics. Between October 2016 and August 2019, 16 SRT surveys were acquired, resulting in the production of 16 V_p and 16 V_s cross-sections spanning a period of 1001 days, close to 33 months. The length of time of the monitoring period allowed data to be collected over two distinct annual climatic cycles, ensuring data were acquired at different subsurface moisture contents, and during multiple wetting and drying phases of the landslide system, capturing temporal heterogeneity in hydrological condition. Data were acquired at an average return interval of 9 weeks, which was deemed to be practicable given the characteristics of the landslide system and long-term monitoring period. A shorter return interval would have been desirable, but this was prevented by the logistical and financial cost of mobilisation, equipment availability and deployment, and acquisition and processing time associated with each survey; surveys typically involved two to three days of fieldwork, followed by several days of data processing.

The SRT surveys were acquired along the same profile location over the duration of the monitoring campaign. The profile comprised of 2 m spaced geophones (i.e., receivers), positioned from the crest of the landslide to the toe (Figure 2).

The location of the survey profile was chosen based on previous geophysical surveys that have been undertaken at the site (see Uhlemann et al., 2016) and position of geotechnical sensors (see Merritt et al., 2013). For both the P- and S-wave surveys, a 48-channel ABEM Terraloc Mk6 was used to acquire seismic refraction data. To acquire contiguous data from the entire spread length (142 m total length, comprising 72 receiver locations), two separate 48 receiver (94 m long) profiles with a 46 m overlap between the surveys were acquired. Receivers used in both deployments were not moved between spread acquisitions, and repeat shots were undertaken so that the overlapping spreads could be processed as a single profile of data.

Vertical geophones with a dominant frequency of 8 Hz were used as receivers for the P-wave survey, and a 4 kg sledgehammer and horizontal steel plate were used as a source. At each shot location, data were recorded for 1 second, in order to acquire both refracted P-wave arrivals and surface wave data (these latter data are not described in this study). Shot records were stacked in the field, and the number of stacked shot records varied between surveys based on environmental conditions, such as wind speed and rain; a minimum of two stacks per location were acquired in optimal conditions (i.e., low or no wind and rain), and up to six stacks per location were acquired in poorer conditions.

For the S-wave survey, horizontal geophones with a dominant frequency of 14 Hz were used as the receivers, and a prism with ~45° inclined face was used to generate S-waves in opposing polarisations, perpendicular to the orientation of the receiver profile (Uhlemann et al., 2016). Data were recorded for 0.5 seconds, and same-polarisation shot records were stacked, with a minimum of two stacked per receiver location saved in optimal survey conditions, and up to a maximum of six shot records per location saved in poor survey conditions.

In both surveys, geophones were buried to a depth of ~10 cm below ground level in an attempt to isolate the receivers from aerial environmental noise, and to provide better coupling with the subsurface. Shots were acquired at every other receiver location (i.e., every 4 m) for the whole of the receiver spread, starting at the first receiver at the crest of the slope. It was not possible to acquire off-end shots at the top of the profile (i.e., before the first geophone) due to access restrictions. For the P-wave surveys, off-end shots at the end of the of the spread were acquired at 4m intervals beyond the penultimate receiver at the toe of the slope to a maximum off-end distance of 22 m beyond the last receiver (i.e., 164m from the first receiver). For the S-wave surveys, off-end shots were acquired at 10 m intervals to a distance of 20 m beyond the last receiver (i.e., 162 m from the first receiver). For both surveys, the same shot locations were used throughout the entire monitoring campaign, ensuring consistent spatial coverage between surveys.

3. Overcoming challenges in long-term SRT monitoring of landslides

In this study, several sources of error in SRT surveys need to be accounted for during data acquisition, and in the subsequent data processing and inversion stages. Some of these sources of error are unique to landslide monitoring. The goal during processing is to minimize transient changes in time-lapse data that may arise from differences in survey set-up and processing of data between surveys, and to preserve changes arising from genuine variations in the properties of landslide materials. As velocity is the ratio of distance and time, the determination of accurate velocities relies on accurate positioning (i.e., determining true distances) and correct picking (i.e., identifying correct travel-times) of data. The major sources of potential error in SRT acquisition and processing are identified in Table 1, along with their solutions presented in this study and stage at which the error should be considered. A workflow to produce a robust seismic velocity time-series, taking into consideration the potential sources of error identified in Table 1, is shown in Figure 3. In this study, the 1001 day monitoring period is considered as a ‘time-series’ of SRT data, with the 16 individual surveys comprising ‘time-steps’ within this time-series. The following sections describe how the stages of the workflow are used to address the potential errors listed in Table 1 that occur within individual time-steps, and across the time-series as a whole.

Potential source of error:	Addressed by use of:	Addressed during:
Inconsistent repositioning of receivers to same locations between surveys	Permanent reference points in the field for repeatable receiver deployment	Data acquisition
Failure to capture accurate 3D locations of receivers deployments, and subsequent differences in positions between surveys	RTK-GNSS systems to obtain accurate 3D receiver positions	Data acquisition
Inconsistent data coverage within surveys and across the time-series	Repeatable field procedures to boost data coverage; common travel-time maxima across time-series to give consistent data coverage	Data acquisition/ data processing
Inconsistent picking of first arrivals	Reciprocal error analysis to identify data for re-processing	Data processing
Not accounting for changes in surface topography between	Unique inversion for each survey (i.e., time-step) derived from individual RTK-	Data inversion

surveys	GNSS acquired topography	
Not incorporating errors in to the inversion	Error model derived from reciprocal error analysis	Data processing/ data inversion
Using inappropriate constraints for initial data inversion	Tests to determine best starting parameters for first stage inversion	Data inversion
Using inappropriate constraints to regularize data over time	Selecting 'best' data from first stage inversion to use as a 'reference model' for second stage of inversion	Data inversion

Table 1: Potential sources of error arising from the acquisition, processing and inversion of time-lapse SRT data.

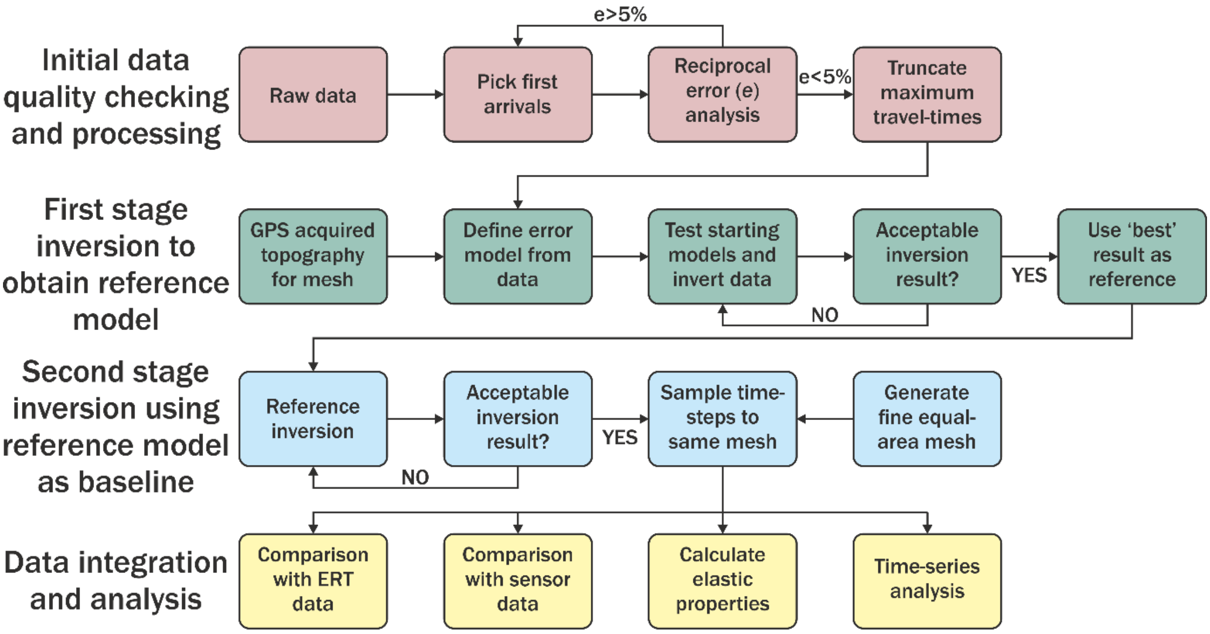

Figure 3: Proposed workflow for processing SRT surveys (i.e., time-steps) to produce a reliable time-series of time-lapse SRT data. SRT data are first processed using reciprocal data analysis for quality control. Individual SRT survey data are initially inverted to determine the best time-step, from which the resulting model us then used as a 'reference model' for all the time-steps in the time-series. Time-lapse SRT images are then created using unique topography acquired at each survey, in order to determine velocity changes in the subsurface between surveys.

3.1. Assessing first arrival quality

Identifying consistent, repeatable first arrivals in SRT data is a recognised challenge with no universally accepted solution. Attempts include using automatic picking algorithms (e.g., Khalaf et al., 2018) and using statistical approaches (e.g., Dangeard et al., 2018) to minimize absolute and relative errors introduced by operators picking time-lapse SRT data. In this study, reciprocal errors between inverse source-receiver configurations are used to identify 'bad' picks that display an unacceptable differential in reciprocal travel-time. Reciprocal measurements require receiver locations to be used as both receiver and source location during the course of the survey. Therefore, reciprocal analysis is undertaken on ~50% of the entire data for any given survey, and is used as a representative sample of the entire survey dataset i.e., a reciprocal error subset. The error (e) in a reciprocal measurement (defined as the mean travel-time of the two measurements) is defined as

$$|e| = 100 \bullet \left(\frac{|t_n - t_r|}{t_n + t_r} \right), \tag{3}$$

in which t_n is the travel-time between a source at position A, and receiver at position B, and t_r is the travel-time between a source at position B and a receiver at position A. Reciprocal measurements cannot always be acquired when intra-survey (i.e., within the same time-step) data coverage is inconsistent. Factors leading to poor data coverage include low signal-to-noise-ratio at greater source-receiver offsets and interference from noise sources, such as wind, rain and amplification of these noise sources through nearby trees (Figure 4a). Large reciprocal errors occur in travel-times with greater source-receiver offsets, and therefore the use of reciprocal measurements as a data quality indicator favors data acquired from the very near-surface (i.e., shots with smaller source-receiver offsets). Across all of the reciprocal error subsets from each time-step in this study, 12.5% of the V_p data and 14.1% of the V_s data are not analyzed due to lack of reciprocal measurements. Remaining reciprocal-pairs of measurements showing a discrepancy e>5% are re-examined and re-picked (Figure 4b). Shot records adjacent to a reciprocal-pair with e>5% are also considered during this procedure. The data are then re-analyzed, and any further measurements with e>5% are re-picked. This iterative process continues until all measurements in the dataset have e<5%.

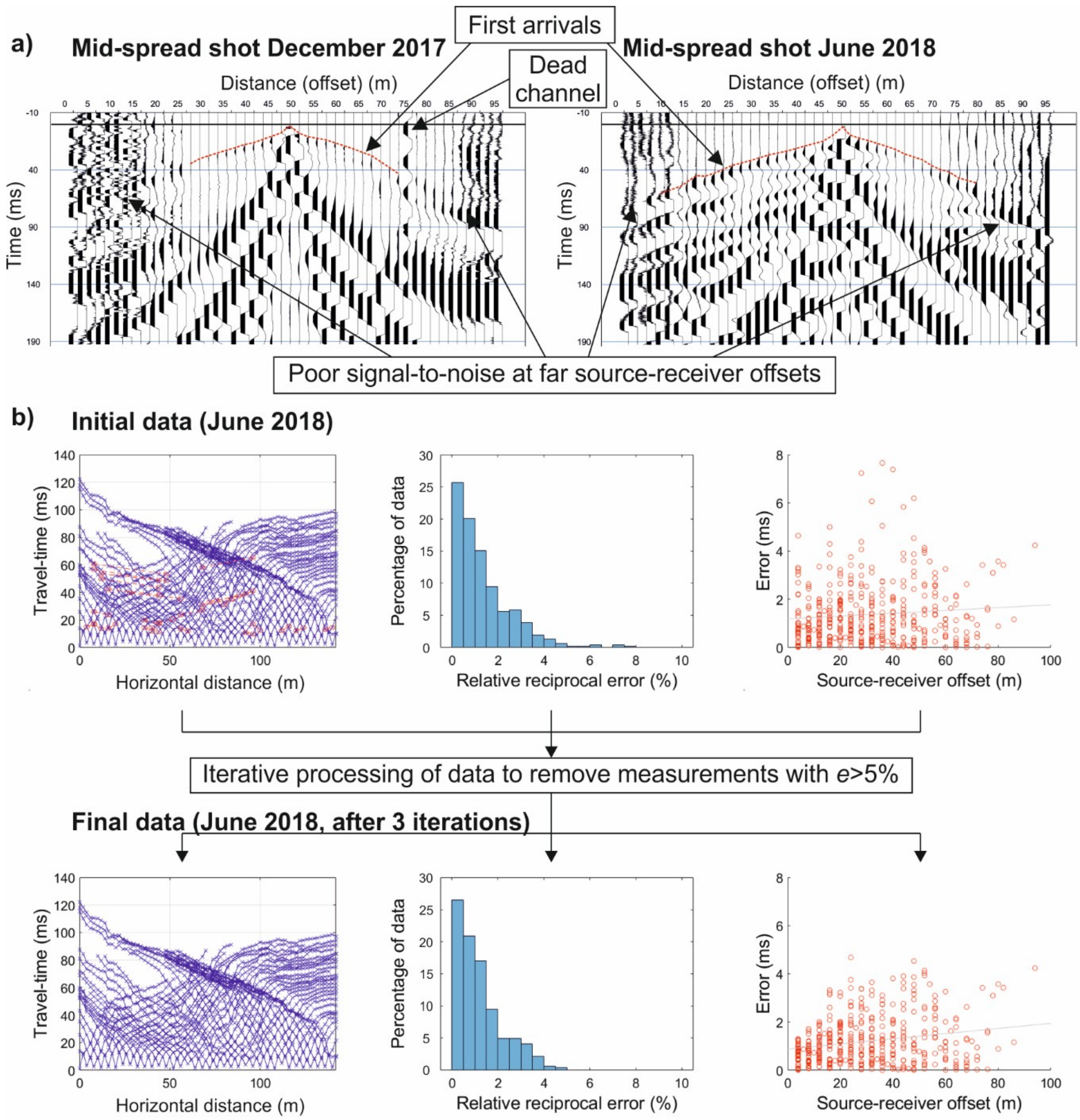

Figure 4: a) Examples of V_p shot records from the same position at the HHLO from December 2017 (left panel) and June 2018 (right panel). Poor signal-to-noise at larger source-receiver offsets prohibits the identification of first arrivals, and prevents acquiring reciprocal pairs for every measurement in the survey. b) The process of identifying reciprocal errors within a subset of the refraction survey data with e>5% *from V_p data from June 2017. Top left panel shows all first-arrival data (displayed as travel-time curves) with pairs of measurements of* e>5% *circled in red. Top centre panel shows the distribution of relative reciprocal errors within the reciprocal error data subset, and the top right panel shows the distributions of absolute reciprocal errors from this data subset as a function of source-receiver offset, indicating that shots with further offsets have higher errors. The corresponding panels below show the effect of iteratively identifying and re-picking data with e>5%, in order to reduce errors across the dataset.*

A further issue arising from implementing a time-lapse approach is achieving consistent inter-survey (i.e., between time-steps) data coverage over time. Consistent coverage cannot always be achieved due to variations in noise sources and environmental factors between surveys. For example, surveys undertaken in periods of increased ground moisture show higher signal attenuation compared to surveys undertaken in drier conditions, and subsequently individual surveys show a significant variation in maximum recorded travel-times. Without normalization of these maximum travel-times, comparison of the inverted sections to determine an appropriate reference model (see Section 3.3) is challenging, primarily due to differences in maximum travel-times inducing significant variations in the maximum depths of coverage in the inverted models. To overcome this, the distribution of all travel-times from across the monitoring period is plotted, and a travel-time value that preserves the majority of the data is chosen as a cut-off. In this case, the chosen cut-off travel-times are 86 ms and 178 ms for the V_p and V_s data, respectively. Data with travel-times over this cut-off are discarded, creating consistency in coverage between time-steps, but reducing the total number of data points. Across all of the time-steps of this study, 1.5% of the V_p data and 17.1% of the V_s data are discarded, giving a common maximum travel-time between surveys. The V_s surveys tend to have better signal-to-noise ratios, which results in a higher number of reliable first arrivals being recorded, however, this ultimately means that more data are discarded in order to match the relatively poorer coverage achieved in the V_p surveys.

3.2. Using accurate topography

Repositioning of receivers to repeatable *x*, *y* and *z* positions on the landslide surface is crucial to ensure that seismic ray paths are sampling comparable domains of the subsurface over time. The positioning error in *x* and *y* can be minimized by deploying receivers relative to permanent markers located outside of the active area of the landslide, and recording absolute *x* and *y* positions for receiver locations. Furthermore, the slope surface (*z*) will change between surveys. This effect cannot be removed by accurate positioning, and therefore needs to be incorporated into the data processing. Variations in *z*, as well as small unavoidable discrepancies in *x* and *y* positions can be captured using accurate geodetic surveying methods.

In this study, receivers are deployed every 2 m, with the first receiver located outside of the active landslide area (i.e., above the backscarp) and deployed at the same absolute position for each survey. A permanent ground peg marks the location of this first receiver, and a tape measure draped across the ground surface is used to deploy the remainder of the survey profile relative to this location. A Real-Time Kinetic Global Navigation Satellite System (RTK-GNSS) is used to capture the absolute positions in x, y and z of all receivers with a precision <0.05 m. With accurate positional data for each survey, the 'line-of-sight' distance (d) between one receiver location with coordinates (x_i,y_i,z_i) and another with coordinates (x_{i-1},y_{i-1},z_{i-1}) can be expressed as

$$d_i = \sum_{1}^{i} \sqrt{(x_i - x_{i-1})^2 + (y_i - y_{i-1})^2 + (z_i - z_{i-1})^2} \quad . \qquad (4)$$

Topographic features between these receiver locations (i.e., those features smaller than the receiver spacing) are not captured in the data.

For accurate 2D seismic travel-time inversion, accurate elevations and surface distances of the receivers are required, as the fundamental problem to be solved is one of distance and time. It is common in SRT surveys for the elevations (z_i) of sensors to be recorded accurately, but for the inter-receiver spacing to be assumed to be a "fixed" nominal surface distance. This is particularly common in surveys on flat or uniformly dipping surfaces, where accurate inter-receiver spacing are easier to measure and control. However, in environments where topography can vary sharply within the receiver array, such as landslides, this approach can lead to errors in the positioning of receivers, which in turn introduces errors in to the generation of subsurface meshes for inversion, ultimately influencing the resulting inverted travel-times. Figure 5 shows the discrepancies that can arise from assuming a "fixed" nominal spacing (e.g., assuming receivers are deployed every 2 m, without accounting for the changes in distance that topography will create) with variable z_i measurements (red points) against using the true x_i, y_i and z_i positions to generate line-of-sight distances using Equation 4 in this study (green points). Using a "fixed" nominal spacing for time-lapse monitoring ignores lateral variations in receiver spacing, and results in an overestimation of array length. Acquiring topographic information at every survey (i.e., time-step) allows for accurate inversion of travel-times.

At the HHLO, the SRT profile is orientated to match the maximum slope profile, which is broadly parallel to the north-south orientation, and therefore the main direction of recorded wave propagation for the SRT survey was also in a north-south direction. Given the alignment of the y coordinate orientation with the direction of wave propagation, greater variations in the y coordinate of the receiver position (i.e., north-south orientation, parallel to slope) introduce larger errors to the results of the seismic survey if not accounted for. Variations in *x* coordinates (i.e., east-west orientation, perpendicular to slope) have a smaller effect on measurement accuracy, as they are perpendicular to the wave propagation direction. Between each survey, the mean variation in receiver repositioning is 0.03 m in the *x* coordinate (1.5% of receiver spacing), and 0.01 m in the *y* coordinate (0.5% of receiver spacing), which is below the nominal accuracy of the equipment used for data acquisition. Across the entire monitoring period, receiver positions vary by an average of 0.41 m in the *x* coordinate (20.5% of receiver spacing) and 0.15 m in the *y* coordinate (7.5% of receiver spacing). The increased accuracy of deployment in the *y* coordinate is due to the use of a tape measure deployed in the same orientation. Some active areas of the landslide experience much greater variations in topographic change due to slope displacements (Figure 5).

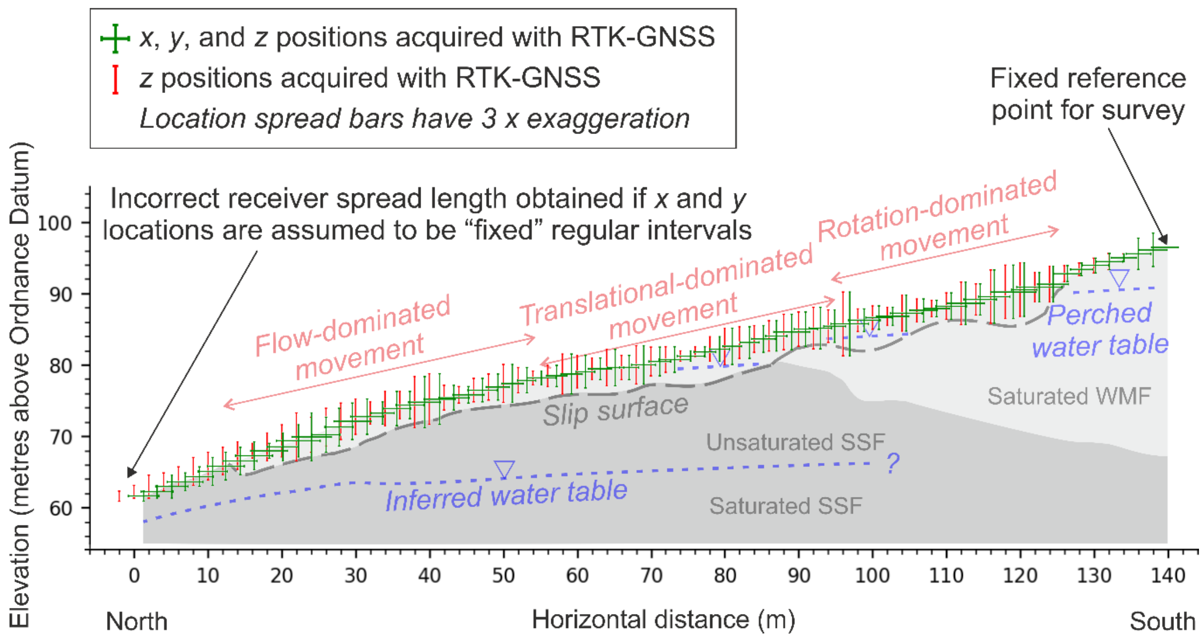

Figure 5: The positions of receivers used in the SRT surveys at HHLO superimposed on to the site conceptual model, and their variation over the monitoring period; the bars indicate the spread of locations over the time-series (with 3 x exaggeration applied). The green points are surveyed positions using an RTK-GNSS system, where Equation 4 has been used to generate true line-of-sight receiver distances. The red points show how errors in positioning can arise if a "fixed" nominal receiver spacing is assumed, resulting in lateral errors in receiver positions, and over-estimation of slope length, which can result in inaccurate inverted seismic velocities.

3.3. Defining appropriate inversion parameters

In this study, 2D inversion of the seismic data is undertaken using the open-source Python based software *pyGIMLi (Rücker et al., 2017)*. This software allows the inclusion of an error model fitted to the distribution of reciprocal errors across the entire time-lapse dataset. A mesh-generation module in *pyGIMLi* produces unique meshes for each time-step inversion, derived from the RTK-GNSS measurements (see section 3.2). The production of unique meshes for each time-step increases the accuracy of the model for each time-step, but presents issues for later time-series analysis; in an ideal monitoring campaign, the inversion meshes for each of the survey time-steps would be identical, allowing for comparison of inverted velocity models on a cell-by-cell basis. However, given the overriding importance of capturing the differences in receiver positions and topography between time-steps, the use of unique meshes is necessary, and this issue is addressed after the data inversion stage.

For this study, a two-stage ‘reference model’ inversion approach is used to constrain the inversion and minimize differences between time-steps (Figure 3). In the first stage, stand-alone inversions of all of the individual time-steps are undertaken, with the aim of identifying the single time-step with the ‘best’ inverted model in terms of data fit. For this first-stage, it is necessary to define a generic velocity gradient model (i.e., velocity increasing with depth) as the starting model for the inversion process. In the first instance, we test the effects of changing different parameters of the starting velocity gradient model by changing these parameters and performing repeat inversions. The parameters we test include the velocity bounds of the starting gradient model (i.e., velocity at the surface and base of the model) and the maximum depth of the starting model. From this process, we obtain a range of inverted models, and assess which starting velocity gradient model parameters provide good results both in terms of data-fit (by considering RMS and χ^2 values) but by also giving consideration to our understanding of the subsurface based on the site conceptual model (Figure 1). There is a risk that the inversion process will introduce artefacts to improve data-fitting, and so qualitative comparison of results with *a priori* site information is as equally important at this stage as considering quantitative metrics of data-fitting. For this study, we are able to rely on several sources of previous intrusive and geophysical data against which to validate the inverted velocity model (see Chambers et al., 2011, Merritt et al., 2013, Uhlemann et al., 2016, Uhlemann et al., 2017). Stand-alone inversions of each time-step are performed using the parameters for the starting velocity gradient model shown in Table 2.

Inversion settings							
Inversi	**Dept**	**Minim**	**Maxim**	**Smoot**	**Maxim**	**Absol**	**Relat**

on parameter	h of mesh	um velocity at surface	um velocity at base	hing factor (lambda)	um travel time	ute data error	ive data error
P-wave inversion input value	40 m	300 m/s	3000 m/s	25	86 ms	0.0242 ms	0.02 %
S-wave inversion input value	4 0m	100	1500	25	178 ms	0.0194 ms	0.006 %

Table 2: The optimal parameters for the staring velocity gradient model used in the first-stage inversion. These were obtained by changing their values and observing their effects on the inverted model output in terms of both data fit and comparison with the site conceptual model. Parameter values resulting in the 'best-fit- model were then used to invert all of the time steps, in order to identify the 'best-fit' model used for the second-stage of inversion.

RMS and χ^2values are calculated for each inverted time-step model. The 'best-fit' model is assessed by looking at the divergence of χ^2 from a 'perfect-fit' model, in which $\chi^2 = 1$. The model with the lowest absolute divergence (i.e., closest to $\chi^2 = 1$) is designated as the 'reference model' for the second inversion stage. As no modelling error is included in the inversion, inversions do not typically converge at $\chi^2 = 1$. Details of the values of χ^2 and divergence from χ^2 for each inversion are shown in (see Table 3, in Appendix). From here, a single model is identified as having the 'best' fit and is taken forward to the second stage; all other resulting inverted models computed up to this stage are discarded.

In the second stage of the inversion process, the inversion of the entire data set is then repeated, but this time using the 'reference model' (i.e., the 'best' fit model) from stage one as the starting model. Using this method provides all of the time-steps a realistic and common starting model that is appropriately constrained and represents the local subsurface seismic properties. Within this second stage of the inversion, a nearest-neighbour lookup function allows the starting model to be sampled from the 'reference model' mesh to the inversion mesh, overcoming issues of topographic variance encountered by using different topography for mesh generation at each time-step. The result is a time-series of inverted seismic velocity models, all inverted using a common, real-world derived starting model.

As a result of incorporating unique topography for each time-step, each time-step uses a different mesh structure. To allow for consistent analysis of inverted velocity models between time-steps, the models are re-sampled and interpolated to a regular, refined, triangular mesh (constructed using the same *pyGIMLi* module), effectively creating a spatially-identical time-series on a consistent mesh (Figure 7a). We use a mesh generated with the most recent topography in the monitoring campaign, in order to better reflect an up to date state of the system. One consequence of this approach is that some cells from earlier surveys, in which the surface positions may now have slipped downslope are not sampled to the resampling mesh. To mitigate against this effect, we use a refined cell size that is smaller than the original cells used for the inversion, purposefully oversampling the inverted data in order to discretize the subsurface, and capture variations in the very near-surface. This enables a range of analyses of the time-lapse dataset (see Figure 3; Data analysis and integration).

4. Topographic induced variations in seismic velocity

In section 3.2, we emphasise the importance of accurately capturing the intra-survey topography by using 2D line-of-sight distances from 3D GNSS surveys, and using topography data acquired for each individual survey in the monitoring campaign. This short section serves to demonstrate how failing to accurately capture variations in topography can have a significant impact on final inverted V_p and V_s measurements.

To demonstrate the effect of temporal topographic variation on seismic velocity, the first 14 V_p datasets (D_0:D_{13}) and accompanying topographic surveys (T_0:T_{13}) are processed according to the workflow in Figure 3, and the text in Sections 3.1 to 3.3. A P-wave travel-time dataset from midway through the monitoring campaign, January 2018 (D_8), is processed and inverted using the surveyed topography (T_8) to produce a 'true' time-step dataset (TTS_8) comprising 2128 subsurface V_p model points. The same seismic dataset (D_8) is then processed using the remaining topographic data in the time-series (T_0:T_7, T_9;T_{13}), resulting in 13 SRT time- steps with 'false' topography (FTS_0:FTS_7, FTS_9:FTS_{13}). The variations present in these 'false' time-steps represent the effect that real-world variations in topography across the monitoring period have on seismic velocity. By normalising all of the time-step data to TTS_8, the results from January 2018 become a baseline against which variations in seismic velocity arising from subtle, but realistic changes in landslide topography are assessed. The result of this analysis is shown in Figure 6. They indicate that topographic variations can have a large impact on the resulting V_p, with 23% of the total data showing velocities greater than ±10% of the true maximum recorded velocity. This has significant implications when trying to identify variations arising from genuine subsurface elastic property changes caused by environmental factors, as these variations can be very subtle (see Section 5).

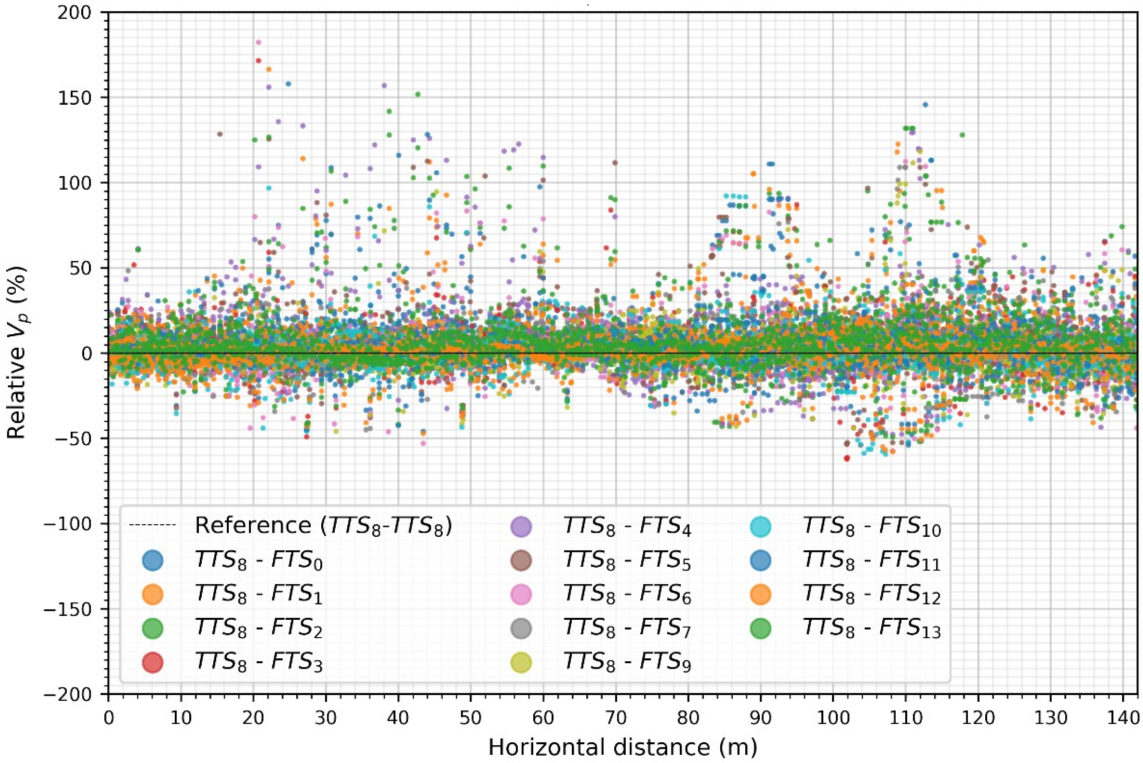

Figure 6: Relative changes in V_p caused by subtle, real-world changes in topography. The solid black line at y=0 represents a normalised baseline (TTS_8-TTS_8). The same seismic dataset (D_8) has been processed using the other time-step topographic data; any variations in V_p are therefore a product of these subtle topographic changes between surveys.

5. Data analysis and results

One approach to analysing time-series SRT data is to look at how the seismic attributes of discrete seismic units respond to changing environmental conditions. The prevalent subsurface lithological discontinuities (i.e., those that are stable in time) are highlighted by plotting the mean values of the individual cells across the 33 month monitoring period (Figure 7). These plots are displayed using the most recent topography in the time-series. The individual cross-sections highlight significant subsurface features, including changes in lithology at depth, and different domains of movement in the near surface. Plots showing the standard deviation of these mean values (Figure 7) indicate the areas of the landslide that show greatest velocity variation across the monitoring period.

Here we concentrate on the sliding layer at the HHLO (extending from the surface to 2 – 4 m depth), which is easily identified by the low V_p and V_s at the surface of the cross-sections. This extends from beneath the break in slope at the bottom of the back scarp (~15 m horizontal distance), to the base of the flow lobes (125 m horizontal distance). At HHLO this surface sliding layer is monitored by several subsurface and surface environmental sensors, recording rainfall and changes in moisture content, allowing direct comparison with inverted cross sections. By selecting grid cells within this layer, it is possible to calculate the change in velocity over time. In our case, the surface layer comprises model cells from both the V_p and V_s time-series datasets (Figure 7b and Figure 7d), the positions of which are

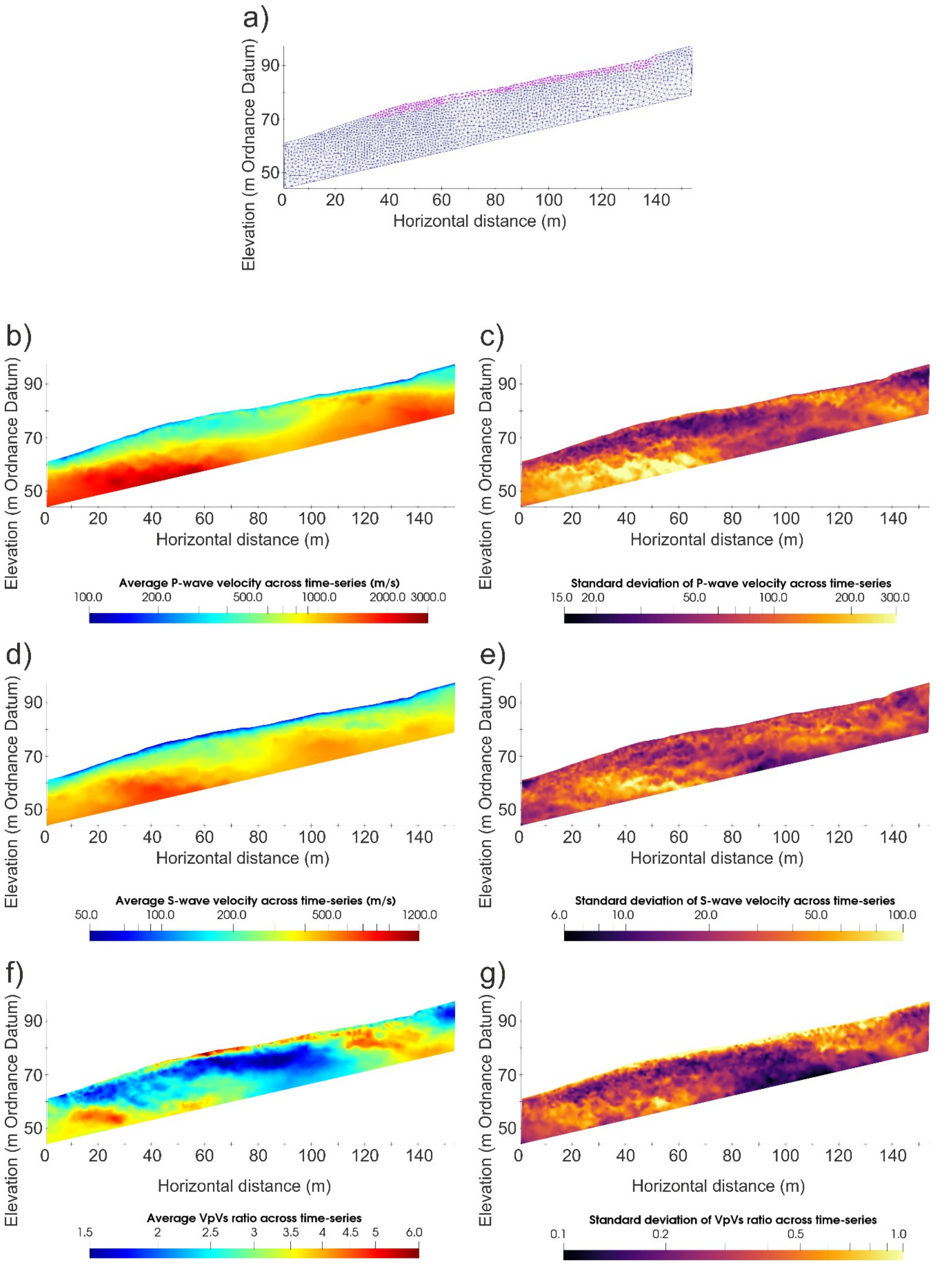

Figure 7: a) The regular mesh used to sample all of the individual time-steps to create spatially comparable datasets for the time-series. The cells highlighted purple in the surface sliding have been used for the analysis in Figure 8. b) Cross-section showing average V_p across the time-series. The value shown in each cell (see Figure 7a) is the mean velocity value from the entire 33 month monitoring period. c) Cross-section showing average V_s across the time-series. The value shown in each cell (see Figure 7a) is the mean velocity value from the entire 33 month monitoring period. d)

Cross-section showing average V_p /V_s across the time-series. The value shown in each cell (see Figure 7a) is the mean ratio value from the entire 33 month monitoring period.

fixed through the use of a common fixed mesh. Figure 8 shows the time-series V_p and V_s inverted values extracted from this surface layer, alongside calculated effective rainfall, soil moisture data from a cosmic-ray sensor measuring shallow (~0.1 m bgl) moisture content across the site. V_p increases and decreases in relation to soil moisture, but with a slight time lag. The lag effect is caused by the difference of the sampling depth of the moisture sensor (<0.1 m bgl) and the depth of the V_p readings (2 – 4 m bgl); the moisture content of the HHLO near-surface changes more quickly in relation to net infiltration and evapotranspiration rates (shown by the hourly soil moisture, faint green line) than the top 2 – 4 m of the landslide, which will be less subject to evapotranspiration processes at depth. It is also worth noting that inverted velocities will be smoothed values from the true velocities, due to the spatial and temporal smoothness constraints used.

Furthermore, the calculated V_p/V_s ratio (Figure 7f), which is an indicator of material saturation (Uyanık, 2011), better reflects changes in moisture content in Figure 8. Crucially, the minimum V_p (350 m/s) in the time-series is 24% less than the maximum V_p (462 m/s). Given that topographic effects alone can cause variations in V_p >±10% (Figure 6), the changes in seismic velocity over time could easily be masked if the data are not processed correctly. This demonstrates the necessity for including accurate topography in long-term SRT monitoring campaigns in landslide settings.

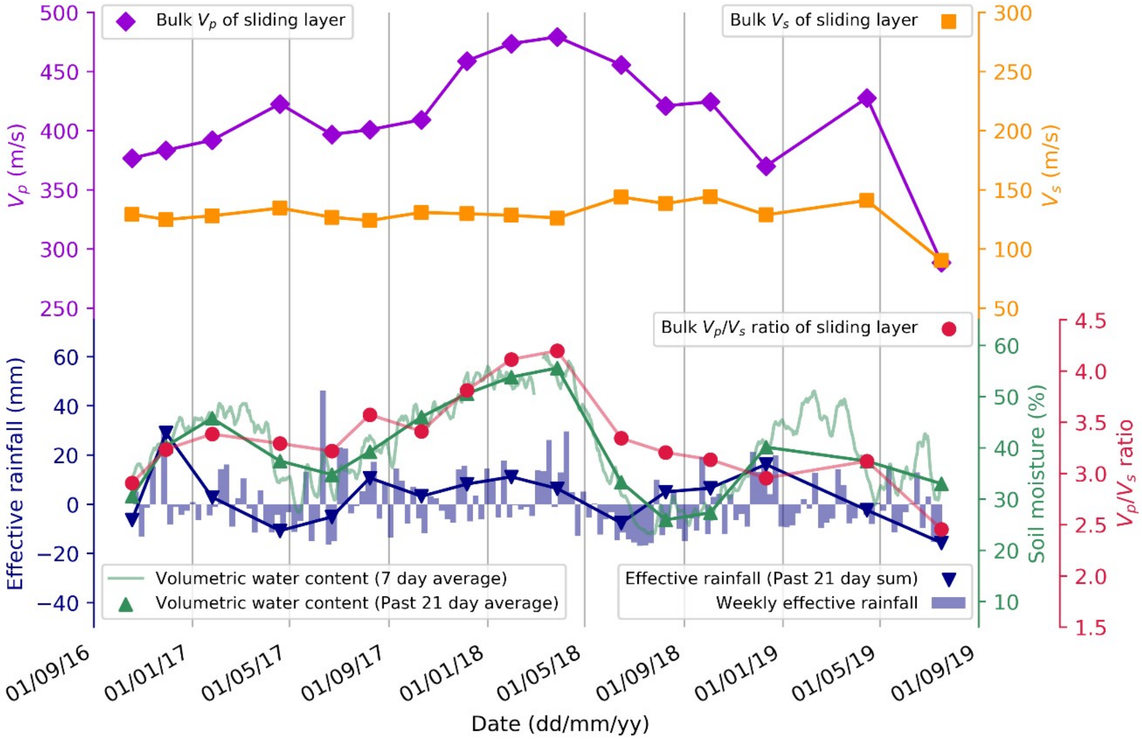

Figure 8: The top panel shows variation in bulk V_p and V_s *readings from the sliding layer at the HHLO (see Figure 7 for location of this layer). The shaded areas are the 1st and 3rd quartiles of the range in* V_p and V_s. *The* V_p/V_s *s ratio, derived from the bulk* V_p and V_s *readings is shown. In the bottom panel, weekly effective rainfall, showing periods of net infiltration/evapotranspiration at the HHLO, and soil moisture from a surface sensor measuring to <0.1m bgl. The variation in* V_p *broadly follows the increases and decreases in moisture content, while* V_s *shows little variation. The derived* V_p/V_s *s ratio shows greatest sensitivity to the moisture content of the surface sliding layer at HHLO.*

6. Conclusions

SRT is rarely used for the long-term assessment of landslides prone to hydrological destabilization, but has great potential for high-resolution spatial monitoring. This is particularly the case in slopes with high spatial heterogeneity in which monitoring data obtained from sparse point observations is insufficient to capture the complexity of the landslide system. Landslide monitoring campaigns using SRT can determine seismic attributes of slipped materials, which provides information on elastic property changes due to temporal variations in moisture content. However, failing to give due attention to the possible sources of error in SRT surveys can lead to artefacts in the time-lapse data, which can easily mask changes arising from genuine variations in the elastic properties of landslide materials, including the underlying rock. In this study, we provide a workflow for addressing the errors associated with producing a reliable time-series of inverted seismic velocity models, and have shown how velocities in the near-surface soil layers are sensitive to variations in moisture content.

Standard approaches to quality assessment and processing of SRT data aid in minimizing individual survey data error. The use of emerging methods to increase picking accuracy, such as automatic detection algorithms, machine learning and statistical approaches will also decrease data errors introduced in to the creation of time-lapse data from standalone surveys. In this study, data from each survey were processed using reciprocal error analysis to ensure e<5% of travel-time for all datasets. However, in the case of producing time-lapse data from these individual datasets, we underscore the importance of using detailed, unique topography data for processing each time-step. This crucial step could easily be overlooked by inaccurate assumptions regarding field setup, receiver spacing landslide surface movement between surveys, even by experienced SRT operators.

For the data considered here, changes in topography lead to >±10% variations in apparent seismic velocities in 23% of the data for the unconsolidated near-surface. Our data exhibits a 24% difference between the fastest and slowest V_p observed in this layer, caused by variations in elastic properties induced by changes in moisture content, underscoring the need to properly account for topography effects. To avoid the errors associated with changes in topography, accurate source-receiver positions are important when processing SRT monitoring data. Several other steps, including the repositioning of receivers in the field, the use of data quality indicators (such as travel-time reciprocity) and robust reference models for inversion further reduces these errors. If these potential sources of error are managed correctly, SRT presents a useful tool for the identification of heterogeneous subsurface conditions and their changing properties over time in active landslide settings.

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Appendix

P-wave inversions																
Starting model	Velocity gradient model (see Table 2 for values)															
Time-step	**0**	**1**	**2**	**3**	**4**	**5**	**6**	**7**	**8**	**9**	**10**	**11**	**12**	**13**	**14**	**15**
χ^2	0.982	1.495	1.952	1.345	1.478	1.138	1.434	1.358	1.848	1.573	1.288	1.316	1.258	0.982	1.157	0.992
RMS	2.970	3.788	4.971	4.310	4.143	4.412	4.721	5.244	5.713	5.431	4.010	3.965	3.727	3.404	4.157	3.492
χ^2 divergence	0.018	0.495	0.952	0.345	0.478	0.138	0.434	0.358	0.848	0.573	0.288	0.316	0.258	0.018	0.157	0.008
Starting model	Reference model derived from results of first stage inversion (i.e., time-step 15)															
Time-step	**0**	**1**	**2**	**3**	**4**	**5**	**6**	**7**	**8**	**9**	**10**	**11**	**12**	**13**	**14**	**15**
χ^2	0.984	1.516	1.904	1.486	1.593	1.440	1.733	1.445	2.238	1.893	1.442	1.351	1.357	1.161	1.330	0.927
RMS	3.157	3.935	4.860	4.466	4.069	5.044	4.729	5.202	5.693	5.515	4.314	3.966	4.050	3.987	4.148	3.385
χ^2 divergence	0.016	0.516	0.904	0.486	0.593	0.440	0.733	0.445	1.238	0.893	0.442	0.351	0.357	0.161	0.330	0.073
S-wave inversions																
Starting model	Velocity gradient model (see Table 2 for values)															
Time-step	**0**	**1**	**2**	**3**	**4**	**5**	**6**	**7**	**8**	**9**	**10**	**11**	**12**	**13**	**14**	**15**
χ^2	1.117	1.416	2.061	1.959	2.654	1.991	2.499	1.563	1.771	3.822	1.614	2.873	2.517	1.910	2.179	2.100
RMS	1.206	1.415	1.764	1.678	2.001	1.685	1.910	1.491	1.704	1.974	1.716	2.068	1.933	1.629	1.770	1.750
χ^2 divergence	0.117	0.416	1.061	0.959	1.654	0.991	1.499	0.563	0.771	2.822	0.614	1.873	1.517	0.910	1.179	1.100
Starting model	Reference model derived from results of first stage inversion (i.e., time-step 0)															
Time-step	**0**	**1**	**2**	**3**	**4**	**5**	**6**	**7**	**8**	**9**	**10**	**11**	**12**	**13**	**14**	**15**
χ^2	0.990	1.520	2.082	1.988	2.629	1.912	2.243	1.713	1.681	1.728	1.812	2.861	2.533	1.913	2.047	2.039
RMS	1.124	1.479	1.657	1.745	2.018	1.532	1.820	1.731	1.724	1.716	2.349	2.106	2.074	1.612	1.889	1.675
χ^2 divergence	0.010	0.520	1.082	0.988	1.629	0.912	1.243	0.713	0.681	0.728	0.812	1.861	1.533	0.913	1.047	1.039

Table 3: The results of the two-stage inversion process for both the Vp *and* Vs *surveys. In stage one, a velocity gradient model with the parameters in Table 2 is used to perform stand-alone inversions of each time-step. The ‘best’ result (highlighted green) is then assessed by looking at divergence from a perfect model fit (i.e., a normalised χ^2 value, or ‘χ^2 divergence’). The ‘best’ model is then used as a ‘reference model’ for the second stage inversion. The ‘reference model’ is used for the inversion of each time-step in the second stage inversion, providing a real-world, common starting model for the time-series.*