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Combining seismic noise techniques for landslide characterization

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Abstract

The strong topographic relief and the presence of weakly consolidated sediments create favorable conditions for the development of landslides around the eastern rim of the Fergana Basin (Central Asia). In summer 2012, a field experiment employing small aperture seismic arrays was carried out on an unstable slope, using ambient vibration recordings. The aim of the study was to constrain the seismic response of a potential future landslide and to map lateral and vertical changes in the shear-wave velocity of the surficial soil layers. Strong variations of horizontal-to-vertical spectral ratios in terms of amplitude and directionality indicate clear differences in local site effects, probably reflecting the stability of different sections of the slope. Results further show resonant frequencies of both the entire unstable block as well as for smaller individual parts. The use of an ad-hoc passive seismic tomography approach based on noise correlograms allows the mapping of the shear-wave velocities of the sliding material, even in cases of significant topography relief. Based on the recording of seismic noise only, we clearly identify a low-velocity body of weakly consolidated claystone and limestone material, which can be interpreted as the landslide body, with laterally varying thickness.

Abbreviated title: Seismic noise for landslide characterization

Key words: seismic noise, landslide, site effects, tomography

1. Introduction

Individual slope failures on landslides are generally not as spectacular or as costly as the consequences of earthquakes, major floods or other natural catastrophes. However, landslides are one of the most widespread hazards and cause thousands of deaths and billions of dollars in damage worldwide each year (Dilley et al. 2005). The global death toll from landslides increased from around 600 per year in the early 1970s (Brabb 1991) to more than 340 per month on average nowadays (Petley 2009). Generally speaking, the number of casualties due to slope failures is higher in developing countries, whereas the industrialized world can be adversely affected by severe economic losses. The increase of both parameters is due to the higher value of endangered structures and, the greater number of people living in hazardous areas, coupled with the possible increase of extreme meteorological events (Schuster and Fleming 1986).

Developing countries always have had difficulties affording the high costs involved in mitigating the effects of natural hazards through major engineering works and rational land-use planning. In addition, owing to the economic recession, many industrialized societies are also reluctant to invest in structural measures to reduce natural risks. Hence, the new procedure is to implement monitoring and warning systems with the aim of minimizing the loss of lives and property without investing in long-term, costly projects of ground stabilization. Therefore, obtaining comprehensive knowledge of unstable slope structures (e.g., detection and inventory of unstable areas, displacement monitoring) is essential for any quantitative hazard assessment and for the design of effective remediation strategies. In other words, defining the 3D geometry of a potential slope failure, its geomechanical and hydrological properties, and its internal discontinuities at a high resolution is highly desirable, but still poses a major economic and scientific challenge (Travelletti and Malet 2012).

Since landslides affect all geological materials and exhibit a large variety of shapes and volumes, the characterization of these phenomena is not a straightforward problem and may require extensive investigation. There are a significant number of methods, which mainly include remote-sensing and aerial techniques, geological and geomorphological mapping as well as geophysical and geotechnical techniques, which have been adapted to characterize landslides (see the pioneering work of Bogoslovsky and Ogilvy (1977) as well as the reviews by Fell et al. (2000) and Jongmans and Garambois (2007)). In particular, the recent emergence of 2D and 3D geophysical imaging techniques has dramatically increased their attractiveness for landslide applications, having the major advantage of giving continuous information about the studied body. Although a clear relationship between geophysical parameters and geological/geotechnical properties is often still unclear, these methods now appear as major tools for investigating and monitoring unstable slope masses. Therefore, for characterizing intrinsic properties like geological layering, material stiffness and material porosity, several different parameters like electrical resistivity and seismic body-wave velocities are of great interest (Grandjean 2006, Jongmans et al. 2009).

The latter parameter might be obtained by active seismic tomography, which consists of inverting first-arrival times to obtain an image of the seismic velocity distribution in the ground and to image a velocity contrast between the unstable mass and the stable layers (Caris and van Asch 1991, Grandjean 2006, Jongmans and Garambois 2007). However, compared to classical seismic refraction, such active methods require much more travel-time data and greater field efforts. Moreover, for the case of a landslide, the results might be difficult or even impossible to be interpreted due to disturbed surfaces and the significant topographic relief.

Passive methods based on the recordings of ambient seismic noise can serve as an efficient alternative. This technique, generally low cost and carried out with instruments easy to deploy, is not bound to any active sources and has become increasingly attractive in a broad

range of seismological disciplines on different scales (Bonnefoy-Claudet et al. 2006a). Regarding the analysis of ambient vibrations on potentially unstable slopes, several studies have been published in the last years. Havenith et al. (2002) and Méric et al. (2007) estimated the resonant frequencies of a rock slope instability using single-station H/V measurements. Danneels et al. (2008) used small seismic arrays to measure the dispersion curve of surface waves and, subsequently, inverted for two 1D shear-wave velocity profiles in the active area. Burjanek et al. (2010) estimated amplification factors and identified the range of resonant frequencies for a large unstable rock mass by noise polarization analysis. Burjanek et al. (2012) used small seismic arrays to identify normal mode motion (standing waves) within an unstable rock mass. Renalier et al. (2010) correlated seismic noise to resolve 2D profiles of the shear-wave velocity structure of a landslide composed of clayey soil deposits. This correlation plays a significant role in the landslide self-excitation process, since the triggering mechanism – besides ground shaking – most often proposed is the partial or total liquefaction of the mass, resulting from an increase in pore water pressure (Picarelli et al. 2005), meaning that the bulk S-wave velocity should decrease dramatically in the vicinity of liquefied zones and also with the extent of damaged (i.e., cracked) subsurface material (Renalier et al. 2010). Finally, Jongmans et al. (2009) showed a clear correlation between surficial displacement rates (related to clay deformation) and S-wave velocity changes (obtained by active seismic refraction) on a landslide in the French Alps.

All these studies suggest that S-wave velocity could be a valuable parameter for the characterization of a landslide body and that it would constitute a well suited alternative to classical prediction methods obtained for slope degradation monitoring. Here, we show the potential of purely seismic noise data by the joint use of both single station noise measurements and a 3D tomography approach. These methods can provide complementary information about the seismic response of an unstable landslide, and allow the identification of layers of different material in the landslide body. After a description of the study area and

the setup of the experiment, we will show that directionality analysis of ambient noise can provide robust estimates of the intrinsic resonance properties of the landslide body. Moreover, the complementary correlation of seismic noise signals can be used to obtain a 3D S-wave velocity image of the investigated landslide body.

2. Study region

Central Asia, in particular the countries of Kyrgyzstan and Tajikistan, are one of the areas with the highest landslide hazard in the world (Nadim et al. 2006). The strong relief of the Tien Shan is one of the main causes of the general instability of the mountains which, in combination with the complex interaction of tectonic, geological and hydro-meteorological factors (Havenith et al. 2006), can cause the development of significant gravitational movements along large mountainous slopes. The first documented observations of slope instabilities in this region date to the end of the 19th century (Ignatiev 1886). However, several of these mass movements were neither documented nor did they cause significant damages, since they occurred in remote and sparsely inhabited areas (Strom and Korup 2006).

The 300 km long and up to 110 km wide Fergana valley, covering territory in Kyrgyzstan, Tajikistan and Uzbekistan, represents an exception. More than ten million people, i.e., 20 % of the population of Central Asia, turn this region into an exceptionally densely populated area. Relatively high levels of precipitations and the presence of Mesozoic-Cenozoic rocks in combination with Quaternary sediments create favorable conditions for landslides, which are mainly concentrated along the eastern rim of the Fergana valley (Roessner et al. 2005). Most of the landslides are mobilized during the rainy season, between fall and spring, causing serious damage to settlements and loss of life. The last big

documented landslide in this region occurred in 1994 in the north-western part of the Fergana valley, killing 115 people and displacing another 27,000 (Moldobekov et al. 1997).

In the south-eastern part of the Fergana valley, the Osh Oblast forms the transitional zone between the Fergana Basin in the north and the Alai Mountain Range in the south, with elevations ranging from 1000 m for the city of Osh up to more than 3000 m in the south. The region is composed of Paleozoic and Cretaceous rocks as well as weakly consolidated clays and limestones of the Neogene in the synclines (Figure 1), while the river valleys are filled with Quaternary deposits (Tursungaziev et al. 2008). The entire area is characterized by a moderate to high seismic hazard (Abdrakhmatov et al. 2003, Bindi et al. 2012).

The studied landslide forms the eastern edge of the village of Papan. The village has around 4000 inhabitants, mainly settled along the Ak-Bur river, which runs from south to north and bisects the city. The entire landslide is around 360 m long, 300 m wide and covers an elevation range of 140 m and is mainly composed of a number of cretaceous carbonate rocks, limestones and claystones. In addition, it is accompanied by smaller micro-landslides. Due to its remote location and sudden occurrence, the Papan landslide is poorly documented. The last larger failures which formed today's configuration of the area occurred in 1994 and on 16th of December 2004. Recently, between May 2009 and September 2010, a small portion (about 60,000 m²) in the northern part of the landslide slid (U. Wetzel and D. Golovko, pers. comm.). However, an effective landslide characterization is only possible with a clear understanding of its geometry and dynamics, hence our aim is to detect material changes where failure can occur.

3. Field experiments

In September 2012, we performed ambient noise measurements on the Papan landslide using a set of 17 wireless sensing units. Each sensing unit is composed of a digitizer board with an effective resolution of 19-bit, onboard storage capability provided by a compact flash card and wireless data transmission capabilities that can be used to assemble a large number of sensing units into a self-organizing network sending the recorded data to a central data center. In particular, if the transmission of the entire recorded data is not required, each sensing unit can also perform a predefined set of calculations on the recorded data and only transmit the calculated parameters to the central acquisition center (Fleming et al. 2009, Picozzi et al. 2010).

Each sensing unit recorded seismic noise at 100 samples per second and was connected to an external PE-6/B 3-component 4.5 Hz geophone, providing an excellent signal-to-noise ratio for frequencies down to ~ 0.9 Hz and allowing appropriate investigations with a general tendency to be valid even down to 0.6 Hz (Strollo et al. 2008). The precise positions of the sensors were measured using differential GPS. We deployed two arrays on consecutive days with varying geometries and apertures up to 90 m within the unstable soil mass (shaded area in Figure 2). Altitude differences within each array were around 30 m. In each case, a reference station was put outside the unstable part of the landslide but still located on soft soil. On each day the recording duration was around three hours. However, the recordings of one station of the first array were unusable due to a connection malfunction.

4. Horizontal-to-vertical spectral ratios and directivity analysis of noise

The H/V spectral ratio, i.e., the ratio between the Fourier spectra of the horizontal and the vertical component of microtremor recordings, was first introduced by Nogoshi and Igarashi (1970, 1971) and was widespread by the publication of Nakamura (1989). Several authors have pointed out the correlation between the H/V peak frequency and the fundamental resonance frequency of a site, and it is widely accepted to use the H/V technique as an indicator of the subsurface structure (e.g., Parolai et al. 2002, Bonnefoy-Claudet et al. 2006b). Although theoretical considerations support the use of the H/V spectral ratio for site effect evaluation only for the case of 1D layering dynamic response, this technique has also been found suitable to obtain basic information about the site specific structure in more complex situations, e.g., in the presence of topographic amplification (Chávez-García et al. 1996).

To study the landslide by means of the H/V method, raw velocity recordings for each site were divided into 60 s long windows and base-line corrected. Each window was cosine tapered (5 per cent) and a Fast Fourier Transformation (FFT) for each component was performed. Spectral amplitudes were smoothed using the Konno and Ohmachi (1998) recording window, fixing the parameter b that determines the degree of smoothing to 40. To study any directional resonance, the horizontal recordings for each measurement site were progressively rotated by an azimuthal step of 5° , and the spectral ratios calculated for each pair of horizontal components, based on Spudich et al. (1996).

Figure 3 presents polar diagrams of the mean H/V spectral ratios for the different array locations. Obviously, the observed directionality pattern of ambient seismic noise is rather complex, showing strong variations between neighboring sites. In common for most of the sites is that for the lowest frequencies (0.5 to ~ 1 Hz) noise polarization follows an east-west trend with a small maxima around 0.8 Hz. Based on field observations, we propose that such a directional resonance might be caused by the resonance of the soft Quaternary material of the

Ak-Bur river valley. In the studied area, the river valley is from several hundred meters up to 1 km wide, and is oriented almost exactly in a north-south direction. We assume that the noise wavefield is dominated by standing waves perpendicular to the valley axis causing a single narrow low-frequency amplification in the H/V ratio which might identify the SV fundamental. Such peaks are found from almost all stations on the landslide body as well as outside of it. Although one might expect a distinct vertical component for the SV modes other studies describing similar directional low-frequency amplification in river valleys (e.g., Steimen et al. 2003, Roten et al. 2006) indicate that peaks of the fundamental mode of SV resonance might not be visible in the vertical component. Due to the apparent heterogeneous structure in elongated sediment-filled river valleys, these patterns cannot be created by horizontally propagating surface waves and cannot be interpreted classically in terms of 1D resonance, but may be explained with a wave front propagating almost vertically at a slight angle. This is the case for standing waves expected in the case of 2D resonance.

The narrow H/V peaks at intermediate frequencies (~2 to 2.5 Hz) show significant directionality with a clear east-west trend for almost all the sites (see Figure 3) which is in excellent agreement with the maximum slope direction of the entire landslide body. We assume that here the ambient noise shows normal mode vibration for the unstable portion of the slope, i.e. it becomes polarized in one direction at one single resonance frequency for all stations on the landslide. Following Del Gaudio and Wasowski (2007) and Burjanek et al. (2010), due to the peaks in the axial Fourier spectra (e.g., S4 and S11 in Figure 4), it is likely that the entire unstable landslide mass, being around 360 m long and 300 m wide, might be viewed as a volume of effectively reduced elastic moduli, analogous to the sediment infill in valleys. Consequently, this mass vibrates in this direction of freedom with a normal mode resonance frequency of around 2.5 Hz, and it is likely that ground motion during a strong event will be amplified in this direction.

Locally enhanced or reduced amplitudes of the H/V spectral ratios as seen in Figure 3 might reflect lateral variations in material properties of the landslide body which might represent differences between slightly more stable and more unstable parts of the slope, although the transition from the stable to the unstable part is rather smooth, meaning that a clear boundary of the unstable part can hardly be mapped (Burjanek et al. 2010). For example, the H/V amplitudes in the northern part of array 2 are rather high for all measurement sites (see e.g., S3 in Figure 3) and we assume that this area might represent an extended unstable area (cf. photo B in Figure 2). This is in agreement with Bourdeau et al. (2004), who proposed a correlation between the lateral extent of a zone of large amplifications inside the slope and the location and the area of the sliding mass (i.e. this can be seen as a surface area approximation of the 3D volume of the landslide). Accordingly, stations installed at the edge or outside of the landslide body show only a weak polarization and rather low H/V amplitude at this frequency, while the horizontal Fourier spectra do not show any clear peak (S7 and S19 in Figures 3 and 4).

The high-frequency directionality (between 10 and 30 Hz) in Figure 3 is more complex to interpret. Generally, the amplification is not fixed to one common frequency and one single direction for all sites, but changes from site to site, with large lateral variations observed over short distances. Therefore, we assume that the high-frequency motion might be controlled by the vibration of smaller portions of the landslide and very shallow heterogeneities with different resonant frequencies and different directions of freedom. For example, this explanation is in line with the direction of maximum resonance at 20 Hz for station S11, which is approximately perpendicular to the trace of a nearby surface crack oriented in a north-south direction. Hence, this high amplification might be related to an area of maximum local instability. However, a more thorough interpretation of this complexity requires longer and denser measurements.

5. Ambient noise imaging

Single station H/V measurements cannot escape the problem posed by the inherent trade-off between S-wave velocity and thickness of the soil layers in inversion schemes. By contrast, geophysical imaging techniques based on ambient noise recorded simultaneously on a dense array can provide supplementary information, and can therefore serve as an efficient alternative. Traditional methods like the SPatial AutoCorrelation (SPAC, Aki 1957) method retrieve surface wave dispersion curves, which, in turn, can be inverted to obtain the S-wave velocity structure, under the assumption that the underlying medium is not varying laterally. In this way, the wavelength of surface waves has to be small enough not to break this assumption, leading to a limited penetration depth and limited lateral extent. However, in complex environments like a landslide with a distinct topographic relief, large variations in material properties are expected both laterally and vertically, hence classical methods like SPAC will provide biased results.

On the other hand, one can expect that local heterogeneities will affect the noise propagation along the ray path between the sensors, and hence can be retrieved by analyzing the Green's function estimated by the cross-correlation of the signals recorded at two different stations. Since the pioneering work of Aki and Lee (1976), this idea has been applied over many scales and for different purposes. However, using this approach for all inter-station paths generally requires the solution of a large inverse problem to obtain a heterogeneous velocity model that is consistent with observations. Here we make use of a new procedure for the passive imaging of near-surface properties by a weighted 3D inversion procedure that also works in case of complex 3D S-wave velocity structures under pronounced topographic conditions (Pilz et al. 2012, 2013). In particular, in contrast to previous landslide imaging studies (e.g., Renalier et al. 2010, Mainsant et al. 2012), we include the effects of topography since the travel times could potentially be biased where the relief is not small compared to the wavelength and

penetration depth of the considered waves. The applied method, based on surface waves only, which can be easily extracted because they dominate the Green's function between receivers located at the surface, can offer a more comprehensive view of the subsurface, having an investigation capacity on greater areal extent and greater depth than traditional methods, such as trenching or drilling.

Before deducing a velocity model, since this method is based on anthropogenic noise with frequencies higher than a few Hertz, a frequency-wavenumber (f - k) analysis is carried out using the maximum likelihood method (Capon et al. 1967) for checking the isotropy of the wave field for the frequency range of interest ($4 \leq f \leq 13$ Hz). Figure 5 shows the distribution of noise sources for array 2 for two different frequencies, 4 and 13 Hz. At 4 Hz, the sources are mainly distributed to the north and south-east, whereas for frequencies higher than 10 Hz, the sources are more scattered over the wave number plane, with much energy coming from the western direction, i.e., the village of Papan. Although a non-uniform distribution of seismic noise sources might bias the reconstruction of the attenuation of the studied media, it will have only a very minor influence on the phase, i.e., on the calculation of the travel times, provided sufficiently long recording time is available in which enough energy is coming from the remaining azimuthal sectors (Derode et al. 2003, Larose et al. 2007, Yang and Ritzwoller 2008, Yao and van der Hilst 2009).

Therefore, following Ohori et al. (2002), for a quasi stochastic and stationary microtremor wave field, the azimuthally averaged correlation function for each angular frequency ω_0 can be expressed in the form of a Bessel function $J_0\left(\frac{\omega_0}{c(\omega_0)}r\right)$ where $c(\omega_0)$ is the frequency-dependent phase velocity and r is the interstation distance. Considering that experimental values of the azimuthally averaged spatial correlation function can be obtained from seismic noise measurements, by fitting these latter values to the theoretical Bessel function values, the frequency-dependent phase velocity can be retrieved.

For calculating the cross-correlation functions between all stations for our study, we consider only the vertical component of the recordings, which are strongly dominated by Rayleigh waves. For each measurement site, the continuous data sets were divided into 240 windows of 30 s. After removing the linear trend from each window, a 5 per cent cosine taper was applied at both ends. A second order Butterworth high-pass filter was applied to these data previously corrected for the instrumental response using a corner frequency of 1 Hz, mainly to improve both the signal-to-noise ratio for the high-frequency range of interest, and also to ameliorate the cross-correlations by oppressing the influence of very low-frequency signals in the cross-correlation functions (Bensen et al. 2007). For each of the chosen frequencies over a range $4 \text{ Hz} \leq f \leq 13 \text{ Hz}$ at intervals of 1 Hz and for all $n(n-1)/2$ possible station pairs, where $n = 16$ is the number of stations, the data were one-bit normalized (Campillo and Paul 2003) and, finally, stacked together which, on average, improves the signal-to-noise ratio and reduces the contribution of the most energetic arrivals. Finally, using the known distances between the sensors, the travel times for all ray paths between the stations were calculated from the estimated phase velocities.

Using 5 different frequency values (4, 6, 8, 10, 13 Hz) the travel times were then inverted by a rapid inversion algorithm to calculate a 3D S-wave velocity model for the respective subsurface below each of the two arrays. Since the S-wave velocity is the dominant parameter influencing changes in Rayleigh-wave phase velocity in the high-frequency range, we perform the inversion considering the observed velocities in the 3D frequency-dependent tomography as being representative of Rayleigh wave phase velocities. Although the ray path is velocity dependent, meaning that travel time inversion is a non-linear problem, deviations of the paths from a straight line will be less than a quarter of the wavelength. Hence, keeping in mind that there is also a small level of error for the input data for shallow seismic surveys, we are sure that a bias of a few percent can be tolerated to keep the solution linear.

Consequently, the medium of interest can be subdivided into a reasonable number of smaller cells, meaning that the problem can be expressed in a simple discrete matrix form to be used in practical applications

$$\mathbf{t} = \mathbf{L}_1 \mathbf{s}. \quad (1)$$

$\mathbf{L}_1$ is an $RF \times MN$ matrix with R being the number of rays for F different frequencies crossing the studied medium that is subdivided into M cells in each of the N horizontal layers (for details see Pilz et al. 2012, 2013). The vector $\mathbf{t}$ corresponds to the travel times between each pair of receivers with a frequency dependent slowness $\mathbf{s}$ (inverse velocity). Starting from a homogeneous 3D velocity model, an iterative procedure using singular value decomposition for minimizing the misfit between the observed and theoretical travel times is adopted. Since shallow blocks are sampled much stronger by higher frequencies where the deeper ones are sampled only by lower frequencies, we introduce a further constraint on the solution by adding a $MNF \times MNF$ weighting matrix which weights the data depending on the number, length, orientation, and vertical penetration depth of each ray path segment crossing each cell. Additionally, to reduce the risk of divergence and to stabilize the iteration process, an adaptive bi-weight estimation (Tukey 1974, Arai and Tokimatsu 2004), which is a type of maximum likelihood estimation, is applied. The final 3D S-wave velocity model is retrieved from the updated slowness $\mathbf{s}$ is by $v_s = (0.92 \mathbf{s})^{-1}$. The S-wave velocity inversion results for the two arrays after 500 iterations in each case are shown in Figure 6.

Using a wide range of frequencies, we can image both shallow and deeper parts simultaneously. Besides the significant topographic variations several prominent features of the S-wave velocity model stand out. In particular, we note that the thickness of the low-velocity body displayed by the thickness of the low-velocity material in Figure 6 varies laterally to some degree. For array 1, the thickness increases towards the north-east and for array 2 it does so from east to west. Remarkably, array 1 shows higher velocities, especially for the lower and flatter parts of the studied area. By contrast, the S-wave velocities for

array 2 are often lower, especially in the north-western part ($v_s < 200$ m/s). In fact, array 1 was located on pebbly limestone material whereas the material below array 2 was mainly composed of clay (see Figure 2). Therefore, consistent with the observations in Figure 3, the area below array 2, in particular the thick north-western parts, is likely to be more unstable, with several significant radial surface cracks also being seen. These features are less obvious for the southern part of array 2 which is characterized by slightly higher S-wave velocities (Figure 6) and smaller H/V amplitudes with smaller directionality (Figure 3).

6. Discussion

The available literature and the presented data sets indicate that the dynamic response and internal structure of potentially unstable slopes can be very complex. There is clear evidence that ground motion on slopes can be considerably amplified with a pronounced directional character. However, strong variations in amplitude of the H/V spectral ratio over short spatial scales, as shown in Figure 3, cannot be explained by topographic effects, i.e., the shape of the topography, alone (e.g. Del Gaudio and Wasowski 2011). Previous numerical simulations have found that topographic amplification factors hardly exceed 2 (Assimaki and Kausel 2007), in particular, pure topographic effects have been found to be considerably smaller than the complex interaction of combined topographic and geologic effects (Geli et al. 1988, Sanchez-Sesma and Campillo 1993, Bourdeau and Havenith 2008). Indeed, a fundamental role seems to be played by shallow heterogeneities and impedance contrasts (see e.g., Moore et al. 2011) in slope material properties. On the other hand, such shallow impedance contrasts are often clearly visible by high-frequency peaks of the H/V spectral ratios (see e.g., stations S11 and S19 in Figure 3 show significant peaks around 25 Hz and clear shallow impedance contrasts in Figure 6).

By contrast, in common for all spectral ratios is a directional amplification at around 2.5 Hz. This specific resonance frequency can be distinguished from both deeper and shallower impedance contrasts (Bozzano et al. 2011). Similar to what is observed across sediment-filled valleys in numerical simulations (e.g., Bard and Bouchon 1985, Frischknecht and Wagner 2004) and in real data (e.g., Tucker and King 1984, King and Tucker 1984, Roten et al. 2006), we expect the resonance frequency for such a normal mode resonance to be approximately equal over the entire extent of the landslide body, regardless of its thickness at each site, alike Burjanek et al. (2012). Using classical 1D transfer function theory and the obtained velocity model (e.g., for S13 and S20, see Figure 6) one might pre-estimate that a thickness of the low-velocity body of around 30 to 35 m will correspond to a fundamental frequency of around 2.5 Hz. Such a maximum thickness seems reasonable in consideration of the shape of the relief. On the contrary, the shallower edges of such landslide bodies undergo very large differential motions and, therefore, the estimation of the landslide's properties using simple relationships like, e.g. the one provided by Haskell (1960), which do not account for the influence of the landslide boundaries and 2D effects, would not yield useful results in such cases.

The strong directionality at this frequency is likely to be caused by the direction of freedom of the landslide body which might be viewed as a volume of effectively reduced elastic moduli. Since most of the energy is distributed along the maximum slope direction this causes an anisotropic decrease of the stiffness of the soil mass which will, in turn, increase the susceptibility of slopes to seismically induced failures. In this context, station S7 of array 1, which was located below the toe of the landslide, shows significant lower directionality and lower amplitudes at 2 Hz, here indicating the reduction of the landslide block's freedom to oscillate.

We interpret the landslide block, which is characterized by low S-wave velocities (bluish color in Figure 6), to be situated above well-consolidated Quaternary material (reddish color

in Figure 6) which is found along the Ak-Bur river valley, similar to the results of Havenith et al. (2000). Given the deficiency of any data on the sediment structure and their thickness, we might infer a sub-vertical interbedded sequence of Quaternary sediments over Neogene material having large implications also for the overlaying soil layers. Since the vertically varying materials most likely have different strengths and weathering properties, the subsurface structure between the high-velocity and low-velocity layers will vary laterally, as already implied in Figure 6. Moreover, the slip of the uppermost soil-layers can also be influenced by the vertically varying layering. The topography of the first impedance contrast is only slightly linked to the surface topography and most likely has more to do with bedrock interbedding and with soil-compression rather than with surface processes. However, we offer a note of caution with regards to its interpretation. Since a set of smoothing parameters has been introduced to the solution, the image will not show a sharp contrast between the two layers, but only a more gradual transition.

To quantify the resolution capacities of the proposed technique we further assess the quality of the tomographic results by means of the spread function, which has been found to be a better way for assessing the resolution than by solely examining the diagonal elements of the resolution matrix. The spread function compresses each row of the model resolution matrix, which is routinely calculated in the inversion and which describes the distribution of information for each grid cell, into a single number describing how strong and peaked the resolution is for that specific cell (Micheline and McEvelly 1991). Although, as shown by Toomey and Foulger (1989), there is no universally applicable value to define the range of acceptable values of the spread function, from its definition, it is clear that in areas of high resolution, the spread function takes small values.

In Figure 7 the resolution of the velocity models of both arrays are displayed by the spread function. Overall, the resolution for the central parts of the model is higher due to the dense ray coverage and the angular distribution of the rays. It has already been shown that where

there is adequate resolution, the S-wave velocity can be reliably mapped (Pilz et al. 2013). However, for the edges of the block, the resolution is generally lower and for the outermost parts the results might be biased by interpolation. Obviously, there is also a decrease in resolution with depth since fewer frequencies sample these parts of the blocks. For the central parts the models can be well resolved down to a depth of around 30 m, being approximately one-third to one-half the wavelength of the lowest frequency. Thereby, deeper structures of the model can only be mapped properly if the array configuration, i.e., the distance between the receivers, allows longer wavelengths to be resolved.

Assuming that in between large sliding events, steady creep will slowly move the uppermost well-resolved soil cover down-slope, with certain parts moving faster while others might be locked, possibly leading to differential stress build-up in the soil layers. Then, if the shear-strength of a soil interface is to drop, it will trigger a shallow landslide, with the material in question being dumped at the base of the steep slopes. On the other hand, the low-velocity clay material along the steep slopes (bluish color in Figure 6), is characterized by a clear fractured structure. Several cracks can be observed at the surface, in particular near S3 of array 2. Such fractures, having preferred orientation perpendicular to the highest slope directions, might generate even stronger amplification effects (Burjanek et al. 2012) and probably enhance instability by a local reduction of the shear strength within the upper layer. Therefore, we suggest that the material in the most unstable part of the landslide body, i.e., the northern part of array 2, is significantly weakened by the cumulative effect of various parallel fractures, causing considerably higher H/V spectral ratio amplitudes.

7. Conclusions

Based on recordings of ambient seismic noise on a landslide, we have shown that with a limited number of stations and recording times of some hours, the fundamental resonance properties of a landslide body as well as detailed images of the local S-wave velocity structure can be obtained, even under pronounced topographic conditions. There is clear evidence that ground motion on potentially unstable slopes can be amplified considerably with a pronounced directional character and with maxima oriented along potential sliding directions. Complementing traditional methods that give only point-wise results, we could show that seismic imaging can add an extra dimension and a deeper perspective down to some tens of meters. In particular, the area and the depth of investigation might be enlarged using more extended array configurations. Since the presented method only uses the phase and not the amplitude information associated with the direct or scattered waves, the scattered part of the correlation functions is stable enough to provide robust velocity estimates not only for imaging purposes, but also allow changes in the medium properties to be monitored. Moreover, the development and the use of wireless sensor networks should ease the efforts necessary for the acquisition of noise data. Likewise, due to wireless data transmission, these units can serve as valuable low-cost tools for monitoring environmental phenomena and are particularly suited for landslide monitoring, since they allow repeated images of the S-wave velocity model. With regards to this, Jongmans et al. (2009) have shown that S-wave velocity is very sensitive to clay consolidation (and the corresponding decrease in porosity) and should tend to zero in the event of liquefaction. Also, a significant drop in v_s of around 10 % was measured at the base of a small active landslide several days before its failure (Mainsant et al. 2012). With the proposed rapid inversion technique, it might be possible to detect the location of changes at the base of a sliding layer. In particular, the presented method can overcome shortcomings like poor spatial resolution, which have been reported for other passive

processing techniques. A combination of tomographic imaging with GPS measurements for displacement tracing might even provide a greater potential for more robust landslide monitoring. As the characteristics of a potentially unstable slope may change rapidly from site, to site the presented results suggest that only a 3D investigation and interpretation of the internal structure of the unstable soil mass can provide a complete picture.

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References

- Abdrakhmatov, K., H. B. Havenith, D. Delvaux, D. Jongmans, P. Trefois (2003), Probabilistic PGA and Arias intensity maps of Kyrgyzstan (Central Asia), *J. Seismol.* **7**, 203-220
- Aki, K. (1957), Space and time spectra of stationary stochastic waves, with special reference to microtremors, *Bulletin of the Earthquake Research Institute* **35**, 415-456
- Aki, W. H., K. Lee (1976), Determination of the three-dimensional velocity anomalies under a seismic array using first *P* arrival times from local earthquakes. A homogeneous initial model, *J. Geophys. Res.* **81**, 4381-4399
- Arai, H., K. Tokimatsu (2004), S-wave velocity profiling by inversion of microtremor H/V spectrum, *Bull. Seismol. Soc. Am.* **94**, 53-63
- Assimaki, D., E. Kausel (2007), Modified topographic amplification factors for a single-faced slope due to kinematic soil-structure interaction, *J. Geotech. Geoenviron. Eng.* **133**, 1414-1431
- Bard, P. Y., M. Bouchon (1985), The two-dimensional resonance of sediment-filled valleys, *Bull. Seismol. Soc. Am.* **75**, 519-541
- Bensen, G. D., M. P. Barmin, A. L. Levshin, F. Lin, M. P. Moschetti, N. M. Shapiro, Y. Yang, M. H. Ritzwoller, (2007), Processing ambient noise seismic data to obtain reliable broadband surface wave dispersion measurements, *Geophys. J. Int.* **169**, 1239-1260

Bindi, D., K. Abdrakhmatov, S. Parolai, M. Mucciarelli, G. Grünthal, A. Ischuk, N. Mikhailova, J. Zschau (2012), Seismic hazard assessment in Central Asia: Outcomes from a site approach, *Soil Dyn. Earthq. Eng.* **37**, 84-91

Bogoslovsky, V. A., A. A. Ogilvy (1977), Geophysical methods for the investigation of landslides, *Geophysics* **42**, 562-571

Bonnefoy-Claudet, S., F. Cotton, P. Y. Bard (2006a), The nature of noise wavefield and its applications for site effects studies: A literature review, *Earth-Science Reviews* **79**, 205-227

Bonnefoy-Claudet, S., C. Cornou, P. Y. Bard, F. Cotton, P. Mozco, J. Kristek (2006b), H/V ratio: a tool for site effect evaluation. Results from 1D noise simulations, *Geophys. J. Int.* **167**, 827-837

Bourdeau, C., H. B. Havenith (2008), Site effects modelling applied to the slope affected by the Suusamyr earthquake (Kyrgyzstan, 1992), *Eng. Geol.* **97**, 126-145

Bourdeau, C., H. B. Havenith, J. A. Fleurisson, G. Grandjean (2004), Numerical modelling of seismic slope stability, in: Engineering geology for infrastructure planning in Europe - a European perspective, R. Hack, R. Azzam, R. Charlier (eds.), Lecture notes in Earth Sciences **1004**, 671-684

Bozzano, F., L. Lenti, S. Martino, A. Paciello, G. S. Mugnozza (2011), Evidences of landslide earthquake triggering due to self-excitation process, *Int. J. Earth Sci.* **100**, 861-879

Brabb, E. E. (1991), The world landslide problem, *Episodes* **14**, 52-61

Burjanek, J., G. Gassner-Stamm, V. Poggi, J. R. Moore, D. Fäh (2010), Ambient vibration analysis of an unstable mountain slope, *Geophys. J. Int.* **180**, 820-828

Burjanek, J., J. R. Moore, F. X. Yugsi Molina, D. Fäh (2012), Instrumental evidence of normal mode rock slope vibration, *Geophys. J. Int.* **188**, 559-569

Campillo, M., A. Paul (2003), Long-range correlations in the diffuse seismic coda, *Science* **299**, 547-549

Capon, J., R. J. Greenfield, R. J. Kolker (1967), Multidimensional maximum-likelihood processing of a large aperture seismic array, *Proc. IEEE* **55**, 192-211

Caris, J. P. T., T. W. van Asch (1991), Geophysical, geotechnical and hydrological investigations of a small landslide in the French Alps, *Eng. Geol.* **31**, 249-276

Chávez-García, F. J., L. R. Sánchez, D. Hatzfeld, (1996), Topographic site effects and HVSr. A comparison between observations and theory, *Bull. Seismol. Soc. Am.* **86**, 1559-1573

Danneels, G., C. Bourdeau, I. Torgoev, H. B. Havenith (2008), Geophysical investigation and dynamic modelling of unstable slopes: case-study of Kainama (Kyrgyzstan), *Geophys. J. Int.* **175**, 17-34

Del Gaudio, V., J. Wasowski (2007), Directivity of slope dynamic response to seismic shaking, *Geophys. Res. Lett.* **34**, L12301, doi: 10.1029/2007GL029842

Del Gaudio, V., J. Wasowski, (2011), Advances and problems in understanding the seismic response of potentially unstable slopes, *Eng. Geol.* **122**, 73-83

Derode, A., E. Larose, M. Tanter, J. de Rosny, A. Tourin, M. Campillo, M. Fink (2003), Recovering the Green's function from field-field correlations in an open scattering medium, *J. Acoust. Soc. Am.* **113**, 2973-2976

Dilley, M., R. S. Chen, U. Deichmann, A. Lerner-Lam, M. Arnold (2005), Natural disaster hotspots. A global risk analysis, The World Bank, Hazard Management Unit, Washington, DC

Fell, R., O. Hungr, S. Leroueil, W. Riemer (2000), Geotechnical engineering of the stability of natural slopes and cuts and fills in soil, in: Proc. GeoEng2000, Int. Conf. on Geotechnical and Geol. Engineering, Melbourne, Australia, Technomic Publishing, Lancaster, **1**, 21-120

Fleming, K., M. Picozzi, C. Milkereit, F. Kuehnlenz, B. Lichtblau, J. Fischer, C. Zulfikar, O. Ozel, SAFER and EDIM working groups (2009), The Self-Organising Seismic Early Warning Information Network (SOSEWIN), *Seismol. Res. Lett.* **80**, 755-771

Frischknecht, C., J. J. Wagner (2004), Seismic soil effect in an embanked deep alpine valley: a numerical investigation of two-dimensional resonance, *Bull. Seismol. Soc. Am.* **94**, 171-186

Geli, L., P. Y. Bard, B. Jullien, (1988), The effect of topography on earthquake ground motion: a review and new results, *Bull. Seismol. Soc. Am.* **78**, 42-63

Grandjean, G. (2006), Imaging subsurface objects by seismic P-wave tomography: numerical and experimental validations, *Near Surf. Geophys.* **4**, 279-287

Haskell, N. A. (1960), Crustal reflection of plane SH waves, *J. Geophys. Res.* **65**, 4147-4150

Havenith, H. B., D. Jongmans, K. Abdrakhmatov, P. Trefois, D. Delvaux, I. A. Torgoev (2000), Geophysical investigations of seismically induced surface effects: case study of a landslide in the Suusamyр valley, Kyrgyzstan. *Surv. Geophys.* **21**, 349-369

Havenith, H. B., D. Jongmans, E. Faccioli, K. Abdrakhmatov, P. Y. Bard (2002), Site effect analysis around the seismically induced Ananevo rockslide, Kyrgyzstan, *Bull. Seismol. Soc. Am.* **92**, 3190-3209

Havenith, H. B., I. Torgoev, A. Meleshko, Y. Alioshin, A. Torgoev, G. Danneels (2006), Landslides in the Mailuu-Suu Valley, Kyrgyzstan – Hazards and impacts, *Landslides* **3**, 137-147

Ignatiev, I. V. (1886), Earthquake in the Tokmak district in 1885, *Proc. of the Imperial Russian Geographic Soc.* **22**, 150-164 (in Russian)

Jongmans, D., S. Garambois (2007), Geophysical investigation of landslides: A review, *Bull. Soc. Geol. Fr.* **178**, 101-112

Jongmans, D., G. Bievre, F. Renalier, S. Schwartz, N. Beaurez, Y. Orengo (2009), Geophysical investigation of a large landslide in glaciolacustrine clays in the Trièves area (French Alps), *Eng. Geol.* **109**, 45-56

King, J., B. Tucker (1984), Observed variations of earthquake motion across a sediment-filled valley, *Bull. Seismol. Soc. Am.* **74**, 137-151

Konno, K., T. Ohmachi (1998), Ground-motion characteristics estimated from spectral ratio between horizontal and vertical components of microtremor, *Bull. Seismol. Soc. Am.* **88**, 228-241

Larose, E., P. Roux, M. Campillo (2007), Reconstruction of Rayleigh-Lamb dispersion spectrum based on noise obtained from an air-jet forcing, *J. Acoust. Soc. Am.* **122**, 3437-3444

Lenti, L., S. Martino (2013), A parametric numerical study of the interaction between seismic waves and landslides for the evaluation of the susceptibility to seismically induced displacements, *Bull. Seismol. Soc. Am.* **103**, 33-56

Mainsant, G., E. Larose, C. Brönnimann, D. Jongmans, C. Michoud, M. Jaboyedoff (2012), Ambient seismic noise monitoring of a clay landslide: Toward failure prediction, *J. Geophys. Res.* **117**, F01030, doi: 10.1029/2011JF002159

Méric, O., S. Garambois, J. P. Malet, H. Cadet, P. Gueguen, D. Jongmans (2007), Seismic noise-based methods for soft-rock landslide characterization, *Bull. Soc. Geol. Fr.* **178**, 137-148

Michellini, A., T. V. McEvilly (1991), Seismological studies at Parkfield, I. Simultaneous inversion for velocity structure and hypocenters using cubic B-spline parameterization, *Bull. Seismol. Soc. Am.* **81**, 524-552

Moldobekov, B., A. Sarangoev, S. Usupaev, A. Meleshko (1997), Prognosis of natural hazards on the territory of the Kyrgyz Republic, 172 p., All-Press, Bishkek, Kyrgyzstan (in Russian).

Moore, J. R., V. Gischig, J. Burjanek, S. Loew, D. Fäh (2011), Site effects in unstable rock slopes: dynamic behavior of the Randa instability (Switzerland), *Bull. Seismol. Soc. Am.* **101**, 3110-3116

Nadim, F., O. Kjekstad, P. Peduzzi, C. Herold, C. Jaedicke (2006), Global landslide and avalanche hotspots, *Landslides* **3**, 159-173

Nakamura, Y. (1989), A method for dynamic characteristics estimations of subsurface using microtremors on the ground surface, *Q. Rep. Railway Tech. Res. Inst. Japan* **30**, 25-33

Nogoshi, M., T. Igarashi (1970), On the propagation characteristics of microtremors, *J. Seism. Soc. Japan* **23**, 264-280 (in Japanese with English abstract)

Nogoshi, M., T. Igarashi (1971), On the amplitude characteristics of microtremor, *J. Seism. Soc. Japan* **24**, 26-40 (in Japanese with English abstract)

Ohori, M., A. Nobata, K. Wakamatsu (2002), A comparison of ESAC and FK methods of estimating phase velocity using arbitrarily shaped microtremor arrays, *Bull. Seismol. Soc. Am.* **92**, 2323-2332

Parolai, S., P. Bormann, C. Milkereit (2002), New relationships between V_s , thickness of sediments, and resonance frequency calculated by the H/V ratio of seismic noise for the Cologne area (Germany), *Bull. Seismol. Soc. Am.* **92**, 2521-2527

Petley, D. N. (2009), On the impact of urban landslides, in: Engineering Geology for Tomorrow's Cities, M. G. Culshaw, H. J. Reeves, I. Jefferson, T. W. Spink (eds.), *Engineering Geology Special Publications* **22**, 83-99, London, UK

Picarelli, L., G. Urciuoli, M. Ramondini, L. Comegna (2005), Main features of mudslides in tectonised highly fissured clay shales, *Landslides* **2**, 15-30

Picozzi, M., C. Milkereit, S. Parolai, K. H. Jaeckel, I. Veit, J. Fischer, J. Zschau (2010), GFZ Wireless Seismic Array (GFZ-WISE), a wireless mesh network of seismic sensors: New perspectives for seismic noise array investigations and site monitoring, *Sensors* **10**, 3280-3304

Pilz, M., S. Parolai, M. Picozzi, D. Bindi (2012), Three-dimensional shear wave velocity imaging by ambient seismic noise tomography, *Geophys. J. Int.* **189**, 501-512

Pilz, M., S. Parolai, D. Bindi (2013), 3D passive imaging of complex seismic fault systems: evidence of surface traces of the Issyk-Ata fault (Kyrgyzstan), *Geophys. J. Int.* **194**, 1955-1965

Renalier, F., D. Jongmans, M. Campillo, P. Y. Bard (2010), Shear wave velocity imaging of the Avignonet landslide (France) using ambient noise cross correlation, *J. Geophys. Res.* **115**, F03032, doi: 10.1029/2009JF001538.

Roessner, S., H. U. Wetzler, H. Kaufmann, A. Sarnagoev (2005), Potential of satellite remote sensing and GIS for landslide hazard assessment in Southern Kyrgyzstan (Central Asia), *Nat. Haz.* **35**, 395-416

Roten, D., D. Fäh, C. Cornou, D. Giardini (2006), Two-dimensional resonances in Alpine valleys identified from ambient vibration wavefields, *Geophys. J. Int.* **165**, 889-905

Sanchez-Sesma, F. J., M. Campillo (1993), Topographic effects for incident P, SV and Rayleigh waves, *Tectonophys.* **218**, 113-125

Schuster, R. L., R. W. Fleming (1986), Economic losses and fatalities due to landslides, *Bull. Am. Assoc. Eng. Geol.* **23**, 11-28

Spudich, P., M. Hellweg, W. H. K. Lee (1996), Directional topographic site response at Tarzana observed in aftershocks of the 1994 Northridge, California, earthquake: implications for mainshock motions, *Bull. Seismol. Soc. Am.* **86**, S193-S208

Steimen S., D. Fäh, F. Kind, C. Schmid, D. Giardini (2003), Identifying 2D resonance in microtremor wave fields, *Bull. Seism. Soc. Am.* **93**, 583-599

Strollo A., S. Parolai, K. H. Jäkel, S. Marzorati, D. Bindi (2008), Suitability of short-period sensors for retrieving reliable H/V peaks for frequencies less than 1 Hz, *Bull. Seismol. Soc. Am.* **98**, 671-681

Strom, A., O. Korup (2006), Extremely large rockslides and rock avalanches in the Tien Shan Mountains, Kyrgyzstan, *Landslides* **3**, 125-136

Toomey, D. R., G. R. Foulger (1989), Tomographic inversion of local earthquake data from the Hengill-Grensdalur central volcano complex, Iceland, *J. Geophys. Res.* **94**, 17497-17510

Travelletti, J., J. P. Malet (2012), Characterization of the 3D geometry of flow-like landslides: a methodology based on the integration of heterogeneous multi-source data, *Eng. Geol.* **128**, 30-48

Trifonov, V. G., O. V. Soboleva, R. V. Trifonov, G. A. Vostrikov (2002). Recent geodynamics of the Alpine-Himalayan collision belt, *Transactions of the Geological Institute RAS* 541, 224 p. (in Russian)

Tucker, B., J. King (1984), Dependence of sediment-filled valley response on input amplitude and valley properties, *Bull. Seismol. Soc. Am.* **74**, 153-165

Tukey, J.E. (1974), Introduction to today's data analysis, in: *Proceedings of the Conference on Critical Evaluation of Chemical and Physical Structural Information*, pp. 3-14, D. R. Lide, M. A. Paul (eds.), *National Academy of Sciences, Washington, D. C.*

Tursugaziev, B. T., O. W. Petrov, W. P. Zubkov, A. G. Konjuchov, G. S. Biskae, N. G. Wlasov (2008), Geological map of the Kyrgyz Republic, ed. by the State Agency for Geological and Mineralogical Resources of the Republic of Kyrgyzstan and the Russian Scientific Research Institute Karschinsky, Saint Petersburg, Russia

Yang, Y., M. H. Ritzwoller (2008), Characteristics of ambient seismic noise as a source for surface wave tomography, *Geochem. Geophys. Geosys.* **9**, doi: 10.1029/2007GC001814

Yao, H., R. van der Hilst (2009), Analysis of ambient noise energy distribution and phase velocity bias in ambient noise tomography with application to SE Tibet, *Geophys. J. Int.* **179**, 1113-1132

Figure captions

Figure 1: Relief of the south-eastern rim of the Fergana Basin and a geological map of the study region. White dots represent recent earthquakes ($M \geq 3.5$) between 2000 and 2012. The main active faults (thin black lines, Trifonov et al. 2002) are indicated.

Figure 2: Configuration of the array measurements (red dots: array 1, black dots: array 2) over the landslide (shaded area; small black box in Figure 1). Stations mentioned in the text are indicated. Black lines are contour lines with 10 m spacing. The coordinates are given in meters of the Universal Transverse Mercator coordinate system. Letters and arrows indicate the positions from which the photos of array 1 (right) and array 2 (left) were taken.

Figure 3: Polar diagrams of the mean horizontal-to-vertical spectral ratios obtained from seismic noise recordings for the measurement sites (top-left labels) of the northern (left) and southern arrays (right). See Figure 2 for the locations of the measurement sites indicated.

Figure 4: Fourier spectra of all components for selected stations of array 2 (top panels) and the array 1 (bottom panels). See Figure 2 for the locations of the measurement sites indicated.

Figure 5: f - k power density function showing the distribution of noise sources for array 2 using the maximum likelihood method for (a) 4Hz and (b) 13 Hz.

Figure 6: Left: 3D S-wave velocity models below array 1 (top) and array 2 (bottom) obtained after 500 iterations. The arrows indicate the approximate locations of the stations whose H/V spectral ratios are shown in Figure 3. Right: Corresponding isobaths indicating the thickness of the low-velocity body (S-wave velocity lower than 500 m/s).

Figure 7: Horizontal cross sections of the normalized spread function for the S-wave velocity model shown in Figure 6 for array 1 (top) and array 2 (bottom).

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Figure 2: Top: Configuration of the array measurements (red dots: array 1, black dots: array 2) over the landslide (shaded area; small black box in Figure 1). Stations mentioned in the text are indicated. Black lines are contour lines with 10 m spacing. The coordinates are given in meters of the Universal Transverse Mercator coordinate system. Letters and arrows indicate the positions from which the photos of array 1 (right) and array 2 (left) were taken.

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