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Measuring and modelling water transport on Skaftafellsheiði, Iceland

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Abstract

Areas with thick basaltic aquifers are used for drinking water supply and irrigation purposes, such as the Columbia River Basalt group in northwest USA and the Deccan Traps in India. However, rainfall-runoff processes in these basaltic areas are poorly understood. Cooling joints can transport large amounts of water, but – due to their limited porosity – they are vulnerable for over-abstraction. On Iceland, in the small Skaftafellsheiði basaltic catchment (4 x 6 km), field data were collected in 2014 and 2015. Two small streams discharge the rain surplus. Precipitation was measured at various elevations on the ridge. Also, the discharge of the streams was measured. A groundwater flow model was constructed in order to get more insight in the physical properties of the basalt aquifer and its rainfall-runoff properties. The field experiments showed that precipitation increases linearly with surface elevation. On average, the precipitation at 800 m+msl was almost double, relative to the precipitation at 200 m+msl. Calculated ET_{pot} was rather high, due to the 19 potential sun hours per day during the Icelandic summer. Field experiments revealed quick discharge response on rainfall events, but also rather constant base flow during dryer periods. This indicates a limited infiltration capacity, but also a considerable storage capacity in the subsequent layers. The peat layer is believed to be the dominant storage/reservoir. Peat, regolith and an organic layer formed the top layer in the GMS-Modflow groundwater model. The thick basaltic aquifer was split in a series of model layers. Best results were obtained by using a decreasing hydraulic conductivity to depth. The transient model overestimated the groundwater levels at the outlet, but managed to reproduce the wet/dry conditions in the catchment rather well. This indicates that it is possible to model complex basaltic aquifers, by taking a large Representative Elementary Volume (REV) as starting point.

Keywords: *Basalt aquifers, MODFLOW, modelling, Representative Elementary Volume*

Rezumat. Măsurarea şi modelarea transportului de apă în Skaftafellsheiði - Islanda

Zonele cu bazalte acvifere groase sunt folosite în scopuri de alimentare cu apă potabilă și de irigare, cum ar fi Grupul Columbia River Bazalt în nord-vestul SUA şi Deccan Traps in India. Oricum, procesele de precipitații-scurgere în aceste zone bazaltice sunt insuficient înțelese. Fisurile de răcire pot transporta cantități mari de apă, dar - datorită porozității lor limitate - acestea sunt vulnerabile pentru captări in exces. In Islanda, în micul bazin hidrografic bazaltic Skaftafellsheiði (4 x 6 km), datele de teren au fost colectate în 2014 și 2015. Două râuri cu debite mici au tranzitat ploaia netă. Precipitația a fost măsurată la diferite altitudini pe cumpănă de apă. De asemenea, au fost măsurate şi debite de apă din râuri. Un model de curgere a apelor subterane a fost elaborat pentru a obține o imagine mai în detaliu asupra proprietăților fizice ale acviferului bazaltic și a proprietăților procesului precipitații-scurgere. Experimentele de teren au arătat că precipitațiile cresc liniar cu înălțimea suprafeței. În medie, precipitațiile la 800 m + MSL au fost aproape duble, în raport cu acela de la 200 m + MSL. ET_{pot} calculată a fost destul de mare, din cauza celor 19 ore potențiale de soare pe zi din timpul verii islandeze. Experimentele pe teren a relevat un răspuns rapid al debitelor asociate unor ploi şi constanța scurgerii de bază, în timpul perioadelor uscate. Aceasta indică o capacitate de infiltrare limitată, dar și o capacitate de stocare considerabilă în straturile subsecvente. Stratul de turbă este considerat a fi dominant în cantonarea resurselor de apă. Turba, regolitul și un strat organic format stratul superior au fost rulate prin programul de modelare a apelor subterane GMS-Modflow. Depozitele acvifere bazaltice au fost modelate prin metoda multistrat. Cele mai bune rezultate au fost obținute considerând scăderea conductivității hidraulice cu adâncimea. Modelul tranzitoriu a supraestimat nivelurile apelor subterane de la priză, dar a reușit să reproducă destul de bine condițiile umed/uscat din bazinul hidrografic. Acest lucru indică faptul că este posibilă modelarea complexă a acviferelor bazaltice, plecând de la un Volum Elementar Reprezentativ (VER) extins.

Cuvinte-cheie: *acvifere bazaltice, MODFLOW, modelare, Volum Elementar Reprezentativ*

Introduction

Areas with extensive and thick basaltic aquifers are used for drinking water production and irrigation purposes. Examples of such regions are the Columbia River Basalt Group in northwest USA and the Deccan Traps in India.

The relation between rainfall, storage and runoff in such basaltic regions is often poorly understood, with over-abstraction and declining groundwater levels as a result (Macdonald et al., 1995, Taylor and Howard, 1998; Hancox et al., 2010; Liu, 1998). Cooling joints in basalt formations can transport large quantities of water, but have a limited porosity.

Domenico & Schwartz (1998) showed that the representatives of physical parameters like porosity and hydraulic conductivity are highly dependent of the sampled volume (Figure 1). Only when the sample includes multiple flows, representative values can be obtained, i.e. the Representative Elementary Volume (REV).

In areas where intensive water abstractions for drinking water production and irrigation purposes led to significant water level decline, attempts have

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been made to measure the water balance terms (precipitation, actual evapotranspiration, storage change, discharge) in relation to the water abstraction (Candel, 2014; Candel et al., 2016).

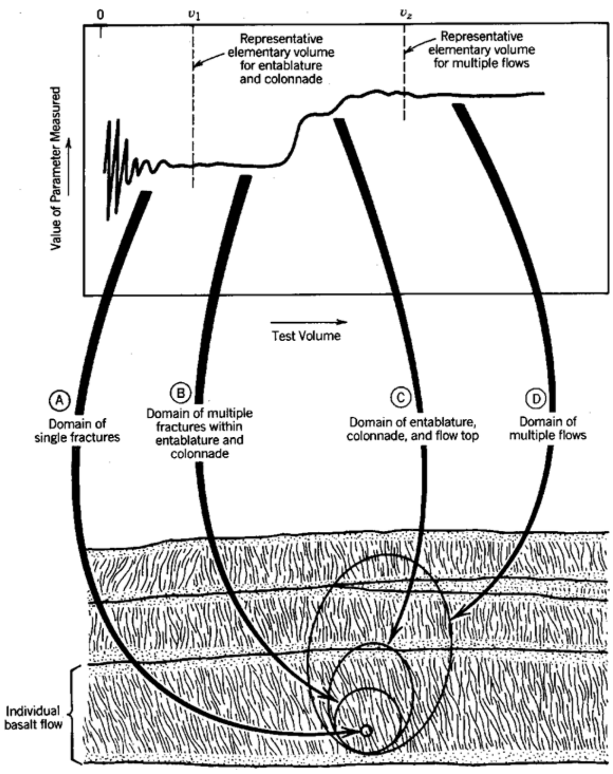

Fig. 1: Representative Elementary Volume in basalt (Domenico & Schwartz, 1998)

In order to calculate scenarios for future water use, often groundwater models are developed for these regions.

A major problem for groundwater modelling is the limited accessibility of the basalt aquifers for measuring its physical properties. Therefore, field experiments were performed at Skaftafellsheiði, southeast Iceland. Skaftafellsheiði is a small basaltic catchment (4 x 6 km), (Pyatt & Ditcham (1983), with elevation levels ranging from 1000 m+msl in the north to 100m+msl in the south (see Figure 2).

The basalt layers origin forms a periglacial period in Late-Tertiary (4-5 MA ago). The flows are interbedded with some sedimentary interbeds (Helgason & Duncan, 2001). During Pleistocene deep gullies were formed through glacial erosion, leaving behind an isolated 1800 m high basaltic ridge with multiple basalt flows. During maximum glaciation, not only the valleys were filled with ice, but also the ridge was almost completely covered by the glacier. The ridge has two main streams which did cut into the upper basalt layers during Holocene. This explains the morphology of the basalt formations, with its rounded top, steep slopes towards the glacier valleys and V-shaped stream incisions. The basalt formations are locally covered with some thin organic layers (peat) and/or some weathered basalts (regolith layer). This makes the area suitable for fundamental research on water flow through basalt formation.

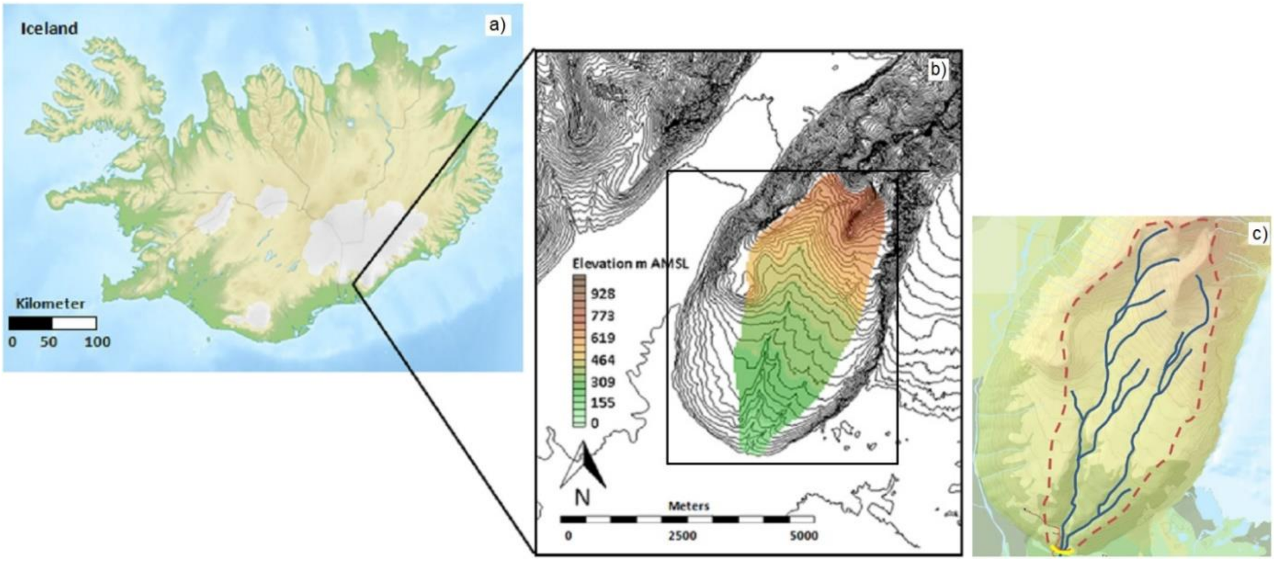

Fig. 2: Map of Iceland (a), elevation map of Skaftafellsheiði and surroundings (b) (Kooi, 2015), adjusted from LMI (2014) and catchment boundaries and streams on Skaftafellsheiði (c)

Field campaign setup

Iceland is known for its pronounced orographic effect on rainfall distribution (VEDUR, 2016). Rainfall ranges from > 4,000 mm·y^{-1} at the upwind side of topographic obstructions (volcanoes, glaciers) to > 400 mm·y^{-1} at downwind regions. In order to get a view on all the water balance components, thus also on the spatial distribution of precipitation on the Skaftafellsheiði, a set of 7 rain gauges (totalizers) was installed (see Figure 3).

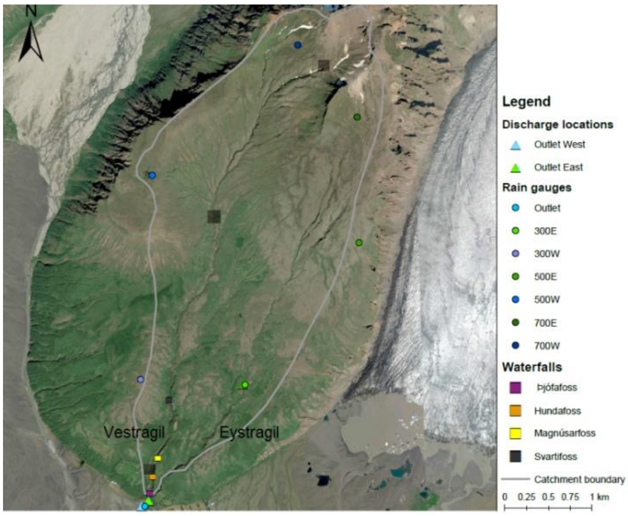

Fig. 3: Locations of rain and discharge measurements sites in the Skaftafellsheiði catchment - letters refer to west (W) and east (E) and numbers to elevation in m+msl; background image adjusted from GoogleEarthPro by Avis (2016)

One gauge was installed at the outlet in the south tip of the catchment, at an elevation of 100 m+msl, and pairs of rain gauges at 300, 500 and 700 m+msl at the eastern stream (Eystragil) and western stream (Vestragil), respectively. All gauges were measured at a daily basis.

Discharge was measured at the outlet of both streams, by using automatic pressure transducers with a measuring interval of 15 minutes. Manual discharge measurements were done at a daily basis. In the highest part of the catchment, snow was present. Since snow melt contributes to the discharge a time lapse camera was installed, imaging every 30 minutes.

Snow coverage was combined with a digital elevation model to estimate the snow volume. A density of 0.6 m^3 water per m^3 snow was used for conversion snow to water equivalent.

Evapotranspiration was derived by using the Hargreaves-Samani equation (Hargreaves & Samani, 1982; Samani, 2000). Reference evapotranspiration (ET_{ref}) was translated to potential evapotranspiration (ET_{pot}) for the different vegetation zones, i.e. deciduous forest, grassland and peat/moss/shrubs. The deciduous forest can be found at the elevation zone 100 – 200 m+msl, and consists predominantly of small birch (*Betula*) trees.

In the range 200 – 400 m+msl the vegetation gradually changes to dwarf shrubs and heathland. Above 400 m+msl an alpine zone is found, with peat, moss and dwarf shrubs.

Table 1 shows the parameters used to get to ET_{pot}.

Table 1: Evapotranspiration parameters related to land cover

Vegetation	Kt (-)	Coverage (-)	Average maximum height (m)	Dominant species
Deciduous forest	0.9	0.9	2	Birch,willow
Peat (willow)	0.9	0.8	0.3	Willow,moss
Peat (bog)	0.9	1	0.3	Moss, grass
Peat (forest)	0.9	1	1.8	Willow, grass
Peat (grass)	0.9	1	0.1	Grass
Moss	0.15	0.6	0.1	Moss, crowberry
Regolith vegetated (west)	0.5	0.6	0.3	Moss, heather, willow, birch
Regolith vegetated (east)	0.4	0.3	0.2	Moss,willow, birch
Regolith bare	0.2	0	0	-

Based on Allen et al. (1998) it was chosen to use an equation with a single crop factor, which was corrected for the relatively large portion of bare rock, see equation 1 and 2.

$$ET_{act} = f_c \times K_t \times ET_0 \quad (1)$$

$$ET_{zone} = ET_{act} + (1 - f_c) \times K_{t,bare\ rock} \times ET_0 \quad (2)$$

where:

ET_{act} = actual evapotranspiration of the vegetation (mm·d^{-1}); K_t = crop factor (-)
ET_0 = reference evapotranspiration (mm·d^{-1})
F_c = vegetation coverage (-)
ET_{zone} = actual evapotranspiration of the vegetation, compensated for bare rock (mm·d^{-1})

An indicative water balance was constructed, with for P and ET a 20 m grid. For the period 19 Jul – 3 Sep 2015 the following water balance was used (Equation 3).

$$\Delta S = P + M - Q - ET \quad (3)$$

where:

ΔS = storage change (m^3)
P = precipitation (m^3)
M = snow melt (m^3); Q = discharge (m^3)
ET = evapotranspiration (m^3)

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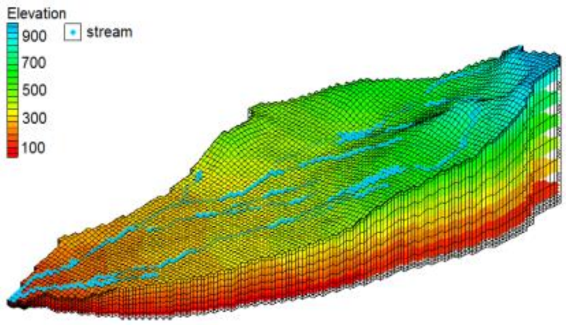
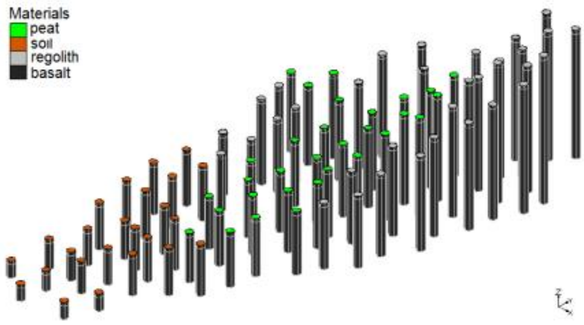
Modelling setup

The model boundary was topography based (LMI, 2014). The western model boundary was a bit uncertain, since many small streams discharge water over the steep western slope towards Morsádalur. Nevertheless, for all boundaries no-flow Neumann boundary conditions were applied. The base of the model was set to 90 m+msl, which is 10 m lower than the outlet level (100 m+msl). The grid consists of cells of 50x50 m and has 10 model layers. See Figure 4 and Table 2 for the model setup and hydrogeological properties of the materials. The basalt was divided in 7 model layers. These are not actual basalt flows, but enabled us to decrease the hydraulic conductivity (m·d^{-1}) to depth.

Hydraulic conductivity for fractured and weathered basalts varies between 10^{-4} and 10^{2} m·d^{-1}, where the highest conductivities were found in Hawaian basalt formation (Singal & Gupta, 2010). Some hydraulic conductivities were derived from fracture desity and fracture with measurements at Svartifoss waterfall and Reynisfjara. The found values of 12 and 16 m·d^{-1} for the horizontal and vertical hydraulic conductivity (k_h and k_v, respectively) are within the abovementioned range. Due to columnar basalt jointing, strong anisotropy occurs, with within the basalt formation a relatively low k_h and a relatively high k_v. For the upper layers, k_h was set to 0.3 m·d^{-1} (Kooi, 2015). Due to a higher pressure on the fractures and therefore a smaller fracture with, the k_h and k_v of the deeper basalt layers was set to 0.01 m·d^{-1} (Singal & Gupta, 2010). At the southern tip of the model the top basalt layer (basalt 1) the k_h was increased to 12 m·d^{-1}. This is in line with the values found at Svartifoss. This adjustment resulted in a reduction of flooded cells.



(a) Elevation (m+msl) and layer distribution.



(b) Materials added to boreholes.

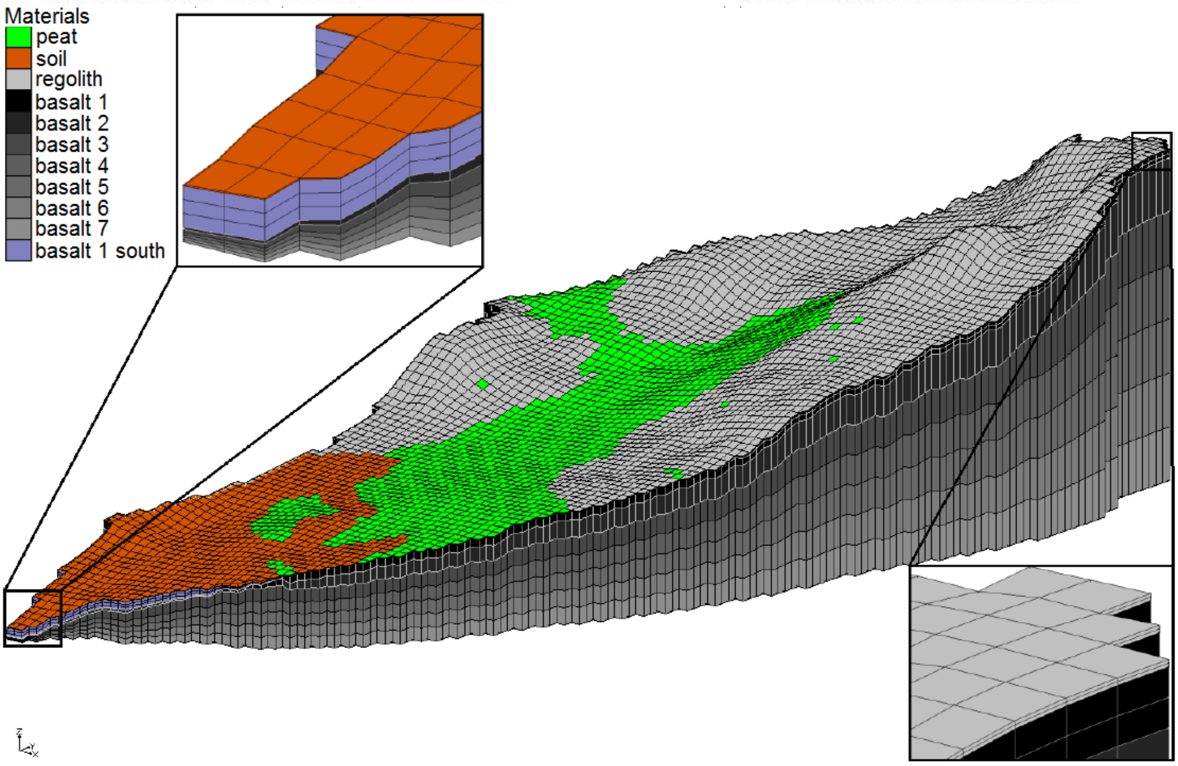


Fig. 4: MODFLOW model setup

Table 2: Materials and physical properties of model layers

Material	Thickness (m)	Horizontal hydraulic conductivity (m·d^{-1})	Vertical hydraulic conductivity (m·d^{-1})	Specific storage (m^{-1})	Specific yield (-)
Organic soil	0.5	5	5	$1\ 10^{-6}$	0.08
Peat	2 – 4	5	5	$1\ 10^{-6}$	0.2
Regolith	2	10	10	$1\ 10^{-6}$	0.4
Basalt 1	1 – 134	0.3	10	$1\ 10^{-6}$	0.4
Basalt 2	1 – 134	0.3	0.3	$1\ 10^{-6}$	0.4
Basalt 3 – 7	1 – 134	0.01	0.01	$1\ 10^{-6}$	0.4

Field campaign results

In Figure 5 the relation between precipitation (mm) and discharge at the outlet of both streams (m^3·s^{-1}) is given. Water level was measured by automatic pressure transducers and by manual measurements, with automatic 3 and 4 measuring the water level at Eystragil and automatic 1 and 2 measuring water level at Vestragil. Water levels were converted to discharge (Q in m^3·s^{-1}) through a rating curve.

The first part of the short intensive measurement campaign was relatively dry. Discharge hardly responded to rainfall events, indicating sufficient storage capacity. From 12 August 2015 onwards a wet period started. First this led to small peaks and an increase of baseflow, but the rainfall events on 22 and 23 August 2015 caused a significant discharge peak. Apparently the storage capacity of the catchment was exceeded at that point.

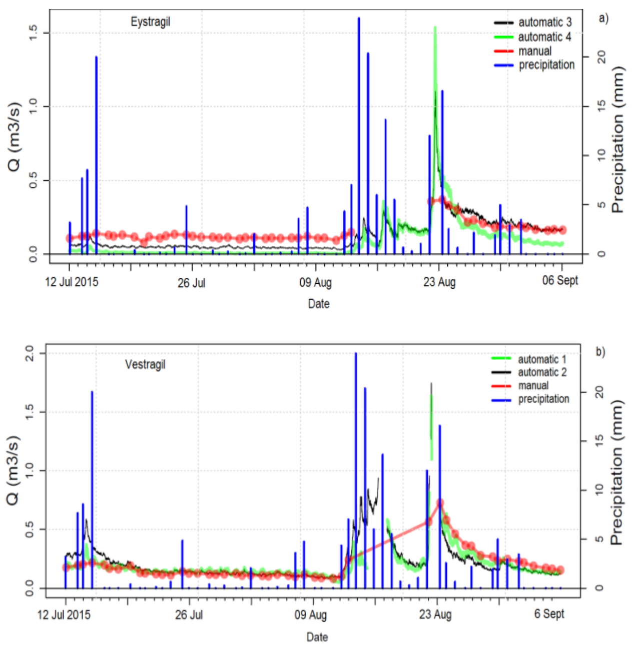


Fig. 5: Precipitation and discharge of: Eystragil – eastern stream (up) and Vestragil (western stream) at the catchment outlet (down)

A significant relation between precipitation and elevation was found. The linear relation appeared to be slightly steeper in the Eystragil catchment than in the Vestragil catchment, see Figure 6. This led to a linear relation for the whole catchment: equation 4.

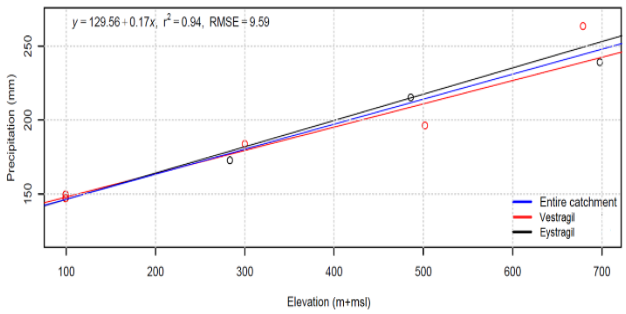


Fig. 6: Relation between elevation and precipitation sums in the period 12 Jul – 6 Sep 2015 for the entire catchment and both sub-catchments

$$P = 0.17H + 129.56 \qquad (4)$$

where:

P=total precipitation in the measurement period (12 Jul – 6 Sep 2015; mm);
H=Elevation (m+msl)

Evapotranspiration was calculated for different vegetation types at 500 m+msl and for deciduous forest in mm·d^{-1} for the range 100 – 1100 m+msl. Figure 7 and 8 show the results of these calculations, respectively. In Figure 7 peat (bog) and peat (forest) are not visible because they coincide with peat (grass). Different vegetation type/land cover combinations resulted in distinct differences in ET_{act}. Peat (grass/bog/forest) proved to have the highest ET_{act} and moss the lowest, with ET_{act} values three times less than peat. The relatively high ET_{act} values in July are due to the combination of relatively high temperatures and long day light. The differences are getting smaller as the day length shortens and the temperature drops.

ET_{act} averaged for the total period, vegetation types and elevations is 1.4 mm·d^{-1}. Averages of ET_{ref} of 3.5 and 2.6 mm·d^{-1} were found for July and August 2015, representative for an elevation of 100 m+msl. Einarsson (1972) reported an ET_{ref} of 3.1 and 2.2 mm d^{-1} for July and August at Kirkjubæjarklaustur, respectively. This indicates that the calculated ET_{ref} and ET_{act} is a reasonable estimation for the catchment.

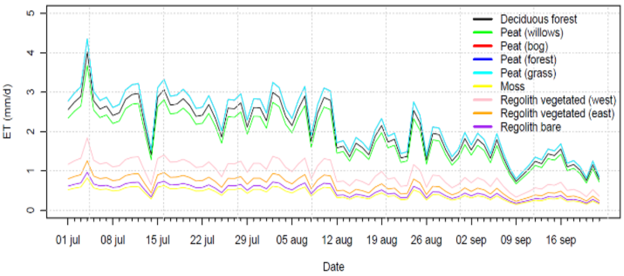


Fig. 7: ET_{act} at 500 m+msl for each vegetation type in combination with underlying material

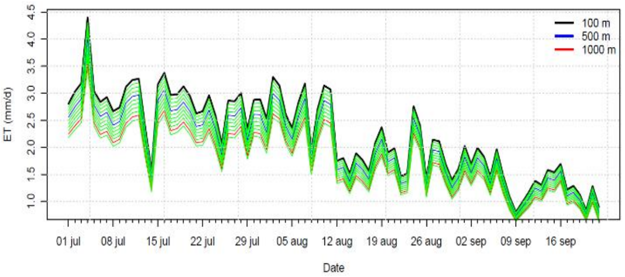


Fig. 8: Calculated Et_{act} for deciduous forest for every 100 m elevation increase

Table 3 shows the simplified water balace for the catchment. Under the (debatable) assumption of accurate values for all balance terms, it can be concluded that there was a small change in storage.

Surface runoff can occur in the catchment, caused by major precipitation events. However, surface runoff feeding the discharge at the outlet seems to be a minor portion of the total.

Table 3: Water balance, period 19 Jul – 3 Sep 2015

Water balance (elements)	P	M	Q	ET_{act}	ΔS
10^6 m^3	2.05	0.24	1.49	0.66	0.14
Values (mm)	206	25	150	67	14

Note: P = precipitation; M = snow melt; Q = discharge; ET_{act} = Actual evapotranspiration of the vegetation; ΔS = storage change.

Modelling results

Steady state modelling results in discharges at the outlet for Vestragil and Eystragil of 22 400 and 7 100 $m^3 \cdot d^{-1}$, respectively, where the field measurements result in average discharges of 18 500 and 13 100 $m^3 \cdot d^{-1}$. Measured and calculated discharges differ per stream, but the total calculated discharge (29 500 $m^3 \cdot d^{-1}$) is in line with the total measured discharge (31 600 $m^3 \cdot d^{-1}$).

The hydrographs of Eystragil and Vestragil show that overland flow can occur during (very) wet conditions, but the relative portion seems to be smaller than the inaccuracies of the other water balance components.

Figure 9 shows the calculated head distribution.

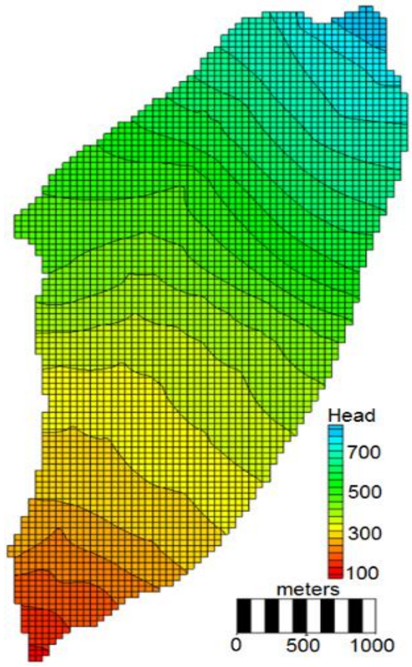


Fig. 9: Head distribution in the model area, using the measured hydraulic conductivities

Fractured rock aquifers often show a large degree of heterogeneity (Wellman & Poeter, 2005). This heterogeneity is dependent on scale (see Figure 1).

When the sampling domain comprises a domain of multiple flows (scale D in Figure 1), then a representative parameter value can be found. At this scale, no detailed information about fracture orientation, fracture density, fracture aperture, degree of connectivity and smoothness of fractures is required, since they are inside the sample volume. At that point, Darcy's law can be applied to describe the water flow.

Conclusion

When discharge observations in streams are combined with groundwater head observations, it is possible to calibrate a rather simple groundwater flow model and obtain reasonable hydraulic conductivity values for the basalt layers. It must be noted that the calibrated hydraulic conductivities depend on how the basalt layers are implemented.

In this study, the model layers do not reflect single basalt layers of sub-aquifers, but represent multiple flows and thus resemble representative elementary units.

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