



Hydrogeological investigation of landslides Urbas and Čikla above the settlement of Koroška Bela (NW Slovenia)

Hidrogeološke raziskave plazov Urbas in Čikla nad naseljem Koroška Bela (SZ Slovenija)

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Abstract

The area above the settlement of Koroška Bela is highly prone to slope mass movements and poses a high risk for the safety of the settlement. To get an insight into the hydrogeological conditions and processes which can affect mass movements in this area, hydrogeological investigations, including hydrogeological mapping, discharge measurements of springs, performance of infiltrometer and slug tests were performed. The results of these investigations show complex and heterogeneous hydrogeological conditions, predisposed by geological and tectonic setting and active mass movements which cannot be uniformly described. Observed large fluctuations in the rate of discharge of springs and groundwater level in observation wells are highly dependent on meteorological conditions. Estimated hydraulic conductivity of the soil is relatively high (2×10^{-4} m/s) and reflects the loose structure and high content of organic matter in the upper part of the forest soil. Hydraulic conductivity of more permeable sections of boreholes is in general higher in the upper parts, in predominantly gravel layers (in range from 2×10^{-3} to 1×10^{-5} m/s), than in the deeper clayey gravel parts (3×10^{-5} to 1×10^{-7} m/s). In the area of the Čikla landslide the average hydraulic conductivity is estimated at 8.99×10^{-4} m/s and is higher than in the area of the Urbas landslide (3.05×10^{-4} m/s).

Izveček

Na območju nad naseljem Koroška Bela je velika nevarnost nastanka pobočnih masnih premikov, kar predstavlja tveganje za varnost naselja. Za razumevanje hidrogeoloških pogojev in procesov, ki lahko vplivajo na masne premike na tem območju, so bile izvedene hidrogeološke raziskave, ki so vključevale hidrogeološko kartiranje, meritve pretokov izvirov ter izvedbo infiltrometrskih in nalivalnih preizkusov. Rezultati raziskav kažejo zapletene in heterogene hidrogeološke razmere, pogojene z geološkimi in tektonskimi značilnostmi širšega območja ter aktivnimi masnimi premiki, ki jih ni mogoče enoznačno opisati. Opazovana velika nihanja pretokov izvirov in gladine podzemne vode v opazovalnih vrtinah so močno odvisna od meteoroloških razmer. Ocenjeni koeficient prepustnosti tal je relativno visok (2×10^{-4} m/s) in odraža rahla tla z visoko vsebnostjo organske snovi, ki so značilna za gozd. Koeficient prepustnosti bolj prepustnih odsekov vrtin je v splošnem višji v zgornjih delih, v plasteh s prevladujočim gruščem (v razponu med 2×10^{-3} in 1×10^{-5} m/s), kot v globljih delih vrtin, v zaglinjenih plasteh (med 3×10^{-5} in 1×10^{-7} m/s). Na območju plazu Čikla je ocenjen povprečni koeficient prepustnosti 8.99×10^{-4} m/s in je višji kot na območju plazu Urbas (3.05×10^{-4} m/s).

Nomenclature

v	infiltration velocity (m/s)
ΔV	volume of infiltrated water during measuring phase when flow is close to steady-state condition (m ³)
A_i	area of inner infiltrometer ring (m ²)
t	time (s)
K	hydraulic conductivity (m/s)
i	hydraulic gradient (-)
z_w	saturated thickness trough which flow occurs (underground depth of infiltrated water reach) (m)
h	hydraulic head (m)
V	total volume infiltrated over the entire duration of the infiltration test (m ³)
θ	volumetric water content - ratio of water volume to total soil volume (-)
θ_s	saturated soil volumetric water content (-)
θ_i	initial soil volumetric water content (-)
$\Delta\theta$	difference between θ_s and θ_i (-)
A	effective area of casing or excavation (m ²)
F	shape factor (-)
L	length of effective intake or filtering zone (m)
D	diameter of effective intake or well point (m)
Q	injection rate (m ³ /s)
m	shape factor (-)
r	well screen radius (m)

Introduction

Hydrogeological conditions have an important role in the stability of slopes. Part of the rainfall that infiltrates into the ground can contribute to the increase of pore pressures and saturation (decrease of suction) of the ground or even to the rise of the groundwater table, all of which may initiate mass movements on slopes. The changes of hydraulic conditions propagate from the ground surface into the subsoil and slope failures are more likely to take place a certain time after a rainfall event (Jemec Auflič et al., 2016; Nilsen et al., 1976). The time rate of propagation depends on the hydrogeological properties of the ground. The response of the slope material relies primarily on the ground material; granular soils are more sensitive to short-duration intense rainfall events, whereas fine sediment (clay-like) materials are sensitive to long-duration and moderate intensity precipitation (Casagli et al., 2006).

Beside the in-situ infiltration, the occurrence of groundwater that affects landslides can result from lateral flow, or exfiltration from the bedrock. Due to the unique groundwater flow pattern in every landslide, triggering mechanisms related to hydrogeology are complex (Brönnimann, 2011). The local hydraulic field is related to the structure of the landslide mass, which is a result of land-

slide activity, the formation of cracks, and increased hydraulic conductivity due to soil saturation. Therefore, the hydraulic conductivity of the landslide mass, which controls the groundwater flow, may be very heterogeneous (Guglielmi et al., 2002).

The area of Koroška Bela settlement in NW Slovenia has experienced severe debris-flow events in the past and a big part of the settlement is even built on an alluvial fan of past debris flows (Jež et al., 2008). Due to the geological, tectonic and hydrogeological conditions, the area above the settlement is highly prone to different slope mass movements, posing a high risk for the safety of the settlement (Peternel et al., 2016). To assess the risk of mass movements and provide supportive information for the planning of safety measures for the protection of the settlement, various series of geological (Jež et al., 2008), engineering (Peternel et al., 2018), geophysical, geomechanical, geomorphological, geodetic, and hydrogeological investigations were performed (Peternel et al., 2017).

In this paper, hydrogeological investigations, which include hydrogeological mapping, discharge measurements of springs, performance of infiltrometer and slug tests in the period from 29 August to 26 October 2017 are presented. The aim of these investigations was to get an insight into the hydrogeological conditions and processes which can affect mass movements in the area above Koroška Bela settlement.

Study area

The areas of the Urbas (85000 m²) and Čikla (8000 m²) landslides lie in the NW part of Slovenia in the hinterland of the settlement Koroška Bela (figs. 1a, 1b). The areas of landslides extend from an elevation of 1150 m to 1300 m on the south to southwest oriented slope with gradient ranging generally from 30° to 70°. Average annual precipitation and temperature in this area range from 2000 mm to 2600 mm and from 3 °C to 5 °C respectively (ARSO, 2018a).

The area is a part of the high east-west extended mountainous ridge of Karavanke, a mountain range of the Eastern Alps. The terrain of the Karavanke Mountains consists of long and prominent ridges, whose slopes fall steeply to the northern and southern side. The ridges are interrupted by long, deep and narrow valleys streams, exhibiting properties typical for high-mountain watercourses: irregular and uneven channels with rapid flows (Brenčič & Poltnig, 2008).

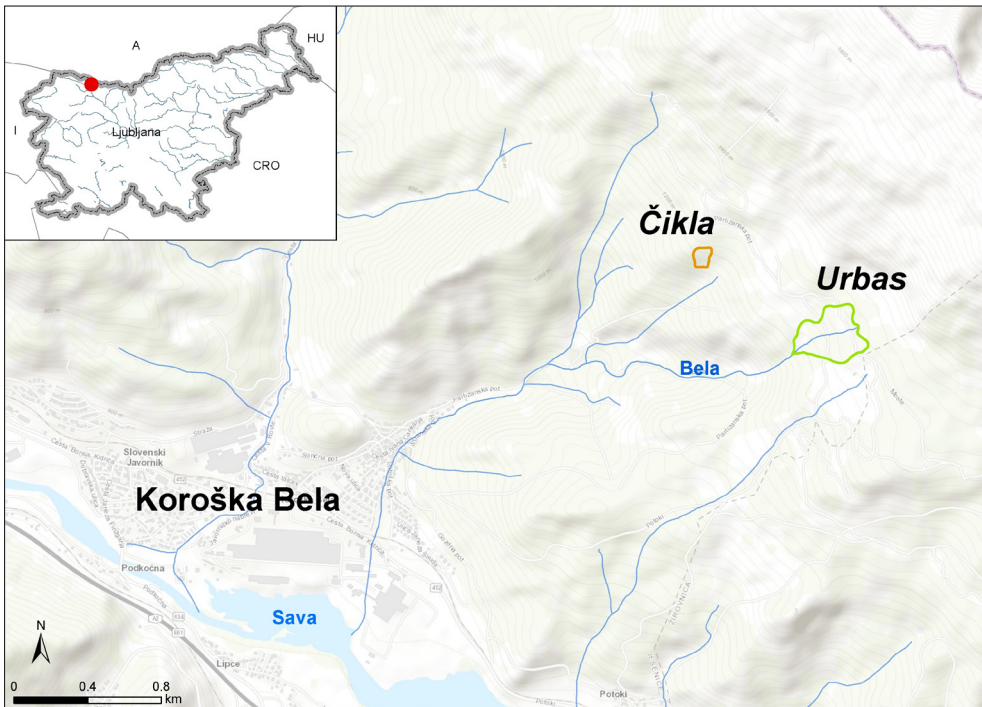


Fig. 1a. Location of study areas.

Regional hydrogeological setting

In general, two types of aquifers characterize regional hydrogeological settings, intergranular and karst-fissured aquifers. Intergranular aquifers are presented by sediments formed as a consequence of intense slope processes typical for steeper mountain regions (slope sediments). In these sediments, groundwater may be retained,

but their spatial extension is limited, and such aquifers mostly represent negligible groundwater sources (Prestor et al., 2008). In some areas, groundwater can be found in sediments which were created by slowly slipping material formed by landslides, such as considered at the study area. In these cases, groundwater occurs at the contact between the above lying sloping sediment and the underlying lower permeable bed-

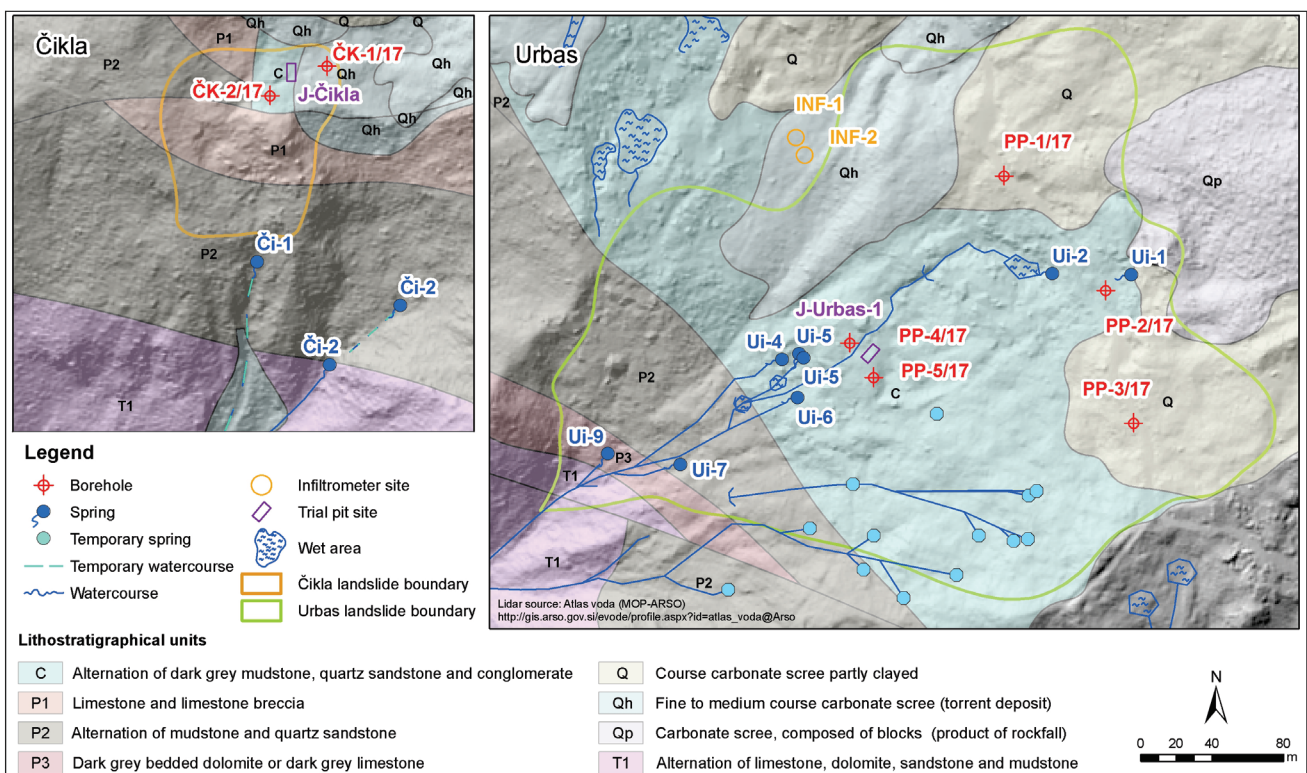


Fig. 1b. The areas of the Čikla and Urbas landslides with the locations of measurements and identified springs.

rock. Within those layers, the groundwater flow is often interrupted by the presence of low permeable clay or mud material (Brenčič & Poltnig, 2008).

The karst-fissured aquifers are represented by carbonate rocks, which also form the drainage area of the investigated landslides. They are characterized by numerous karst springs of very diverse outflow regime. When considering karst-fissured aquifers in the Karavanke Mts. area, it is necessary to also note the differences in the hydrogeological characteristics of limestones and dolomites, which also influence the development of karst phenomena. The permeability of dolomites comparing to limestone is usually lower. The hydrogeological properties of dolomites depend also on the type of diagenesis, or whether dolomites were formed as a result of primary or secondary dolomitization. Most karst-fissured aquifers in the Karavanke Mts. are unconfined, characterized by direct recharge of infiltrated precipitation. In the South Karavanke Mts., springs occur at the (tectonic) contact of extensive aquifer with low permeable layers and represent the high-water overflow from this aquifer. The groundwater level within the karst-fissured aquifer is strongly related to precipitation and is highly variable. This is also reflected in the high discharge fluctuations of springs. In dry periods in summer, some springs can almost dry out, while in periods of intense rainfall they can reach discharges of a few 100 l/s (Brenčič & Poltnig, 2008).

Methods

The field investigation in the two months period (29. 8. 2017 – 26. 10. 2017) started with the mapping of springs and other hydrogeological phenomena in the study area. Based on hydrogeological mapping locations of discharge measurements, new observation wells and in situ measurements of hydraulic conductivity were defined (fig. 1b).

Discharges of Ui-1 and Či-2 springs were estimated with using bucket and a stopwatch. The flows of other springs could not be collected in a bucket; therefore discharges were only visually estimated. Field measurements of pH value, electrical conductivity and temperature (Table 1) were carried out with measuring instrument pH/Cond 340i, and measurements of ORP (Oxydation Reduction Potential) and dissolved oxygen with instrument Multi 3410 SET C, both products of the WTW company.

In situ measurements of hydraulic conductivity

Infiltrometer tests

Infiltrometer tests enable the estimation of infiltration rate (infiltration capacity) as the maximum rate at which soil will absorb water impounded on the surface at a shallow depth, when adequate precautions are taken regarding boundary effects (Richards, 1952; Johnson, 1963). In the study area, a double ring infiltrometer was used which creates vertical (one-dimensional) flow of water beneath the inner ring and simplifies interpretation of measurements (Köhne et al. 2011) (fig. 2a). The method is based on the Darcy law of groundwater flow through intergranular porous material at steady state conditions (constant-head and constant infiltration velocity). The volume of water used during each measured time interval was converted into the incremental infiltration velocity using the equation (EN ISO 22282-5-2012, 2012):

$$v = \frac{\Delta V}{A_i \Delta t}$$

from Darcy law follows:

$$v = Ki$$

hydraulic gradient can be approximated as:

$$i = \frac{z_w + h_i}{z_w}$$

saturated thickness (z_w) through which flow occurs can be determined as:

$$z_w = \frac{V}{A_i \Delta \theta}$$

where $\Delta \theta$ is difference between the saturated soil volumetric water content (θ_s) and the initial soil volumetric water content (θ_i).

To assure horizontal surface and avoid surface cracks and fissures, typical for landslide mass, infiltrometer tests were performed at two locations close to the boundary of the Urbas landslide (fig. 1b), which could be accessed by car and supplied with the required amount of water. At these locations is present a soil typical for coniferous forest which covers a large part of the study area (fig. 2b). The rings were inserted into the ground to a penetration depth of 0.05 m and a constant-head was maintained in both rings ($h_i=0.05$ m) with the supply of water with Mariotte's bottle (fig. 2a). The tests were performed on 26. 10. 2017, three days after a moderate rainfall event.



Fig. 2a. Double ring infiltrometer with Mariotte's bottle (photo: Zmago Bole).



Fig. 2b. The top soil at location INF-1 (photo: Zmago Bole).

Falling and constant-head tests

The constant and falling-head tests in the cased boreholes or trial pits are commonly-used field methods for estimating the hydraulic conductivity of porous material. The review of existing practical engineering procedures for the performance of constant and falling-head tests was presented by Brenčič (2011), while the theoretical background can be found in literature on hydraulic tests (e.g., Batu, 1998; Butler, 1998; Bouwer & Rice, 1976; Cedergren, 1989; Kruseman & De Ridder, 1990).

Falling-head tests

The widely used interpretation of falling-head tests is the Hvorslev method (Hvorslev, 1951), because it provides a quick and inexpensive procedure for obtaining a relatively reliable estimation

of hydraulic conductivity from a single well. This method is based on the interpretation of hydraulic head changes in time and is suitable for wells that are open in a short section at their base. Hvorslev (1951) found out that the return of the hydraulic head to the original, static hydraulic head occurs at an exponential rate with the time and is dependent on the hydraulic conductivity of the porous material. The Hvorslev method is valid for confined aquifers. However, Bouwer (1989) observed that the water table boundary in an unconfined aquifer has little effect on test performance results, unless the top of the well screen is positioned close to the water table. Therefore, in many cases, we may apply the Hvorslev solution for confined aquifers to approximate unconfined conditions. The basic equation for unsteady conditions is as follows (Hvorslev, 1951):

$$K = \frac{A}{\Delta t \times F} \times \ln \frac{h_1}{h_2}$$

where the shape factor F is expressed as:

$$F = \frac{2\pi \times L}{\ln\left(\frac{2L}{D}\right)}$$

or when the water flow is limited only through well walls:

$$F = \frac{2\pi L}{\ln\left(\frac{2L}{D}\right)} 2.75D$$

or when the water flow is limited only on the bottom of tube:

$$F = 2.75D$$

Hvorslev method (Hvorslev, 1951) is valid only if the length of the well screen is more than 8 times larger than its radius ($L/r > 8$). The transient solution omits storativity of the formation and assumes a quasi-steady-state flow between the control well and the tested formation. The following assumptions should be applied to the use of the Hvorslev method: the aquifer has infinite areal extent; the aquifer is homogeneous and of uniform thickness; the aquifer potentiometric surface is initially horizontal; a volume of water is injected or discharged instantaneously (Hvorslev, 1951).

The second method used in this investigation was proposed by Schneebeli (1987). This method shows the same results as the Hvorslev method, if small diameter wells (negligible well storage) are used. In the case of large diameter wells or objects (i.e., excavated trial pit), the results of the

methods differ, since the way that the geometric shape factor in Schneebeli's equation is determined is more robust and its value is proportional to the geometry of the tested object. The shape factor suits the recharge with semiellipsoidal shape (Dachler, 1936; Hvorslev, 1951). Schneebeli (1987) suggested the equation as follows:

$$K = m \cdot A \cdot 2.3 \cdot \frac{\log\left(\frac{h_1}{h_2}\right)}{t_1 - t_2}$$

where m represents a geometric shape factor:

$$m = \frac{\alpha}{D}$$

and α is expressed as:

$$\alpha = \frac{\ln\left(\frac{L}{D} + \sqrt{\left(\frac{L}{D}\right)^2 + 1}\right)}{2 \cdot \pi \cdot \left(\frac{L}{D}\right)}$$

However, it should be noted that there are some other limitations that affect the reliability of falling-head tests in large diameter wells or objects. For example, the hydraulic conductivity of the material should not be high, or the changes of hydraulic head should be fast, and in large diameter wells the capacity effect can greatly slow down the reduction of hydraulic head (Mace, 1999).

Constant-head test

The constant-head test is based on steady state conditions, where the hydraulic head in the borehole is stabilized with a constant water injection. The stabilized hydraulic head reached in a borehole or other tested object in which a constant injection is given by the following expression (Custodio & Llamas, 1983):

$$h = \frac{Q}{FK}$$

where F is a shape factor of tested object and is expressed as:

$$F = \frac{2\pi L}{\ln\frac{L}{r}}$$

From the test conditions and the injection rate to stabilize water level, the application directly

obtains the hydraulic conductivity as follows:

$$K = \frac{Q}{2\pi Lh} \ln\frac{L}{r}$$

Application of hydraulic conductivity measurements in the field

Falling-head and constant-head test were performed in three different types of hydrogeological objects: trial pits, open-bottom tubes, and boreholes. In all objects, the water level was measured with a pressure probe (PPI 200) with data



Fig. 3. Trial pit for performing falling-head test (photo: Zmago Bole).



Fig. 4. Open-bottom tubes for performing falling-head test (photo: Zmago Bole).

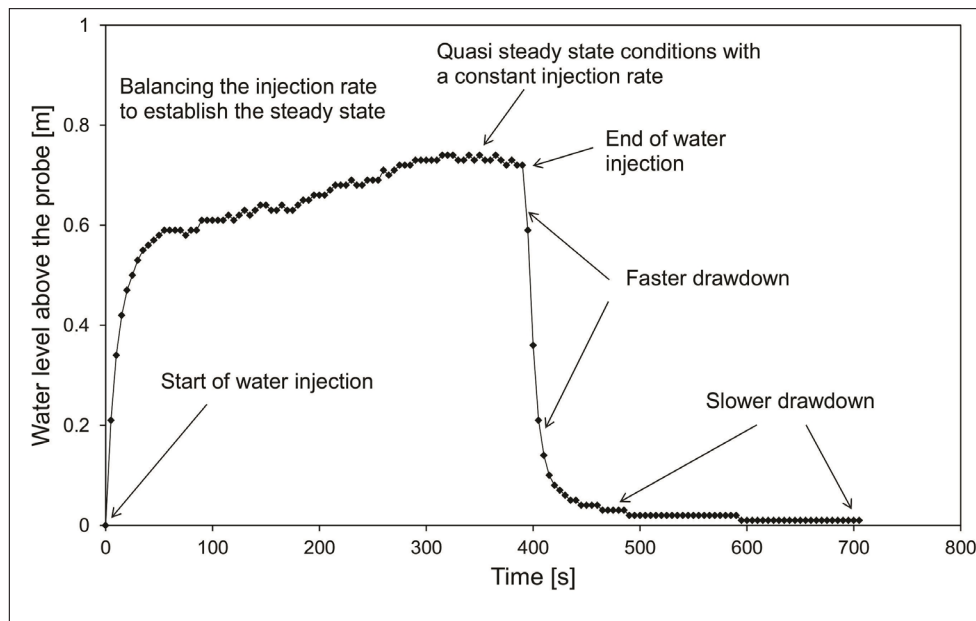


Fig. 5. An example of falling-head test performed in the well (PP-4/17).

logger (GSR 120NTG), produced by the Eltratec company. The pressure probe measures to an accuracy of 1 to 10 mm and in a range from 0 to 30 m of water column. Each test was carried out within specific field conditions; therefore the injection rate and duration were different for each individual tested object. Water level measurements were recorded at 5 to 10 s intervals.

Falling-head tests in trial pits were carried out in rectangular-shaped excavations made with a digger (fig. 3). Falling-head tests in open-bottom tubes were carried out with a tube (internal diameter of 100 mm), inserted into the ground surface or bottom of trial pit. The J-Urbas trial pit was 2.7 m deep. In the deepest part, it was 1.5 m long and 0.6 m wide (fig. 3). In its upper part, two shelves were made, at the depths 2.0 and 0.6 m. On the upper shelf, an open-bottom tube test was performed (fig. 4). Tests at this location were made on 8. 9. 2017. The J-Čikla trial pit was 2.4 m deep, 1.6 m long and 0.6 m wide. Near the trial pit, two open-bottom tube tests were performed (Table 3). In both locations, the upper soil was removed, in the first location down to 0.4 m (Tube-1) and in the second location down to 0.2 m (Tube-2). Tests at this location were made on 18. 9. 2017.

The borehole falling-head test was performed on more permeable sections of boreholes, which were determined on the basis of lithological borehole logs (Peternel et al., 2017). On these sections, drilling with a core drill was carried out after the completion of the technical column. Measurements were performed in 1 – 2 m open hole sections. Before the test started, the occurrence of groundwater was checked.

At locations where we could provide enough water, the performance of falling-head test was carried out in two steps. The second step or second injection was performed after the stabilization of the water level in previous step. Moreover, at suitable borehole sections we also performed a constant-head test. In the study area, the most permeable layers are presented in the form of several-metre-thick coarse (gravel) sediments. In most cases, the tested borehole section included different types of lithology. The common example of the falling-head test performance is shown in the fig.5. The example shows that in the upper 0.7 m of the borehole, in the gravel layers, the drawdown is rapid, while in the lower part, in clayey material, drawdown is significantly slower. In such cases, two different lines were used to interpret the slope of the curve, for the purposes of the processing and estimation of the coefficient of hydraulic conductivity.

Results

In the area of Urbas landslide, seven permanent springs were identified and one in the broader area of Čikla landslide, beside permanent springs several temporal springs were observed (fig. 1b). Spring Urbas (Ui-1) is captured for the water supply of nearby mountain huts Potoška planina and Valvasorjev dom pod Stolom. Discharge measurements during the field campaign show a large fluctuation of discharge, from up to 25 l/s shortly after rainfall, to very low or even no discharge in dry conditions (figs. 6a and 6b). Basic water parameters measured at permanent springs on 29. 8. 2018 are presented in Table 1.

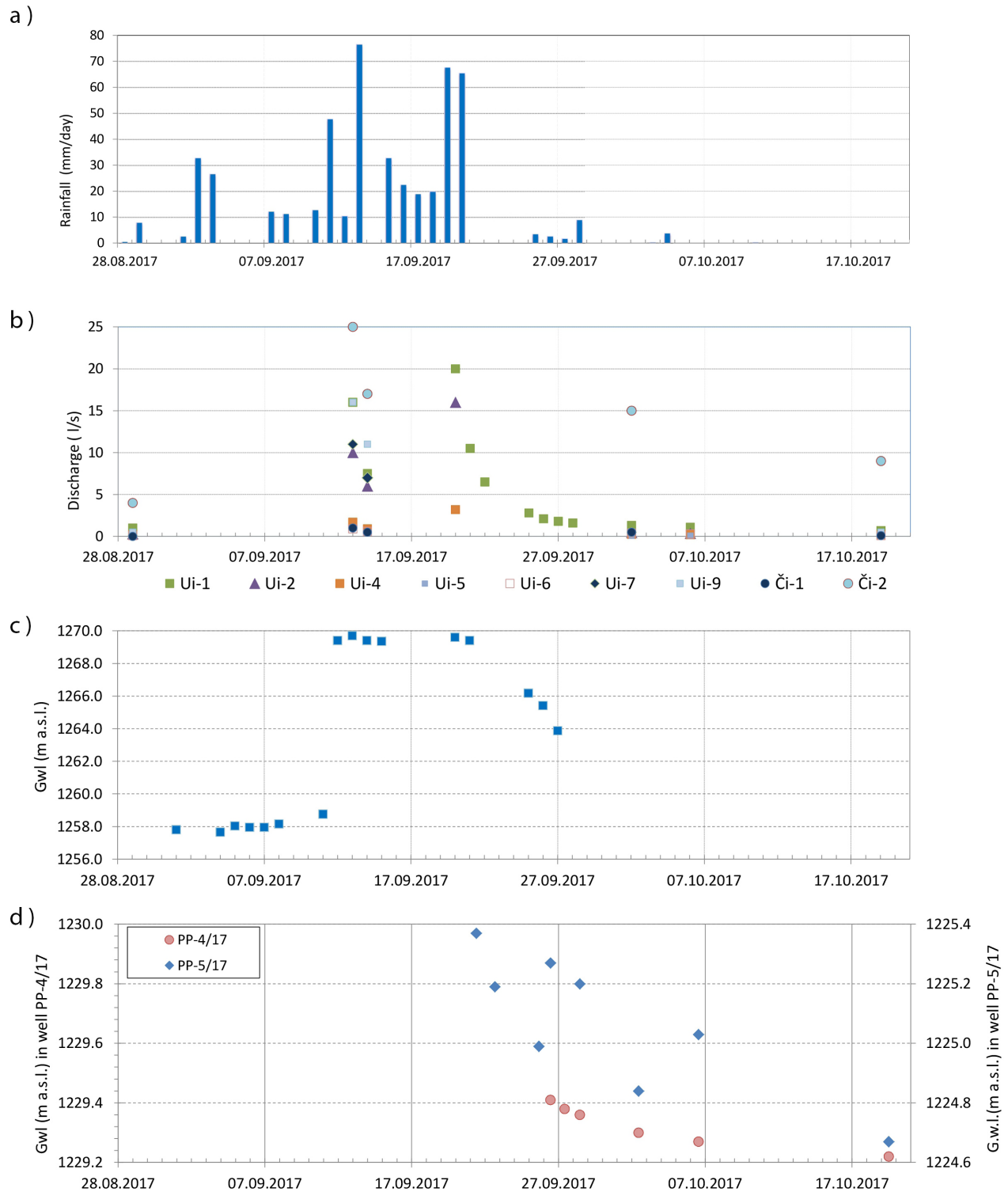


Fig. 6. a) Precipitation measured at meteorological station Planina pod Golico (ARSO, 2018b), b) Observed discharge at springs, c) Groundwater levels in well PP-2/17, d) Groundwater levels in wells PP-4/17 (circles) and PP-5/17 (squares).

Occasional manual measurements of groundwater level during the field campaign were performed in 7 boreholes (Peternel et al., 2017, fig. 6c). After the field campaign, continuous measurements with a pressure probe were recorded in wells PP-4/17 and ČK-1/17 (fig. 7).

Results of two infiltrimeter tests performed in the study area (fig. 1b) are summarized in Table 2.

Hydraulic conductivity of upper ground, estimated with falling-head tests in trial pit and open-bottom tube is presented in Table 3. In the area of landslide Urbas hydraulic conductivity of upper silted gravel layers is in the range between 1.65×10^{-5} m/s and 1.96×10^{-5} m/s (avg. 1.8×10^{-5} m/s). Similar values with an average of 3.24×10^{-5} m/s were estimated for upper clayey sandy layers with gravel at Čikla location.

Table 1. Basic water parameters: pH, electrical conductivity (EC), temperature (T), dissolved oxygen (DO), redox potential (Eh), and discharge (Q), measured at permanent springs on 29.8.2018.

Spring	Elevation (m a.s.l.)	pH	EC ($\mu\text{S/cm}$)	T (oC)	DO (mg/l)	Eh (mV)	Q (l/s)
Ui-1	1267	8.3	186	4.0	11.9	434	1.0
Ui-2	1260	8.5	240	11.1	10.1	417	0.3
Ui-4	1233	7.7	226	5.9	10.4	506	0.2
Ui-5	1242	7.8	228	6.8	10.6	469	0.1
Ui-6	1237	7.7	234	6.1	10.6	515	0.1
Ui-7	1199	7.8	255	8.0	10.4	435	0.1
Ui-9	1176	8.0	228	5.7	11.1	441	0.5
Či-2	1027	8.2	290	6.8	11.3	393	4.0
Mean	1205	8.0	236	6.8	10.8	451	0.3
Range	240	0.8	104	7.1	1.8	122	3.9

Table 2. Results of infiltrometer tests.

Parameter	INF-1	INF-2
Δt (s)	1007	664
ΔV (m^3)	0.017	0.015
Δ (-)	0.2	0.2
V (m^3)	0.022	0.026
z_w (m)	1.43	1.40
K (m/s)	2.10×10^{-4}	2.81×10^{-4}

Table 3. Hydraulic conductivity of upper ground, estimated with falling-head tests in trial pit and open-bottom tube.

Location	Soil type	Classification symbol	Testing object	Pouring water step	Unsteady K [m/s] - Hvorslev	Unsteady K [m/s] - Schneebeli
J-Urbas	Silted gravel	GM	Trial pit		/	1.96×10^{-05}
			Tube		1.65×10^{-05}	/
J-Čikla	Clayey sandy silt with gravel	ML/GC	Trial pit		/	2.91×10^{-05}
Near J-Čikla			Tube-1	1.	2.55×10^{-05}	/
				2.	5.74×10^{-05}	/
Tube-2			1.	2.72×10^{-05}	/	
	2.	2.30×10^{-05}	/			

Hydraulic conductivities of more permeable sections of boreholes ČK-1/17 and PP-4/17 (fig. 8), estimated with performance of falling-head and constant-head tests in boreholes are presented in Table 4 and Table 5.

Estimated hydraulic conductivities in well ČK-1/17 range between 2.64×10^{-3} and 7.78×10^{-6} m/s. The Hvorslev and Schneebeli methods give practically identical results. The

average value calculated for all three tested depth using only results of Hvorslev (or Schneebeli) and steady state method is 8.99×10^{-4} . In well PP-4/17 values of hydraulic conductivities ranges between 1.67×10^{-3} and 1.02×10^{-7} m/s. The average value calculated for all five tested depth intervals using results of Hvorslev (or Schneebeli) and steady state method is 3.05×10^{-4} m/s.

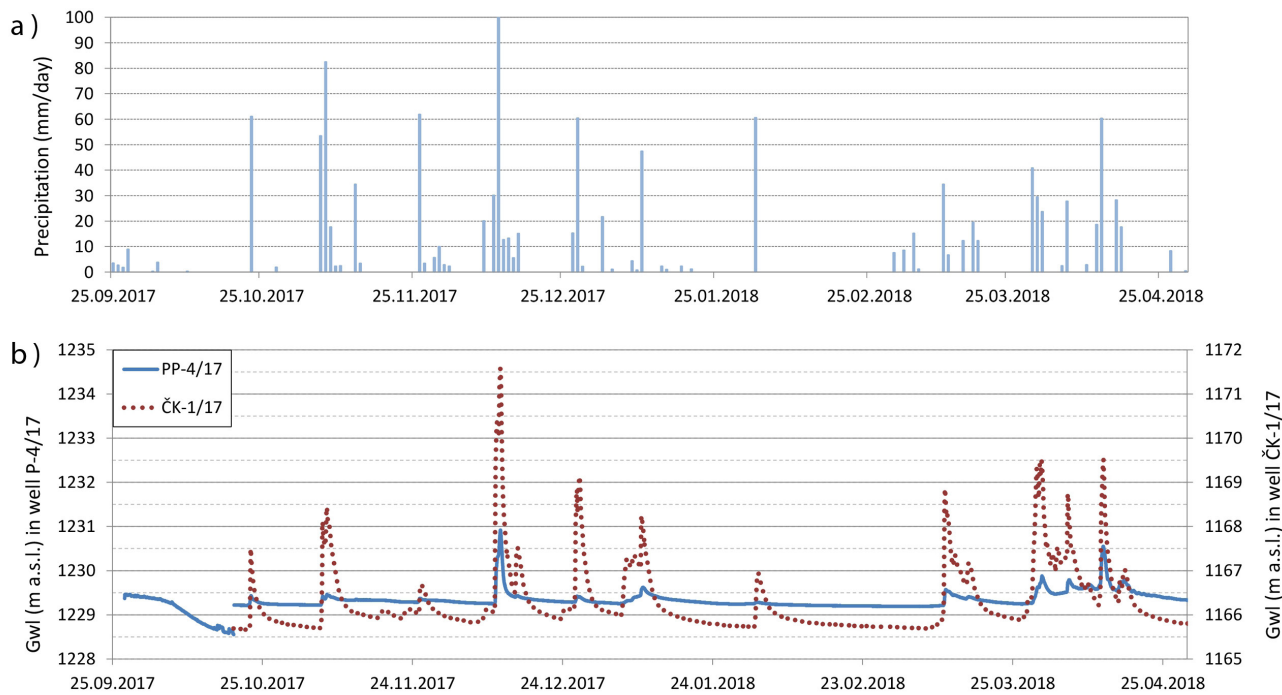


Fig. 7. a) Precipitation measured at meteorological station Planina pod Golico (ARSO, 2018b) in comparison with b) Groundwater level fluctuation in wells PP-4/17 (solid line) and ČK-1/17 (dotted line).

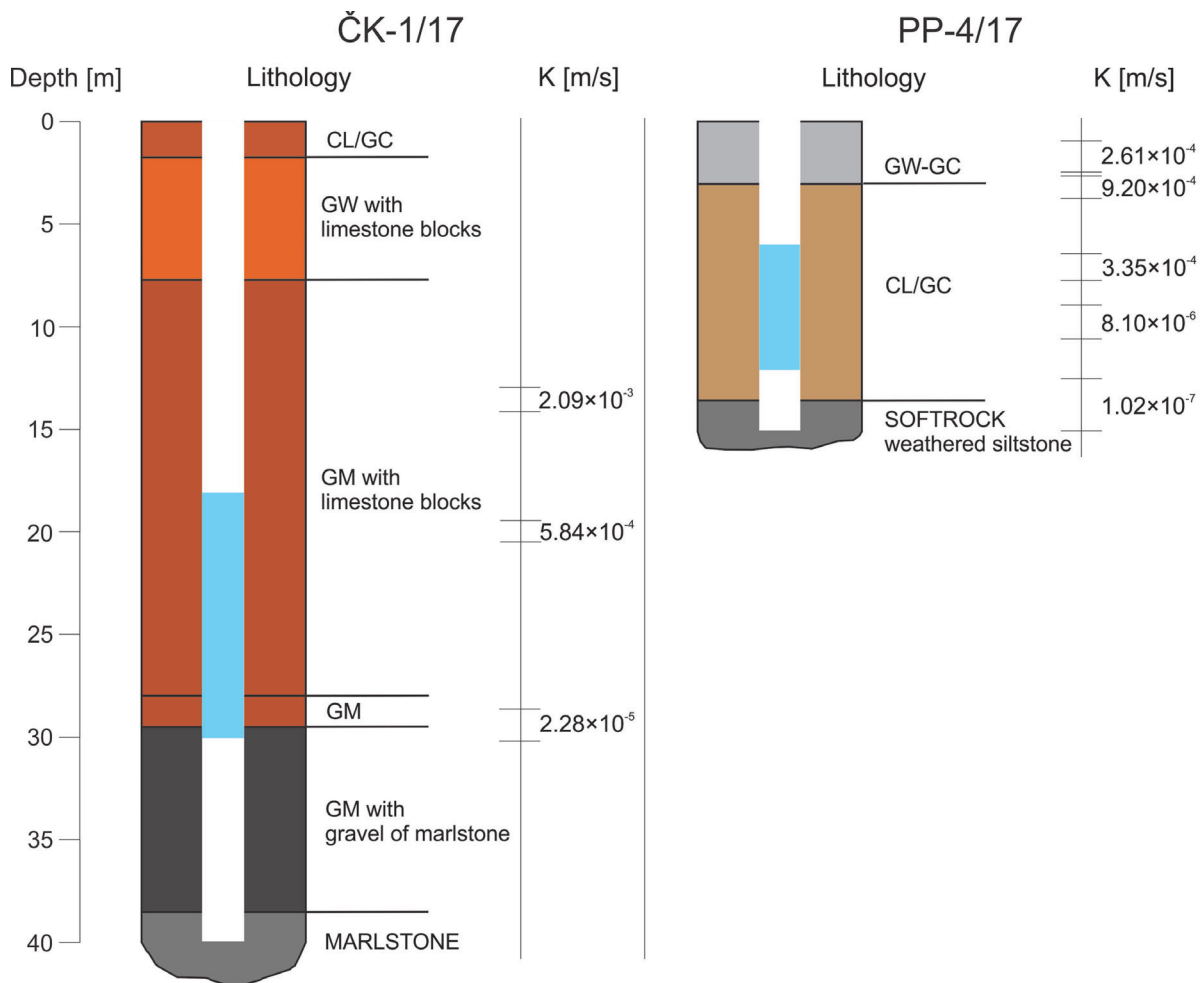


Fig. 8. Simplified lithological profiles of piezometers ČK-1/17 and PP-4/17 (segment with blue colour represents the perforated area).

Table 4. Estimated hydraulic conductivity of three sections in well ČK-1/17.

Tested section [m]	Soil type	Classification symbol	Injection step	Interpreted line slope	Unsteady K [m/s] - Hvorslev	Unsteady K [m/s] - Schneebeli	Steady K [m/s]
13.00 – 14.12	Silted gravel	GW – GM	1.		2.17×10^{-3}	2.17×10^{-3}	9.14×10^{-4}
			2.		2.64×10^{-3}	2.64×10^{-3}	/
19.50 – 20.28	Clayey sandy silt with gravel	GM - GW	1.	1.	7.83×10^{-4}	7.83×10^{-4}	/
				2.	2.39×10^{-4}	2.39×10^{-4}	/
			2.	1.	1.25×10^{-3}	1.25×10^{-3}	/
				2.	6.20×10^{-5}	6.21×10^{-5}	/
28.70 – 30.25	Silted gravel	GM	1.	1.	3.23×10^{-5}	3.23×10^{-5}	/
				2.	7.47×10^{-6}	7.46×10^{-6}	/
			2.		2.57×10^{-5}	2.56×10^{-5}	/

Table 5. Estimated hydraulic conductivity of four sections of borehole PP-4/17.

Tested section [m]	Soil type	Classification symbol	Injection step	Interpreted line slope	Unsteady K [m/s] - Hvorslev	Unsteady K [m/s] - Schneebeli	Steady K [m/s]
1.07 - 2.52	Gravel with clay and sand in lower part	GW-GC	1.	1.	3.36×10^{-4}	3.36×10^{-4}	5.40×10^{-4}
				2.	1.15×10^{-5}	1.15×10^{-5}	
			2.	1.	4.46×10^{-4}	4.46×10^{-4}	/
				2.	6.68×10^{-6}	6.68×10^{-6}	/
2.70 - 3.87	Gravel with sand and silt	CL/GC	1.	1.	1.67×10^{-3}	1.67×10^{-3}	5.72×10^{-4}
				2.	5.18×10^{-4}	5.18×10^{-4}	/
6.55 - 7.88	Gravel with gravel clay	CL/GW	1.		/	/	3.32×10^{-4}
			2.		/	/	3.37×10^{-4}
9.00 - 10.70	Clayey gravel with sand	CL/GW	1.		8.10×10^{-6}	8.09×10^{-6}	/
12.6 - 15.17	Weathered siltstone	SOFTROCK	1.		1.02×10^{-7}	1.02×10^{-7}	/

Discussion

The observed large fluctuation of discharge of springs and occasional dry outs are highly dependent on meteorological conditions, which reflects the low storage capacity of aquifers or locally limited recharge areas of the springs. A strong relation between meteorological and groundwater conditions is reflected also in the groundwater level fluctuation, observed in the wells. The fast rise of groundwater level after rainfall is observed in all wells, however, the amplitudes of fluctuation differ (figs. 6 and 7). The largest amplitude (12 m) was observed in the well PP-2/17. At this location, low permeable permo-carboniferous layer overlay the aquifer which consists of marly limestone and breccia.

It seems that the aquifer is a part of an aquifer system that has a large catchment and is partly drained out at the Urbas spring. Due to the low permeable upper layer, semi confined conditions are created.

Also continuous measurements of groundwater level fluctuation in wells ČK-1/17 and PP-4/17 show fast changes of groundwater level which are strongly related to precipitation (fig. 7). Amplitudes are higher, up to 6 m, in well ČK-1/17 than in well PP-4/17, where the maximum amplitude of groundwater level is below 2 m. These differences could be attributed to the higher hydraulic conductivity, larger thickness (Table 4 and Table 5), and probably also larger spatial extent of more permeable layers at location of well ČK-1/17.

The results of infiltrometer tests show relatively high hydraulic conductivity of the soil. Presumably, this reflects the loose structure and high content of organic matter in the top soil found in the forest which covers a big part of the study area (fig. 2b). The tests were performed at locations where the ground has not been deformed by mass movement. Due to the presence of cracks and fissures, locally higher infiltration could be expected in the landslide body.

The planning of the falling-head test in wells assumed that significant groundwater flow can be established only through more permeable layers. Consequently, falling-head test were performed in more permeable gravel sections of boreholes. Therefore, calculated hydraulic conductivities reflect properties of more permeable parts of the ground in the study area. In general, higher hydraulic conductivity is observed in the upper parts of the boreholes. Those layers are predominantly represented by gravel and have a medium hydraulic conductivity, while the lower clayey gravel parts have a low hydraulic conductivity. In the area of Čikla landslide, values of hydraulic conductivities were estimated approximately half order of magnitude higher than in the Urbas landslide. The differences in hydraulic conductivities could be attributed to the lithological composition of the ground, which, in general, shows a larger share of coarse material in upper parts of the ground and in the area of Čikla landslide. The groundwater occurrence was identified in the last tested interval (28.70–30.25 m), while the first two tested intervals were in an unsaturated zone. In borehole PP-4/17 all tested intervals were in a saturated zone.

Conclusions

The observations and hydraulic tests performed in the area above Koroška Bela settlement have shown complex and heterogeneous hydrogeological conditions, predisposed by geological and tectonic setting and active mass movements. Therefore, the observed hydrogeological environment cannot be uniformly defined. To adequately address such conditions, an approach is required which combines various hydrogeological methods, partly already performed in the presented study.

The performed investigations enabled a very rough insight into the landslide hydrogeological mechanism and provided the first data on the hydraulic conductivity of the material in the landslide masses, the groundwater level, the infiltration capacity of the ground, the occurrences of

the springs and their discharges. However, still many uncertainties exist about the hydrological processes occurring in observed landslides. In order to evaluate the role of groundwater and hydrogeological processes on landslides movements continuation of hydrogeological monitoring (groundwater level and temperature measurements) and additional investigations (e.g., hydrogeochemical and isotope analysis) for better defining recharge mechanism and groundwater flow patterns in the landslide bodies and their catchment areas are proposed.

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