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Hydrological processes in small catchments of mountain headwater lakes: The Tatra Mountains

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Abstract: This study evaluates runoff and different methods for the estimation of water balance and runoff genesis in four small alpine catchments, which lie outside the standard network of hydrological and climate networks. These test catchments, whose size ranges between 2.3 and 110 ha, are located above the timberline at elevations between 1,784 and 2,380 m. Their land surfaces consist of lakes, rock formations, debris deposits, and alpine meadows. Hydrological data were collected for the water year 2001. The catchments were instrumented by three automatic weather stations recording global and net solar radiation, air temperature, humidity, precipitation, and soil temperature. Lake water levels were registered with staff gauges and runoff evaluated from water storage calculations. Runoff genesis was investigated by means of chemical tracers (Rhodamine WT and Lithium chloride). Hydrological process estimations were made using standard methods including: the input of precipitation and snowmelt, both potential and actual evaporation, which was estimated by the approaches of HAMON, PRIESTLEY-TAYLOR, PENMAN and GRINDLEY, and runoff calculated from the lake storage, were compared with results of the conceptual hydrological BROOK90 model. The empirical results show that hydrological processes are governed by the temperature-dependent regime of high mountain snowmelt. However, the major differences in water yield and runoff genesis between watersheds were due to differences in the morphologies of the lakes and their basins, the soil-vegetation complex, and runoff routing. Evaluating approaches to estimation, PENMAN's combination of both aerodynamic and energy balance method provides the best fit to observed data with observed evapotranspiration being 78 to 99% of the potential calculated. The deterministic BROOK90 model is effective for precipitation-runoff genesis studies in small headwater catchments. In the Ľadové pleso basin predicted and observed water yield show close correlation. The annual sum of actual evapotranspiration calculated by BROOK90 (352 mm) corresponds closely to that estimated by the approach of Penman (386 mm).

Key words: Alpine catchments, precipitation, evaporation, runoff genesis, water balance, Slovakia, Poland.

Introduction

Alpine regions are supposed to represent the least disturbed environment in Europe. Generally, there is relatively low direct human impact. Alpine environments face several global impacts: atmospheric pollution, acid atmospheric deposition and climate change. In the last decade, EU funded projects (ALPE, MOLAR and EMERGE) have focussed, particularly, on geo-chemical and biological studies in remote alpine lakes (FOTT et al., 1994).

High altitude watersheds are critical for the recharge of water resources (the mountains are supposed to be water towers of the world), and alpine headwater systems are strongly related to environmental security (HAIGH & KŘEČEK, 2000). Water is the transport medium of solid particles, chemicals, and biota, so alpine lake environments are controlled by hydrological processes and interactions within lakes and their catchments. However, the landscape complexity typical of alpine catchments means that accurate empirical investigation demands a dense network of precipitation and stream gauging stations. Unfortunately, often, these distant mountain areas are considered marginal to national meteorological and hydrological networks and these kinds of investments in instrumentation and monitoring are unusual.

The aim of this paper is to contribute to the better

Table 1. Morphology of studied basins.

Basin	Basin area (ha)	Lake area (ha)	Elevation range (m a.s.l.)	Prevailing face	Land slope (%)	Basin order	Shape index A/L^2
Starolesnianske pleso	2.26	0.73	1986–2030	NE	65	0	0.45
Ľadové pleso	13.80	1.72	2057–2350	S	69	1	0.34
Długi Staw Gąsienicowy	64.84	1.59	1784–2301	N	62	2	0.63
Nižné Terianske pleso	110.01	4.91	1941–2428	NW	58	2	1.05

Table 2. Land cover in watersheds.

Basin	Lake		Grassland		Rock formations		Debris cones	
	(ha)	(%)	(ha)	(%)	(ha)	(%)	(ha)	(%)
Starolesnianske pleso	0.73	32	0.69	31	0.46	20	0.38	17
Ľadové pleso	1.72	13	3.60	26	3.04	22	5.44	39
Długi Staw Gąsienicowy	1.59	2	12.65	20	37.95	58	12.65	20
Nižné Terianske pleso	4.91	4	31.53	29	42.04	38	31.53	29

Table 3. Water balance at the nearest stream gauging stations: mean annual values of precipitation (P_a), runoff (R_a) and evapotranspiration ($ET = P_a - R_a$).

Stream	Profile	Evidence number	Basin area (km^2)	P_a (mm)	R_a (mm)	ET (mm)
Biely Váh	Nad Mlyničnou vodou	4-21-01-030	34.8	891	571	320
Poprad	Nad Malým Popradom	3-01-02-013	48.3	1285	865	420
Biela Voda	Nad Javorinkou	3-01-01-001	77.5	1459	1041	418

understanding hydrological processes in the high altitude catchments of the High Tatra Mountains (Mts). Since 2000, water balance and runoff genesis have been studied in four previously un-gauged headwater catchments which lie outside the standard hydrological network.

Material and methods

The research site

Four investigated catchments are located in the High Tatra Mts forming the highest massif in the Carpathians. The basins of lakes Ľadové pleso, Nižné Terianske pleso, and Starolesnianske pleso are situated in Slovakia, and the basin of the lake Długi Staw Gąsienicowy in Poland. The latitude of catchments varies from 49.17 to 49.22 N, longitude from 20.01 to 20.17 E, elevation from 1784 to 2380 m a.s.l., and catchment area between 2.3 and 110 ha (Tab. 1).

The geology is formed by crystalline bedrock with shallow podsolic soils (depth varies from 0.2 to 0.6 m) with high content of stones. All those watersheds are located above the timberline. The land surface consists of lakes, alpine meadows, rock formations and debris cones (Tab. 2). The percentage of alpine meadows in watersheds varies from 20 to 31% of the watershed area. The vegetation is represented by dry tundra with dominant species of *Calamagrostis villosa, Festuca picta*, and *Luzula luzuloides*) with patches of dwarf pine *(Pinus mugo)*. The percentage of rock (bare or covered with lichens, particularly, *Rhizocarpon, Acarospora oxytona*, and *Dermatocarpon luridum*) increases with the elevation. Grazing or other land-use activities have been prohibited since the early 1950s, when the National Park of the Tatra Mts was proclaimed.

The investigated area can be characterised by the climate type 'Dfc' of the Köppen classification system (Henderson-Sellers & Robinson, 1989). In the region of the High Tatra Mts, Lajczak (1996) reports negative gradient of mean annual air temperature of –0.6 °C and positive gradient of the mean annual precipitation of 50 mm per 100 m of elevation, respectively. Thus, the estimates of mean annual air temperature over the investigated area might vary from –2.6 to 1.6 °C, and precipitation from 1200 to 1550 mm. However, precipitation sums are generally higher in the northern part of the mountains, and some valley can reach more than 2000 mm per year. Tomlain (1985) reported the range of mean annual evapotranspiration in the High Tatra Mts from 250 to 300 mm by the negative gradient of –18 mm per 100 m of elevation.

The basins of lakes Długi Staw Gąsienicowy, Ľadové pleso and Starolesnianske pleso belong to headwaters of the Visla River, the basin of the Nižné Terianske pleso is drained by the Váh River. In all, the drainage channel network is poorly developed. The basins of Długi Staw Gąsienicowy, Nižné Terianske pleso and Starolesnianske pleso are drained by both seepage and complicated stream channel system, while the basin of Ľadové pleso is drained only by the lake seepage.

Historically, the standard network of precipitation and stream gauging stations in the High Tatras has been focussed on watersheds larger than 40 km^2 (HMU, 1968). Therefore, the research sites are included only in the hydrological analyses of larger watersheds (Tab. 3). Extrapolating the water balance into smaller areas, thus, in elevations of

Table 4. Parameters observed at the automatic weather stations.

Watershed Weather Station	Ľadové pleso NOEL 2000		Długi Staw Gąsienicowy NOEL 2000		Nižné Terianske pleso DELTA-T-LOGGER	
Parameter	Height (m)	Interval (h)	Height (m)	Interval (h)	Height (m)	Interval (h)
Global radiation	5	0.5	5	0.5	5.0	0.5
Reflected radiation	–	–	–	–	5.0	0.5
Net radiation	–	–	–	–	5.0	0.5
Air temperature	5	0.5	5	0.5	3.5	0.5
Humidity	5	0.5	5	0.5	3.5	0.5
Air pressure	–	1.0	–	1.0	3.5	1.0
Wind speed	10	0.5	10	0.5	5.0	0.5
Wind direction	10	0.5	10	0.5	5.0	0.5
Soil temperature	–0.02	0.5	–0.02	0.5	–0.02	0.5
Precipitation	4.5	0.5	–	–	2.0	24.0
Water level	–	0.5	–	0.5	–	1.0

1,800–2,000 m, the mean annual values of precipitation (P_a) and evapotranspiration (ET) are characterised by the range of P_a = 1800–2000 mm, and ET = 150–200 mm (DUB & NĚMEC, 1969). Later, for smaller basin of the Studený potok brook (profile Nad Tatranskou Lesnou, basin number: 3-01-02-056, area of 18.5 km^2), VOLOŠČUK et al. (1994) reported the runoff coefficient k_o = 0.92 by the annual evapotransipration ET = 150 mm.

Instrumentation

Investigated catchments were instrumented in the year of 2000, and the detailed observation was carried out in the hydrological year of 2001 (1 October 2000 – 30 September 2001). Three automatic weather stations were installed in the neighbourhood of the lakes Ľadové pleso, Długi Staw Gąsienicowy (NOEL 2000) and Nižné Terianske pleso (DELTA-T-LOGGER) in the elevation between 1,784 and 2,057 m a.s.l. Parameters of global radiation, air temperature and humidity, air pressure, soil temperature, wind speed and direction, rainfall (tipping-bucket recorder, 200 cm^2) and water level at the lakes were registered in 0.5, 1.0 or 24 hour intervals (Tab. 4). The reflected and net radiation was observed only at the Nižné Terianske pleso. All the observed meteorological parameters represent near the ground climate above the grass-rock formations. Meteorological data (including net radiation above grassland) of the nearest standard weather station of Skalnaté pleso (elevation of 1,762 m) were also provided.

Additional 8 storage rain gauges (315 cm^2) were installed in the studied catchments to be collected weekly in the summer and monthly during the winter. The rain gauges exposed in the field were not provided with windshields. In the watersheds, the snow-pack (depth, density and water equivalent of snow) was observed in 20 m intervals following the snowlines (from 1 to 3) across the basin area.

The water level in the lakes was registered by stages (vertical staff gauge) and pressure gauges registered at the automatic weather stations. For each lake, the volume-depth (bathymetric) curve was plotted, and, than, the volume increment per time estimated (LJUNGGREN, 2002).

The registration of catchment outflows by traditional stream-flow gauging is almost impossible because of complicated morphology and not fully developed stream channels. Therefore, tracers were applied to investigate the genesis and pathways of runoff in those small headwater basins. Tracers of Rhodamin WT and Lithium chloride (LiCl) were applied into the lakes of Ľadové pleso (24 October 2000 and 19 July 2001) and Starolesnianske pleso (24 October 2000). The tracers were applied from the boat over the water surface. After the application, concentrations of both tracers in the lakes were monitored (depth of 0, 3, 8, 13 m under the water level, and 0.5 m above the bottom) in daily, weekly and later in two-week intervals.

Data processing

Precipitation

Precipitation sums in hourly, daily and monthly intervals were evaluated by the hypsometric method – an extrapolation of the observation network with elevation (SHAW, 1991). The gauges exposed in the field were not provided with windshields. Therefore, the wind error was compensated by coefficients k_r = 1.16 (collected rain), and k_s = 1.65 (collected snow) according to the wind speed observed in the range of 2–4 (m s^{-1}) (LINSLEY et al., 1975).

Possible additional forms of precipitation (dew, occult fog-drip, cloud interception etc.) were not studied in the investigated catchments. In the year of 1998, STRUNECKÝ (1988, in LOPATOVÁ, 2003) reported the annual sum of additional occult precipitation by only 25 (mm/year) for grassland (mean grass height of 0.30 m) in the neighbouring basin of the Skalnaté pleso. Later, LOPATOVÁ (2003) found only 39 days with a significant fog-drip at the standard weather station "Skalnaté pleso" (1,766 m a.s.l.) during the last four years. From the above results we can consider the additional sum of precipitation in the year of 2000/2001 by an approximate range of 20–30 mm.

Evapotranspiration

In the vegetation period, the estimates of potential evapotranspiration (EP, mm per a unit of time) were calculated by empirical formulae of HAMON (temperature-based, *E-HAM*), PRIESTLEY-TAYLOR (radiation-based, *E-PRT*) and PENMAN (combination of both aerodynamic and energy balance method, *E-PEN*) described in CHOW et al. (1988) and SHAW (1991):

$$EP - HAM = 2.98Ne_d/(T_a + 273.3) \quad (1)$$

$$EP - PRT = 1.3\Delta/(\Delta + \gamma)H_T \quad (2)$$

$$EP - PEN = \frac{(\Delta/\gamma)H_T + E_{at}}{(\Delta/\gamma) + 1} \quad (3)$$

where, N – maximum possible daily sunshine (hours/day), T_a – mean daily air temperature (°C), e_a – vapour pressure in the atmosphere (mm), e_d – saturated vapour pressure (mm), Δ – slope of 'saturated vapour pressure – air temperature' curve (mm/°C), γ – psychrometric constant (mm/°C), u_2 – wind speed at 2 m above the vaporizing surface (m s^{-1}), H_T – evaporation equivalent of net radiation (mm), E_{at} – evaporation equivalent of aerodynamic conditions (mm) expressed by equation

$$E_{at} = 0.35(1 + u_2/u_2 100)(e_a - e_d) \quad (4)$$

The evaporation equivalent H_T can be exactly evaluated from net radiation R_n (W m^{-2}), measured at weather station of the Nižné Terianske pleso (or at the standard weather station "Skalnaté pleso"), and latent heat of vaporization L_v (J kg^{-1}) by the equation (5).

$$H_T = R_n/L_v \quad (5)$$

In research sites of Ľadové pleso and Długi Staw Gąsienicowy, the evaporation equivalent (H_T) was calculated by simplification (6):

$$H_T = (1 - \alpha)G/L_v \quad (6)$$

where, G – global radiation (W m^{-2}), L_v – latent heat of vaporization (J kg^{-1}), and α – albedo (–). Values reported by BURROUGHS (1991) were used for albedo: $\alpha = 0.1$ (water), $\alpha = 0.25$ (grassland), $\alpha = 0.3$ (rock or debris cones), and $\alpha = 0.95$ (snow).

In the absence of observed radiation parameters, PENMAN (SHAW, 1991) recommends empirical equations to calculate the evaporation equivalent of net radiation H_T (mm) in daily intervals:

$$H_T = R_i - R_o \quad (7)$$

$$R_i = R_a(1 - \alpha)(0.16 + 0.62n/N) \quad (8)$$

$$R_o = \sigma T_a(0.47 - 0.075 e_d^{0.5})(0.17 + 0.83n/N) \quad (9)$$

where, R_i – vaporizing equivalent of incoming and R_o – outcoming radiation of the vaporizing surface (mm), R_a – vaporizing potential to the latitude and season (mm), σ – Sefan-Boltzman constant (5.67×10^{-8} W m^{-2}K^{-4}), n – actual daily sunshine (hours day^{-1}), and N – maximum possible daily sunshine (hours day^{-1}).

The radiation income into terrestrial parts of a basin was calculated from the radiation income into the horizontal surface (data measured by sensors or calculated by equations 7, 8, 9) and the mean slope and face (HENDERSON-SELLERS & ROBINSON, 1989).

Generally, in mountain areas in summer, temperature inversions are rare and the ambient temperature decreases approximately linearly with altitude (BURROUGHS, 1991). Therefore, the linear decrease of the air temperature of 1.4°C per 100 m (reported by ŠPORKA et al., 2006) was accepted in the temperature-based approach (equation 1).

Considering the water surface of lakes, the vaporizing potential (EP, equations 1–4) corresponds to the actual evaporation (ET). In the case of grassland, the actual evapotranspiration (ET) was calculated by the method of GRINDLEY (SHAW, 1991) balancing the potential evapotranspiration (EP) and soil moisture deficit (SMD) given

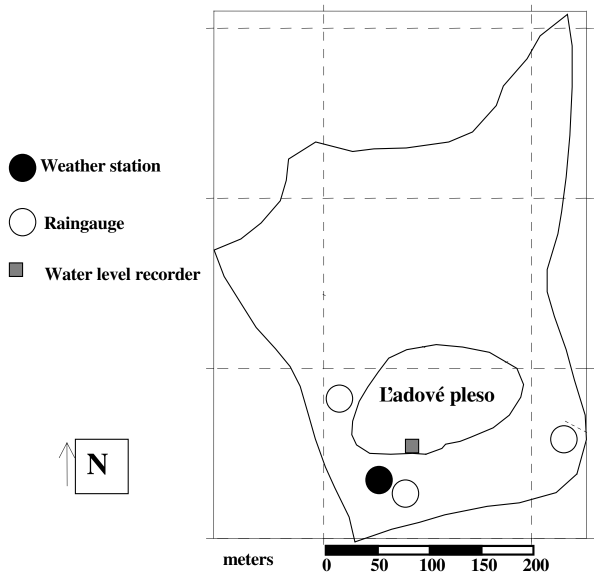

Fig. 1. Instrumented catchment of Ľadové pleso.

as the difference of vaporizing potential (EP), precipitation (P) and root constant (RC)

$$SMD = EP - P + RC \quad (10)$$

since the beginning of the vegetation period. The root constant of the presented grassland (shallow alpine soils) was assumed as $RC = 15$ mm.

The vaporization from the surface of rock formations or debris cones includes namely the process of surface interception ($E\text{-}INT$, mm per a unit of time), KŘEČEK & RUSNOK (1993). Considering daily time steps, the model of daily interception loss (HALL et al., 1992) can be accepted

$$E\text{-}INT = \beta(1 - \exp(-\delta P_d)) \quad (11)$$

where, P_d – daily rain (mm), β and δ – empirical parameters (recommended: $\beta = 1.2$ mm, and $\delta = 0.7$ mm^{-1}).

In the period of air temperatures below zero, KEMEL (1972) reported a simple temperature-based approach from $ET = 1$ (mm day^{-1}) by $T_a = (0\,°C)$ to $ET = 0.1$ (mm day^{-1}) by $T_a = -20$ (°C). Monthly values of ET (mm) versus air temperature T (°C) were approximated by the polynomial equation of second order (Fig. 4):

$$ET = 30 + 3.1T + 0.09T^2 \quad (12)$$

Runoff

The residence time in lakes corresponds to the inflow-outflow relationship. Changes in tracer concentrations of the lake water M_t (kg m^{-3}) with the time t (days) can be expressed by equation (12), COLE & PACE (1998, in TUREK, 2002):

$$M_t = M_o e^{-k} \quad (13)$$

where, M_o – initial tracer concentration at the application (kg m^{-3}), and k – dilution rate (day^{-1}). Thus, the residence time τ (days) can be expressed by equation (13):

Fig. 2. Automatic weather station near Ľadové pleso (photo: J. Křeček, 2001).

$$\tau = 1/k = (\ln M_\mathrm{o} - \ln M_\mathrm{t})/t \qquad (14)$$

The morphology of the Ľadové pleso basin (Fig. 1) indicates drainage of the catchment mainly by processes of lake seepage. Therefore, this catchment seems to be suitable for verification of a precipitation-runoff model. Water yield from the Ľadové pleso basin was calculated from the water balance model of Ľadové pleso by LJUNGGREN (2002). Here, outflow from the catchment is supposed to proceed by the seepage from the lake, which is related to the water level in the lake and season.

The deterministic model BROOK90 (FEDERER, 1992) was applied to simulate precipitation-runoff genesis at the Ľadové pleso basin in daily time steps. BROOK90 is a lumped parameter hydrological model operating in daily or hourly intervals. The model simulates land phases of precipitation, evapotranspiration and stream-flow of a small uniform watershed. Water is stored in the model either as intercepted rain or snow, snow on the ground, soil water, and groundwater. Water outputs are evaporation, deep seepage and stream flow. Stream flow is generated at the same time interval as precipitation input. Water is stored in the model either as intercepted rain or snow, snow on the ground, soil water in two layers and groundwater. The approach of PENMAN-MONTEITH (equation 3 modified by both atmospheric and canopy resistance) is used to calculate the evaporation potential. The actual evapotranspiration consists of five components: evaporation of intercepted

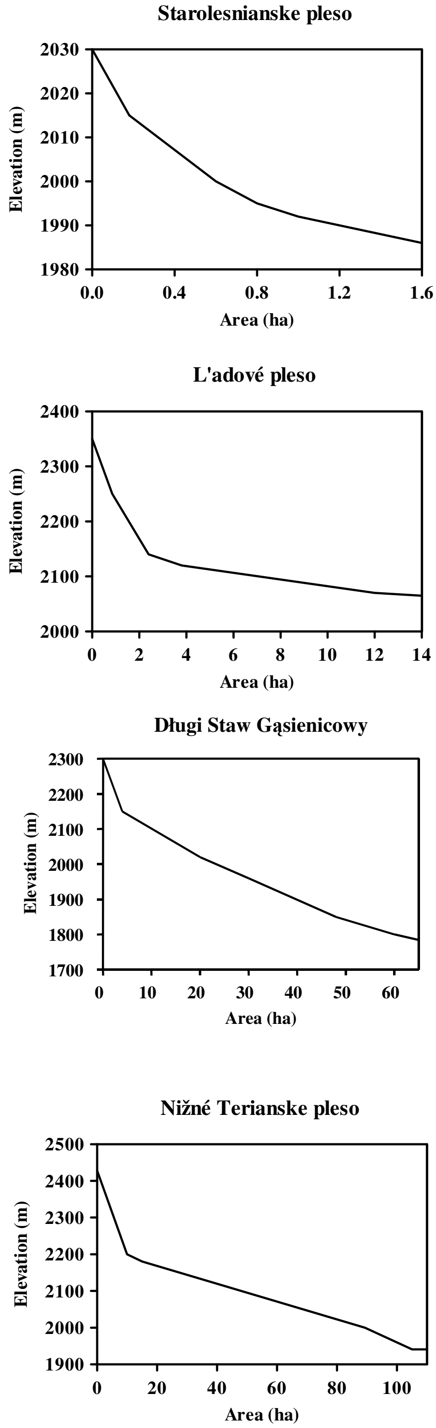

Fig. 3. Hypsometric curves of studied basins.

rain and snow, evaporation of snow, soil evaporation from the top soil layer and transpiration from soil layers containing roots. The model takes into consideration the degree of the slope, which is important for the evaporation and the water movement in the soil and surface. The canopy is characterised by maximum leaf area index LAI = 3.2 in the high summer, albedo $\alpha = 0.25$. Three soil layers were considered representing the alpine meadow: A – horizon (depth of 0.14 m, stone volume fraction of 0.2), B – horizon (depth of 0.33 m, stone volume fraction of 0.3), C – horizon (depth of 0.45 m, stone volume fraction 0.7). The matrix porosity

Table 5. Monthly sums of aerial precipitation over the investigated basins, 2000–2001.

Month	Starolesnianske pleso	Ľadové pleso	Długi Staw Gąsienicowy	Nižné Terianske pleso	Climate normal (1931–1960)
X	85	87	92	106	115
XI	98	123	102	131	110
XII	51	61	53	74	105
I	46	78	67	86	105
II	30	41	21	43	90
III	21	78	85	53	95
IV	125	135	152	45	125
V	127	139	120	77	165
VI	465	423	529	329	215
VII	416	407	459	380	225
VIII	139	173	230	189	225
IX	248	269	302	81	170
SUM	1851	2014	2212	1594	1745

Table 6. Monthly air temperature observed at instrumented weather stations, 2000–2001.

Month	Ľadové pleso (2066 m)	Długi Staw Gąsienicowy (1791 m)	Nižné Terianske pleso (1948 m)	Climate normal (1931–1960) (2000 m)
X	5.5	7.3	5.7	2.0
XI	0.2	2.1	0.4	–2.0
XII	–3.1	–3.8	–3.6	–5.0
I	–5.9	–5.1	–6.0	–8.0
II	–9.6	–4.7	–7.1	–3.0
III	–3.1	0.0	–3.2	–2.0
IV	–1.8	–0.1	–0.9	0.0
V	3.3	5.4	4.0	6.0
VI	3.6	5.6	4.8	8.0
VII	8.7	10.7	9.6	10.0
VIII	9.7	11.3	10.1	10.0
IX	2.8	3.9	3.1	6.0
Mean	1.6	2.9	1.5	1.7

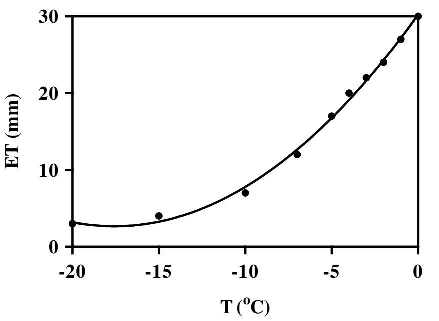

Fig. 4. Monthly evaporation ET (mm) from the snow-pack related to the mean monthly air temperature T (°C) in the range from 0 to −20 (°C): $ET = 30 + 3.1\ T + 0.09\ T^2$.

0.45 and water potential at field capacity −12 (kPa) (sandy loam soil) are assumed. The fine root mass is considered by 400 (g m^{-2}) to calculate the rhizosphere resistance. In the rock formations and debris cones, a crashed rock of surface area index of SAI = 1.6 and SAI = 2.5 uniformly distributed is assumed. The evaporation from the rock and debris cones is supposed to be formed only by the surface interception. In the winter, the evaporation from a smooth snow cover over the watershed area is supposed following the vaporizing potential. The degree-day snowmelt factor is assumed by value of 4.5 (mm day^{-1} °C^{-1}).

Results and discussion

Precipitation

Mean monthly and annual sums of aerial precipitation of studied catchments in the hydrologic year of 2001 are compared with the data of climate normal 1931–1960 (HMU, 1962) extrapolated to the focussed area (Tab. 5). In comparison to the climate normal (1961–1990), the estimates of annual precipitation differ from −9 to 27%. Particularly low winter precipitation of 2001 is evident; precipitation in the period XII–III is only 8–12% of the annual sum versus 22% of the normal.

From a relatively low number of observed foggy days in the weather station of Skalnaté pleso (39 events during four years, Lopatová, 2003) it is possible to consider in studied basins the additional sum of occult precipitation in the year of 2000/2001 by the range of 20–30 mm (1–2% of annual precipitation). However, in this region, HMU (1962) reported 100 foggy days per year of the climate normal 1931–1960. Worldwide, high values of cloud interception (10% of annual precipita-

Table 7. Monthly estimates of PENMAN's potential evaporation calculated in hourly ($PE\text{-}PEN_{(h)}$), daily ($PE\text{-}PEN_{(d)}$) and monthly ($PE\text{-}PEN_{(m)}$) intervals, grassland near the lake of Ľadové pleso, 2000–2001.

Month	$PE\text{-}PEN_{(h)}$ (mm)	$PE\text{-}PEN_{(d)}$ (mm)	$PE\text{-}PEN_{(m)}$ (mm)
X	46.6	44.9	37.5
XI	14.4	14.6	36.3
XII	18.2	18.8	0.0
I	7.4	7.9	0.0
II	9.7	9.4	0.0
III	21.1	21.1	0.0
IV	25.7	22.1	0.0
V	60.7	59.3	61.0
VI	44.5	43.5	59.0
VII	50.5	49.0	32.6
VIII	59.9	59.2	64.7
IX	26.6	25.7	62.6
SUM	385.3	375.5	353.7

tion) in alpine catchments are considered (GONZALES, 2000) particularly in forest stands. KŘEČEK & RUSNOK (1993) reported a significant role of occult precipitation in the process of interception at debris cones in the Zagros Mountains (Iran). Therefore, the considered values of additional precipitation forms in studied basins (20–30 mm per year) might underestimate the real situation.

Evapotranspiration

Monthly air temperatures registered at the instrumented weather stations well correspond with the climate normal of 1931–1960 (Tab. 6).

At the weather station of Ľadové pleso (grass-rock, elevation of 2066 m), monthly and annual (2000–2001) estimates of PENMAN's potential evaporation calculated in hourly ($PE\text{-}PEN_{(h)}$), daily ($PE\text{-}PEN_{(d)}$) and monthly ($PE\text{-}PEN_{(m)}$) intervals are given in Table 7. Annual values ($PE\text{-}PEN_{(h)} = 385$ mm, $PE\text{-}PEN_{(d)} = 376$ mm and $PE\text{-}PEN_{(m)} = 354$ mm) differ in 9 mm (2%, hourly versus daily steps) and 31 mm (8%, hourly versus monthly steps), respectively. Monthly values differ from 0 to 3.6 mm (0–9%, hourly versus daily steps), and from 0 to 25 mm (3–100%, hourly versus monthly steps), respectively. The *PE*-estimates of hourly and daily steps show relatively good agreement in both monthly and annual data, monthly *PE*-values underestimate the vaporizing potential particularly in the winter period (110 mm in the period of XII–IV). The annual estimate of evaporation from the snow-pack by the simple empirical approach (equation 12) in the winter period XII–IV is 78 mm (Tab. 8), 29% lower in comparison with $PE\text{-}PEN_{(h)}$.

Comparing the annual estimates of *PE-PEN* with potential evapotranspiration calculated at the nearest neighbouring standard weather station "Skalnaté pleso" (grassland, elevation of 1762 m, $PE\text{-}PEN_{(h)} = 494$ mm, and $PE\text{-}PEN_{(d)} = 474$ mm, LJUNGGREN, 2002), a negative gradient of −30 mm per 100 m of elevation is evident. Anyway, this gradient is not possible generalize because of complicated mountain morphology at the research area.

Annual estimates of vaporizing potential from water surface in studied catchments vary according to the formula applied: *E-HAM* = 232–261 mm (temperature-based equation of HAMON), *E-PRT* = 350–423 mm (radiation-based equation of PRIESTLEY-TAYLOR), and *E-PEN* = 401–537 mm (combination equation of PENMAN). The approach of PRIESTLEY-TAYLOR (equation 2) considers the aerodynamic part of vaporization by constant 30% of the total evaporation loss. The Penman's combination of both aerodynamic and energy balance method seems to provide the most realistic fit

Table 8. Evaporation from snow-pack, 2000–2001.

Month	Starolesnianske pleso	Ľadové pleso	Długi Staw Gąsienicowy	Nižné Terianske pleso
XII	18	18	18	21
I	13	13	14	15
II	8	7	10	12
III	20	18	22	22
IV	26	22	24	27
SUM XII–IV	85	78	88	97

Table 9. Estimates of potential annual evapotranspiration in studied watersheds, 1.10.2000 – 30.9.2001.

Watershed	*PE-HAM* (mm year^{-1})	*PE-PRT* (mm year^{-1})	*PE-PEN* water (mm year^{-1})	*PE-PEN* grass (mm year^{-1})	*PE-PEN* debris cones (mm year^{-1})
Starolesnianske pleso	246	423	436	354	321
Ľadové pleso	232	410	428	344	316
Długi Staw Gąsienicowy	260	350	401	326	302
Nižné Terianske pleso	261	391	537	458	331

Table 10. Actual annual evapotranspiration (mm year^{-1}) from studied watersheds, 1.10.2000 – 30.9.2001.

Period	Starolesnianske pleso	Ľadové pleso	Długi Staw Gąsienicowy	Nižné Terianske pleso
XII–IV	85	78	88	97
V–XI	326	308	309	326
Annual	411	386	397	423

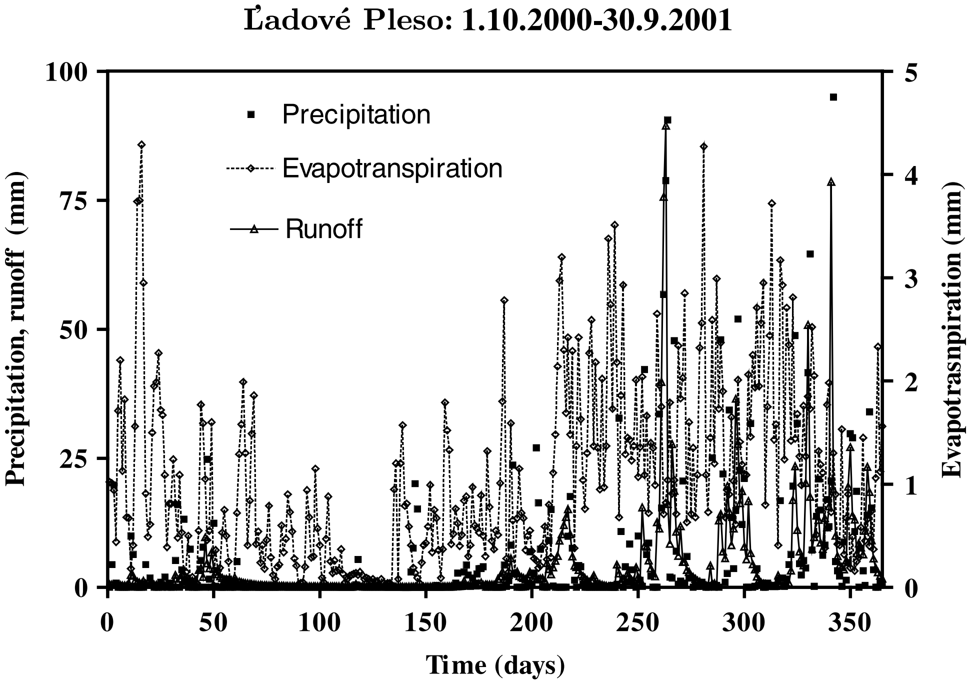

Fig. 5. Daily values of evapotranspiration and runoff predicted by BROOK90 (Ľadové pleso catchment, 2001).

to observed data in these for small headwater basins (Tab. 9). The actual evapotranspiration in catchments (Tab. 10) varies from 78 to 99% of the potential calculated for the vaporizing water surface.

The calculated values are higher than the regional estimates in the High Tatra Mts reported by Tomlain (1985): mean annual evapotranspiration 250–300 mm by gradient of –18 mm/100 m. Miklanek & Meszaros (1998) reported daily evapotranspiration in a small catchment (23 km^2) of Jalovecká dolina valley (elevation up to 2100 m) in April from 1–2 mm (north face) to 3–4 mm (south face). Those data are close to our results. However, the aim of our experiment was to contribute to better knowledge of water balance in high-elevated headwater basins above the timberline.

Runoff

From tracer experiments in 2000–2001, at the Ľadové pleso l, the maximum decrease of tracer concentrations were observed in lower layers of the lake (Turek, 2002). The residence time fluctuated from 32 to 313 days (by the weighted annual average of 136 days). The residence time in the Ľadové pleso (Tab. 11). In comparison to the other investigated lakes, in the basin of the Ľadové pleso, a relatively slow subsurface flow from the basin into the lake dominates, streaming inflows are not developed, and outflow occurred only by the lake seepage. Therefore, the basin of the Ľadové pleso seems to be suitable for application of precipitation-runoff models to study the runoff genesis.

Table 11. The residence time in Ľadové pleso.

Date	Time since application (days)	Dilution (day^{-1})	Residence time (days)
21/12/2000	55	0.0185	54
25/04/2001	125	0.0032	313
18/07/2001	84	0.0193	52
27/10/2001	101	0.0310	32
Annual average	365	0.0074	136

The deterministic model Brook90 applied here seems to be an effective tool to study precipitation-runoff genesis in small headwater catchments. At the basin Ľadové pleso, the predicted and observed water yield (Figs 5–6) show relatively good correlation:

Pearson $r = 0.72$ ($r_{\text{crit}} = 0.66$, $n = 365$, $P = 0.05$). The annual sum of actual evapotranspiration calculated

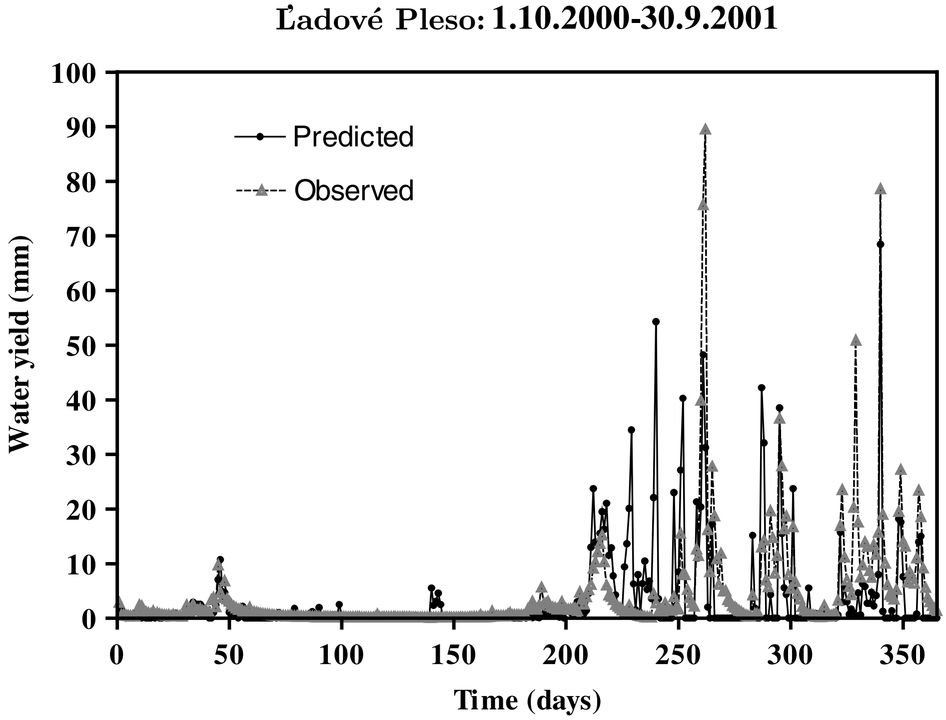

Fig. 6. Daily water yield of Ľadové pleso basin: predicted by BROOK90 versus observed (calculated from the water balance of the lake).

by BROOK90 (352 mm) also good corresponds to the value estimated by the approach of Penman (386 mm).

The data of watersheds registered in the standard hydrological network (Tab. 3) include lower elevations with different ecosystems (namely a high percentage of forest cover from 30 to 50% of the basin area). So, it is difficult to compare these data with water balance estimates of basins above the timberline. The low value of actual evapotranspiration ($ET = 150$ mm) reported by Vološčuk et al. (1994) for the catchment of Studený potok (Nad Starou Lesnou, 18.5 km^2, 30% forested) seems not to be realistic.

The empirical hydrological results show that hydrological processes are governed by the temperature-dependent regime of high mountain snowmelt. However, they also highlight major differences between both water yield and runoff genesis in the watersheds, which are caused by the differences in the morphologies of the lakes and their basins, in the soil-vegetation complex, and in runoff routing.

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