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Hydrologická funkce horských vrchovišť a vlastnosti rašelinných vod v pramenné oblasti Vydry

Hydrological function of peat bogs and peat water properties of the Vydra River headwaters

Disertační práce

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Abstrakt

Retenční potenciál krajiny a vodní režim pramenných oblastí jsou v současné době vyznačující se vysokou četností hydrologických extrémů, zejména sucha, důležitým směrem výzkumu. Práce je zaměřena na výzkum horských vrchovišť, která vzhledem ke svému plošnému rozšíření v pramenné oblasti řeky Vydry tvoří významnou složku v rámci srážko-odtokového procesu zájmového území. Velmi specifické hydropedologické vlastnosti organozemí zajišťují velkou retenční kapacitu daného území, avšak jejich uplatnění a míra zapojení do odtokového procesu je závislá na komplexu fyzickogeografických faktorů. V posledních desetiletích se výrazně měnily názory na hydrologickou funkci rašelinišť. Vzhledem k tomu, že v rámci vrcholových partií Šumavy je formování odtoku vázáno na organogenní a hydromorfní půdy, důležitým faktorem je proto aktuální nasycenost povodí. Organozemě jsou bezpochyby významnou zásobárnou vody a mají zásadní vliv pro okolní krajinu, nicméně zejména během extrémních událostí mohou zvyšovat extremitu odtoku. Zásadní část práce je proto založena na detailním pozorování dynamiky hladiny podzemní vody, která je nejvýznamnějším faktorem vývoje těchto unikátních oblastí, ale i nejdůležitějším prvkem k pochopení hydrologického režimu rašelinných stanovišť. Dalším významným prvkem při hodnocení odtokových poměrů horských vrchovišť je sledování fyzikálně-chemických vlastností rašelinných vod. Jejich specifické složení lze stopovat ve vodních tocích a tím získávat informace o zapojení horského vrchoviště do odtokového procesu. Vliv organozemí v pramenné oblasti Vydry na hydrologické procesy a retenci vody v krajině je nezpochybnitelný, nicméně množství infiltrované vody, způsob jejího proudění a dotování vodních toků jsou v současnosti velmi aktuálními otázkami, jejichž studium přispívá k poznání hydrologických vazeb v krajině.

Klíčová slova: horské vrchoviště, hydrologická funkce, retenční potenciál, hladina podzemní vody, hydrologický režim, organozem, Šumava, Vydra

Abstract

The retention potential of landscapes, along with the water regime of spring areas, are important hydrological topics of research, particularly in the current context of increasing extreme drought frequencies. The present work is focused on monitoring the mountain peat bogs, which, due to their overall frequency of occurrence in the spring area of the Vydra river, represent a significant constituent of the rainfall-runoff process of the area of interest. The specific hydropedological features of the organogenous soils (Histosol type soil) provide the high retention potential of the area, however, the influence of these soils on the runoff process is determined by complex physicogeographical factors. The general opinion on the hydrological function of the peat bogs has changed in recent years and the most important factor in the runoff formation in the mountain area of the Šumava Mts. is now thought to be the actual saturation of the headwater, which is predominantly composed of hydromorphic and organogenous soils.

The organogenous soils are significant water reservoirs and have an important impact on the landscape. However, they may also intensify the extreme values of the watercourses during extreme precipitation events. The fundamental part of this work focuses on detailed observations of the groundwater level dynamics, which is the key factor for the future development of these precious sites and for the comprehension of the hydrological regime of the peatlands. Evaluation of the runoff processes in the mountain peat bogs also requires a detailed observation of the physico-chemical peat water properties. The specific properties of peat waters can be identified in the watercourses and, thus, the involvement of the mountain peat bog in the runoff process can be proven. The impact of organogenous soils on the ongoing hydrological processes in the spring area of the Vydra river is undeniable. Nevertheless, the amount of infiltrated water, the means of water flow, and the supply of the watercourses raise important questions leading to recognition of the hydrological links in the landscape.

Key words: peat bog, hydrological function, retention potential, groundwater level, hydrological regime, Histosol, Šumava Mts., Vydra river

Obsah

1. Úvod a cíle práce

Vzhledem ke zvyšující se frekvenci meteorologických a hydrologických extrémů, zejména častého výskytu sucha, se retence vody v krajině stává velmi aktuálním tématem. Horská vrchoviště jsou díky svým retenčním schopnostem významným hydrologickým a krajinným prvkem v rámci centrální Šumavy.

V kontextu katastrofických povodní a extrémního sucha v posledních letech bylo nezbytné řešit ochranu před povodněmi a zároveň zavést opatření ke zvýšení průtoků v suchých obdobích. Z tohoto důvodu je nezbytné zaměření se na pramenné oblasti s vysokou retenční schopností (Janský, Kocum 2008). Jednou z nejdůležitějších složek retence vody v krajině je samotná retenční schopnost půdy. Fyzikální vlastnosti půdního prostředí rozhodují, za jakých okolností dojde k povrchovému odtoku, kolik vody se může infiltrovat do půdy během srážkové události, nebo jak dlouho se dokáže voda v půdě zadržet během suchých period (Vlček 2017).

Výrazné plošné zastoupení organogenních půd v pramenné oblasti Vydry má zásadní vliv na hydrologický režim této zájmové oblasti. Otázka hydrologické funkce horských vrchovišť není v současné době zcela jednoznačně zodpovězena, a v posledních desetiletích se názory na ni navíc výrazně lišily. Výzkum šumavských rašelinišť započal v druhé polovině 20. století (Špirhanzl 1951; Ferda 1969; Ferda et al., 1971). Tyto práce si kladly za cíl popsat hydrologický režim rašelinných komplexů, který byl označován jako jednoznačně pozitivní s ohledem na zadržování vody v krajině. Významné období výzkumu rašelinných oblastí nastalo v 70. a 80. letech. Ve světové literatuře se však objevují práce, které naznačují, že horská vrchoviště se mohou projevovat hydrologicky negativně během extrémních srážkových událostí, a mohou tak zvyšovat extremitu vodních toků (Baden, Eggelsmann 1970; Burke 1975; Moklyak et al., 1975). Zároveň lze toto období označit za hlavní fázi melioračních prací. Touto činností bylo postiženo přibližně 70 % rozlohy všech rašelinných stanovišť na Šumavě, což mělo za následek postupnou degradaci těchto cenných ekosystémů (Bufková et al., 2010).

Téma retenčního potenciálu pramenných oblastí a s tím spojený výzkum horských vrchovišť se dostalo do popředí zájmu počátkem druhého tisíciletí. Práce se přiklánějí k názoru, že rašelinné komplexy jsou oblastmi s vysokou retenční kapacitou a unikátními hydropedologickými, vegetačními či topografickými vlastnostmi. Jednotlivá horská vrchoviště se však mohou vzájemně lišit, proto je nezbytné přistupovat k výzkumu těchto rašelinných komplexů individuálně. K poznání hydrologické funkce horských vrchovišť přispěla i řada výzkumů zaměřených na pramennou oblast řeky Vydy, které byly realizovány v rámci hydrologické sekce Katedry fyzické geografie a geoekologie, Přírodovědecké fakulty Univerzity Karlovy v Praze (Janský, Kocum 2008; Čurda et al., 2011; Kocum 2012; Kocum et al., 2018; Vlček et al., 2012; Vlček et al., 2016; Vlček 2017, Doležal et al., 2020). Téma je tak stále v popředí zájmu vědeckého výzkumu, a to zejména s ohledem na značnou retenční kapacitu a velmi specifické vegetační poměry horských vrchovišť. Významným tématem jsou také revitalizační opatření rašelinišť, které napomáhají zadržení vody v krajině, což je vzhledem ke zvyšujícímu se výskytu sucha jednou z možností, jak zmírňovat jeho negativní dopady na krajinu.

Tato disertační práce je syntézou jednotlivých výzkumů a klade si následující cíle:

- Zhodnocení dynamiky hladiny podzemní vody horského vrchoviště
- Zhodnocení retenčního potenciálu horských vrchovišť
- Popis formování odtoku rašelinných stanovišť
- Popis dynamiky sledovaných fyzikálně-chemických vlastností rašelinných vod

2. Současný stav poznání

Význam rašelinných stanovišť je značný. Nacházejí se po celém světě od tropů až po vysoké zeměpisné šířky, a přestože zabírají přibližně 3 % zemského povrchu, je v nich uloženo až 10 % světových zásob sladké vody a až jedna třetina uhlíku uloženého v půdě. Jejich hodnota spočívá i v jedinečné biologické rozmanitosti (Rubec 2005). V Česku se dle průzkumů nachází přibližně 10 400 ha horských vrchovišť. Největší zastoupení mají vrchoviště na Šumavě, v Krušných a Jizerských horách, nicméně drobnější plochy nalezneme i v dalších okrajových pohořích Českého masivu (Pošta et al., 2003).

2.1 Odtokový proces v rašelinných komplexech

Pramenné oblasti vodních toků představují zdrojová území formování odtoku. Pramenné oblasti jsou z hlediska fyzickogeografických poměrů velmi heterogenní území. Území české strany Šumavy není výjimkou. Významným, ale nikoliv jediným specifikem je existence ploch vrchovištních komplexů, které tak představují ekologický fenomén zdejší krajiny. V území pramenné oblasti Vydry je proto klíčovým faktorem studium vlivu tohoto jevu na hydrologický režim zdejší krajiny a tvorbu odtoku (Kocum 2012). Až po soutok řek Vydry a Křemelné zaujímají rašeliniště a hydromorfní půdy přibližně 15 % rozlohy v součtu celkové plochy obou povodí, přičemž pouze v povodí Vydry se jedná téměř o jednu třetinu (Janský 2004). Je tedy zřejmé, že se setkáváme s regionálním specifikem, které má na formování odtoku výrazný vliv.

Obecně lze odtokový režim pramenné oblasti Vydry označit jako jednoduchý s výrazným maximem v období jarního tání sněhové pokrývky. Průtoková minima se pak vyskytují na konci února a druhotně v říjnu. Variabilita průměrných denních průtoků je nejvyšší v období tání sněhu, významnou rozkolísanost odtoku lze pozorovat i během letního období, a to vzhledem k větší četnosti výskytu intenzivních dešťových srážek (Kocum 2012). Nicméně tento popis odtokových charakteristik odpovídá celému povodí, které obsahuje celou řadu půdních typů včetně minerálních půd, a stejně tak mohou být heterogenní i vegetační poměry. To poté vede k variabilitě formování odtoku a diferenciaci podpovrchového odtoku. V případě malých dílčích subpovodí, kde výrazně dominují rašelinné a hydromorfní půdy, nemusí být tyto obecné odtokové charakteristiky zcela platné (Hümann et al., 2011).

Výzkum hydrologického režimu rašelinných oblastí na Šumavě započal již na konci 60. let 20. století (Ferda 1969; Ferda et al., 1971). V té době byla rašeliniště chápána jako ideální regulátor průtoků, který v době vysokých srážkových úhrnů vodu akumuluje a v době sucha jí dotuje vodní toky. Tato teorie byla s přibývajícími poznatky později upravena a zahraniční literatura se přikláněla k názoru, že rašelinné oblasti spíše umocňují odtok v období srážek, jelikož jsou po celý rok téměř zcela nasyceny a v době sucha toky nedotují (Burke 1975; Moklyak et al., 1975). Novodobý výzkum rašelinných oblastí na Šumavě byl iniciován hlavně kvůli častému výskytu povodní v 90. letech 20. století, kdy bylo nezbytné detailní studium oblastí s vysokým retenčním potenciálem (Janský 2004; Čurda et al., 2011). Dalším impulzem byla prováděná revitalizační opatření rašelinišť (Bufková 2006; Bufková et al., 2010., Bufková 2013). V současné době je pak výzkum směřován zejména na zkoumání retenční kapacity pramenných oblastí ve vztahu k prohlubujícímu se suchu, případně vegetačním změnám (Janský, Kocum 2008; Kocum 2012; Bufková, Stíbal 2012; Kučerová et al., 2009).

Odtok vody z rašelinných oblastí je nyní obecně charakterizován rychlými nárůsty i poklesy odtoku a rozkolísaností průtoků. Zároveň rašeliniště vykazují velmi nízký bazální odtok v období sucha a naopak intenzivní nárůst průtoku během srážkových událostí (Evans et al. 1999; Bragg 2002). Nicméně je třeba brát v úvahu specifické fyzickogeografické podmínky daného stanoviště včetně aktuálního nasycení povodí a dalších významných faktorů, které ovlivňují srážko-odtokový proces. Z těchto důvodů je při studiu jednotlivých komplexů vhodné volit individuální přístup.

2.2 Dynamika výšky hladiny podzemní vody horských vrchovišť

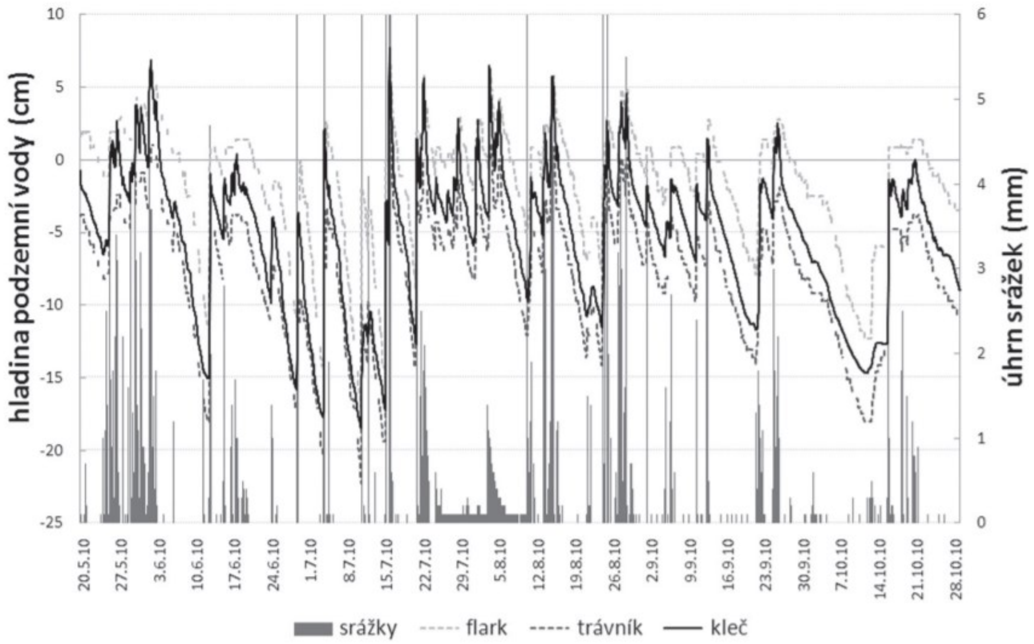
Klíčovým faktorem pro vývoj horského vrchoviště je výška a kolísání hladiny podzemní vody. Pohyb vody v akrotelmu přímo ovlivňuje průtoky, chemismus a druhovou skladbu stanoviště (Labadz et al., 2010; Allott et al., 2009). Kolísání hladiny podzemní vody je velmi důležitým faktorem při hodnocení odtoku vody. Hladina podzemní vody se nachází po většinu roku velmi blízko k povrchu a fluktuace vody je výrazně omezena (Holden et al., 2011). Hladina podzemní vody má v řadě typů rašelinišť výraznou sezónní i meziroční dynamiku. Relativně stabilní hladina bývá jen na rašeliništích sycených pramennou nebo artézskou vodou. Naopak vrchoviště a zalesněná rašeliniště závislá pouze na dotaci srážkovou vodou vykazují během letních přísušků výrazné poklesy (Kučerová et al., 2009). Během dlouhodobého sucha a nízké hladiny podzemní vody může docházet k nevratným fyzikálním změnám rašeliny, a tím i k negativnímu ovlivnění hydrologických procesů probíhajících v krajině (Evans et al., 1999). Dynamika hloubky hladiny podzemní vody během vegetační sezony je značná a na její změně se projeví i menší srážka. Rychlost pohybu hladiny může dosahovat 2 až 3 cm za den (Vlček et al., 2012).

Velmi silná vazba existuje mezi výškou hladiny podzemní vody a příslušným vegetačním typem. Rašeliništní vegetace reaguje velice citlivě na sebemenší změny vodního režimu. Dlouhodobý pokles či vzestup hladiny může vyvolat sukcesní změny (Kučerová et al., 2009). Existuje vysoký stupeň zpětné vazby mezi růstem vegetace a výškou hladiny podzemní vody. S vyšším pokrytím vegetací se zlepšuje samoregulační schopnost rašeliny. Dochází totiž k udržení vyšší vlhkosti vzduchu pod vegetací, a tím se snižuje i evaporace (Binet et al., 2013). Vegetace horských vrchovišť je přirozeně adaptována na vysokou hladinu podzemní vody. Hladina tak určuje ekologické niky rostlinných druhů (Kværner, Snilsberg 2011). Vegetační pokryv je rovněž určující pro energetickou bilanci horského vrchoviště, čímž spoluutváří vodní režim krajiny (Brom et al., 2009). Jednotlivé typy rašelinišť, či jejich části (lagg, centrální část vrchoviště) tak mají charakteristické průměrné hodnoty podzemní vody, na něž je vázána specifická vegetace. Tyto hodnoty byly stanoveny na základě dlouhodobého monitoringu i pro rašeliništní biotopy Šumavy, viz tab. 1 (Bufková 2012).

Tab. 1: Přirozená hladina podzemní vody rašelinných biotopů (Bufková 2012)

Typ biotopu	Hladina podzemní vody (cm pod povrchem)
Vrchoviště (*Sphagnum medii, Oxycocco-Ericion, Leuc-Scheuchzerion palustris*) a lagg vrchoviště	5 (10)
Rašelinné smrčiny (*Sphagno-Piceetum*)	15-20
Podmáčené smrčiny (*Bazzanio trilobatae - Piceetum, Soldanello - Piceetum*)	20-35
Přechodová rašeliniště (*Sphagno recurvi - Caricion canescentis*)	5-10
Luční rašeliniště (*Caricion fuscae, Caricion demis*sae)	10-20

Pro zjednodušení je v některých studiích taktéž sledována hladina podzemní vody v rámci jednotlivých vegetačních typů (např. kleč, trávníky, podmáčená smrčina), nebo jsou porovnávány jednotlivé mikrotopografické prvky rašeliniště (např. flarky, šlenky, bulty). Tyto gradienty jsou v horských vrchovištích snadno rozpoznatelné. Zpravidla v centrální části vrchoviště je hladina podzemní vody vyšší a stabilnější než při okrajích, vegetace je tak tvořena především rašeliníky. Naopak na okrajích se vyskytuje více živin, voda se nachází hlouběji v půdním profilu a je mineralizovaná, což poté umožňuje růst klečového a stromového patra (Faubert 2004). Podobné studie byly realizovány i v rámci Šumavy, viz obr. 1 (Vlček et al., 2012; Kučerová et al., 2009).



Obr. 1: Průběh hladiny podzemní vody v různých částech rašeliniště - Rokytecké slatě (Vlček et al., 2012; Kučerová et al., 2009)

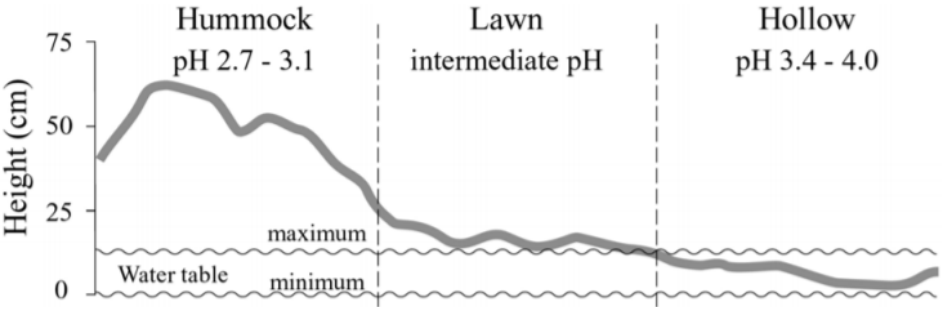
Velmi často je v případě Šumavy také diskutována problematika výšky a variability hladiny podzemní vody s ohledem na prováděná revitalizační opatření (Bufková 2006; Bufková et al., 2010., Bufková, Stíbal 2012; Bufková 2013), nebo případně ve spojení se studiem variability odtoku z rašelinných povodí (Janský, Kocum 2008; Čurda et al., 2011, Kocum 2012; Vlček et al., 2012; Kocum et al., 2018).

2.3 Fyzikální a chemické vlastnosti rašelinných vod

Důležitým směrem studia rašelinných stanovišť je monitoring fyzikálních a chemických vlastností povrchových a podpovrchových vod. Specifické vlastnosti rašelinných vod mohou přispět k objasnění míry zapojení horských vrchovišť do vodního režimu. Chemické parametry rašelinných vod jsou také sledovány při hodnocení stavu (případně míry degradace) horských vrchovišť nebo ve spojitosti s prováděnými revitalizačními opatřeními.

Ombrotrofní vrchoviště sycená srážkovou vodou se vyznačují poměrně stálými hydrochemickými vlastnostmi. To ovšem nemusí platit pro celé vrchoviště, například lagg může být ovlivněn minerálními půdami z okolí rašeliniště. Míra ovlivnění horského vrchoviště více mineralizovanou vodou z okolí poté závisí především na jejím pH, elektrické konduktivitě a na množství rozpuštěného vápníku a uhličitanů (Howie, Meerveld 2011).

Vody horských vrchovišť jsou kyselé, jejich pH se pohybuje zpravidla mezi hodnotami 3,3 – 5,5. Zmíněné hodnoty však závisí na řadě faktorů, např. na klimatu, výšce hladiny podzemní vody nebo biologické aktivitě (Bergsma, Quinlan 2009; Holden et al., 2004). Dalším faktorem ovlivňující pH může být míra antropogenního ovlivnění. Horská vrchoviště postižená odvodněním zpravidla vykazují v průměru nižší pH, které však může vykazovat vysokou variabilitu (Wind-Mulder et al., 1996). Jako výrazný faktor změn pH je třeba zmínit také vyplavování organického uhlíku. Koncentrace organického uhlíku mají silný sezónní průběh související s evaporací vody a produkcí organické hmoty. Přirozeně vyšší obsah organických kyselin společně s nízkou celkovou mineralizací má za následek nízké pH (Kocum 2012). Obecně se v případě rašelinných oblastí jedná o vztah, kdy během období nízkých hladin podzemí vody se zmenšuje množství vyplavovaného organického uhlíku a roste pH (Seibert et al., 2009). Vzhledem k mikrotopografii terénu je tedy třeba uvažovat i rozdílné fyzikálně-chemické vlastnosti jednotlivých částí rašeliniště (bulty, trávníky, šlenky). Liší se zejména svým zamokřením, pH, distribucí živin či rychlostí dekompozice (obr. 2). To vše je spojené i s charakteristickým výskytem vegetace na všech úrovních - mechorosty, traviny, cévnaté rostliny (Campbell, Rochefort 2001; Faubert 2004).



Obr. 2: Mikrotopografický gradient s charakteristickými hodnotami pH (Faubert 2004)

Se změnami pH v rašeliništi souvisí i změny elektrické konduktivity. Elektrická konduktivita se obecně mění s množstvím obsažených iontů. V případě kationtů jde o prvky rozpuštěných solí, nejčastěji Mg, K, Na, ale také Fe, Mn nebo Al. V případě velmi kyselého pH pak vyšší hodnoty elektrické vodivosti mohou způsobit volné anionty vodíku (Worrall et al., 2006). Měření konduktivity v horských vrchovištích přináší velmi variabilní hodnoty, a to od několika jednotek až po stovky $\mu S.m^{-1}$. Konduktivita závisí nejen na saturaci rašeliny, ale také na povrchu pevných zrn, jež jsou s vodou v kontaktu, a lineárně koreluje s množstvím rozpuštěných látek ve vodě. Konduktivita je také ovlivněna teplotou vody, a vlastnostmi rašeliny, jako jsou např. kationtová výměnná kapacita, organický obsah, pH či pórovitost (Ponziani et al., 2011).

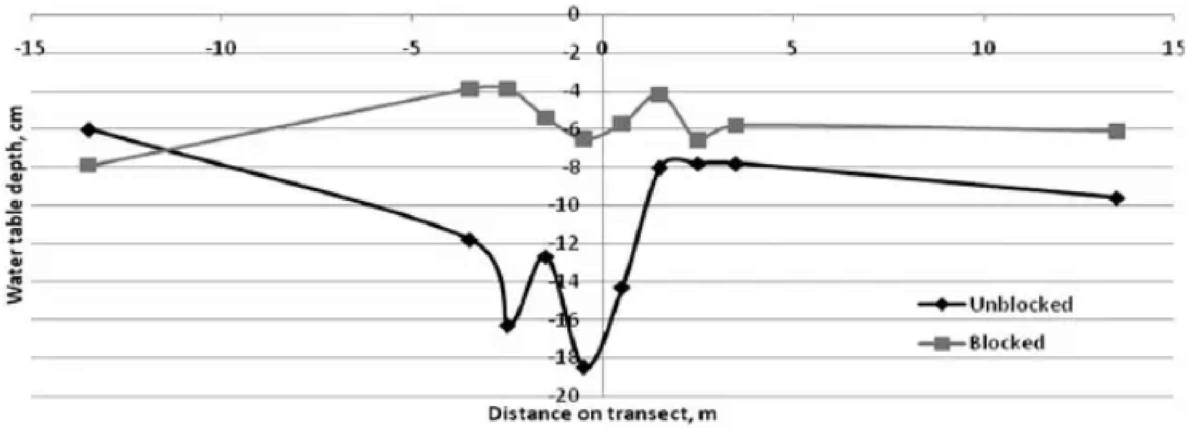
Dalším významným parametrem vody, který je v kontextu rašelinných stanovišť sledován, je rozpuštěný kyslík. Je nezbytný pro aerobní dýchání na všech trofických úrovních. Kyslík je do vody dodáván difuzí z atmosféry, promícháváním vody a vodními organismy, které kyslík produkují. Významným faktorem v případě rozpustnosti kyslíku je teplota. A to jak přímým vlivem, jelikož v chladnější vodě se kyslík rozpouští snadněji, tak i nepřímým vlivem, kdy má teplota signifikantní vliv na metabolické procesy (Yvon-Durocher et al., 2010). V případě organogenních půd jsou vstupy kyslíku omezeny, což podporuje anaerobní procesy, a to v důsledku vede k pomalému rozkládání organické hmoty. Změny hladiny podzemní vody jsou tedy spojeny s posuny aerobní a anaerobní zóny (Estop-Aragonés et al., 2012).

2.4 Revitalizační opatření horských vrchovišť

Výrazné narušení rašelinišť lidskou činností vedlo k odvodnění a následné postupné degradaci těchto jedinečných stanovišť. Na Šumavě probíhaly meliorační práce ve dvou hlavních fázích a to na přelomu 19. a 20. století a poté v 70. a 80. letech 20. století, v rámci procesu tzv. intenzifikace. Inventarizační průzkumy provedené v uplynulých letech ukázaly, že odvodněním je na Šumavě v různé míře poznamenáno téměř 70 % rašelinných stanovišť. Okolí rašelinišť i samotná rašeliniště byla tradičně odvodňována za účelem těžby rašeliny, kultivace zemědělské půdy nebo pro zvýšení produkce dřeva v podmáčených lesních porostech (Bufková 2013). Po detailní inventarizaci rašelinišť na území Šumavy jsou od roku 1999 postupně realizovány jednotlivé revitalizační projekty založené na přehrazování odvodňovacích rýh dle konceptu cílové hladiny podzemní vody (Bufková, Stíbal 2012).

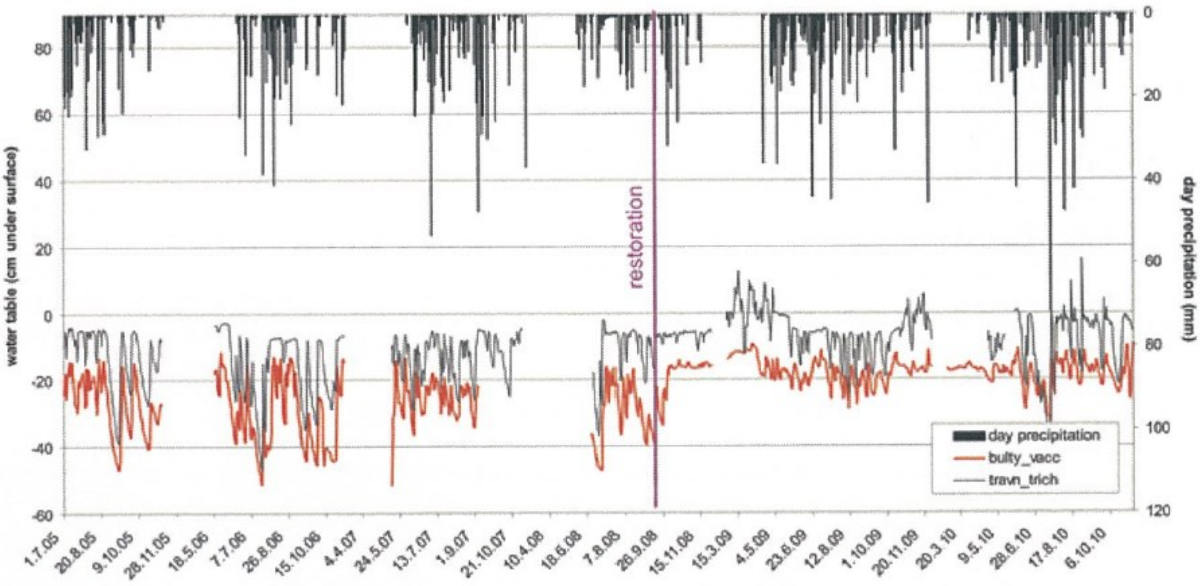
Vlivem revitalizačních opatření rašelinišť na vodní či vegetační poměry v rámci Šumavy se zabývá řada prací (Bufková 2006; Bufková 2013; Bufková et al., 2010; Čurda et al., 2011; Kocum 2012). Odvodňování rašelinišť a následná revitalizační opatření však nejsou specifika, která by se týkala pouze Šumavy. Výrazně byla tato stanoviště ovlivněna lidskými zásahy i v dalších částech Evropy a světa, tudíž je tato tematika často diskutována i v rámci zahraniční literatury. Popis narušení rašelinišť, následná revitalizační opatření a jejich vliv na vegetační, hydrologické, či hydrochemické poměry nalezneme v pracích zejména z Velké Británie (Allott et al., 2009; Holden et al., 2004; Holden et al., 2011; Wilson et al., 2011; Worrall et al., 2007), Skandinávie (Haapalehto et al., 2011; Kværner, Snilsberg 2011) nebo z Kanady (Ketcheson, Price 2011; Price et al., 2003).

Výzkumy poukazují především na velmi pozitivní efekt zvýšení hladiny podzemní vody po provedení revitalizačních opatření. Jedná se o zvýšení hladiny podzemní vody o několik centimetrů až decimetrů v závislosti na typu rašeliniště, vegetaci, hloubce odvodňovacích kanálů a topografických poměrech konkrétního stanoviště (Bufková 2013; Holden et al., 2004; Price et al., 2003; Worrall et al., 2007). Revitalizační opatření mají plošný dosah i několika metrů. Lze uvést příklad z Velké Británie, kde došlo na sledovaných transektech ke statisticky významnému zvýšení hladiny podzemní vody až v 5 metrech od odvodňovacího kanálu – viz obr. 3 (Armstrong et al., 2010).



Obr. 3: Průměrná hladina podzemní vody sledovaná v transektech na přehrazené a nepřehrazené části odvodňovacího kanálu (Armstrong et al., 2010)

Obdobné výzkumy byly realizovány i v rámci Šumavy (obr. 4). Příznivé výsledky revitalizace byly například zaznamenány na lokalitě Schachtenfilz. Zvýšení hladiny podzemní vody a snížení její amplitudy jsou klíčovými cíli revitalizace, kterých bylo v tomto případě dosaženo zejména v keříčkovitých porostech (Bufková, Stíbal 2012).



Obr. 4: Kolísání hladiny podzemní vody před a po revitalizaci na středně narušeném otevřeném vrchovišti (lokalita Schachtenfilz), bulty vacc – keříčková vegetace s dominantními druhy rodu *Vaccinium*, travn trich – trávníky se suchopýrem *Trichophorum cespitosum* (Bufková, Stíbal 2012)

3. Metodika

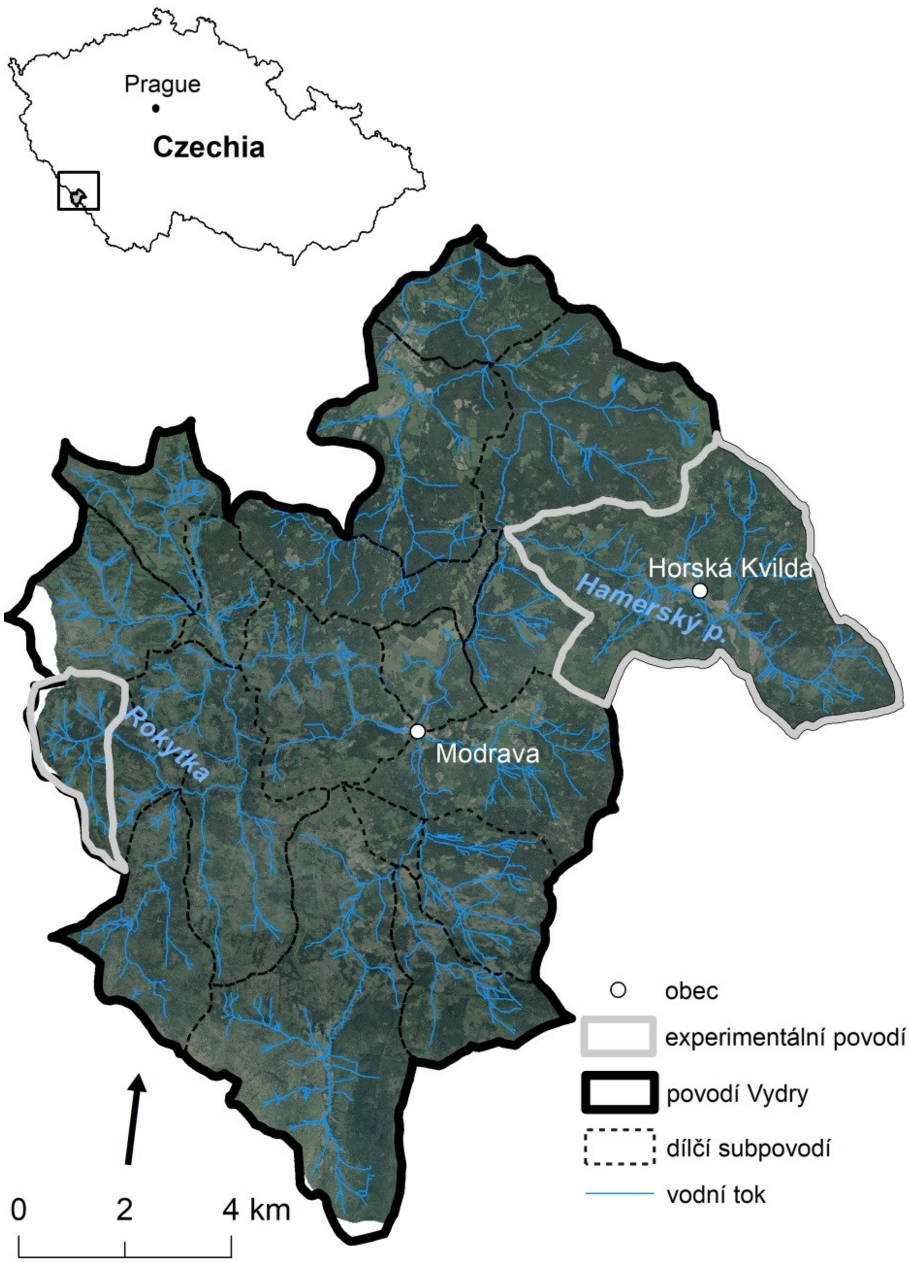
Jednotlivé články byly realizovány na základě vlastních terénních měření a za využití dat z automatických měřicích stanic Univerzity Karlovy v Praze, Přírodovědecké fakulty, Katedry fyzické geografie a geoekologie. Jednotlivé studie se zaměřují na dílčí subpovodí v rámci povodí řeky Vydry.

3.1 Zájmové území

Zájmová oblast byla vybrána na základě charakteristických fyzickogeografických podmínek, které byly předmětem studia. Konkrétně se jedná o malé rašelinné části v rámci Rokytecké a Mezilesní slati, které spadají do experimentálních povodí Rokytky a Hamerského potoka (obr. 5). Důležitým aspektem výběru těchto oblastí byla přítomnost horského vrchoviště a dalších rašelinných biotopů, které lze od sebe snadno odlišit. Tím je možné studovat daná stanoviště v detailním měřítku a přesněji popsat sledované jevy.

Obě povodí náleží do oblasti centrální Šumavy, která spadá do povodí řeky Vydry. Průměrná nadmořská výška povodí horní Vydry činí 1078 m a má vějířovitý tvar. Oblast má charakter náhorní plošiny se zarovnaným povrchem a poměrně nízkou sklonitostí svahů (Kocum 2012). Z hlediska půdního krytu se jedná o půdní region kambizemí oligobazických až rankerů výrazněji svažitých poloh a region kryptopodzolů až podzolů (Šerfna 2004). Hlavním specifikem povodí je však značný výskyt hydromorfních a organogenních půd (Šefrna 2004; Vlček et al., 2012; Vlček 2017). Dalším charakteristickým rysem povodí jsou klimatické poměry. Jedná se o území s relativně vysokými ročními úhrny srážek, zejména na návětrných stranách (stanice Březník, Modrava), a to až v rozmezí 1300 – 1600 mm. Zároveň se jedná o jeden z nejchladnějších regionů Česka. V případě vegetačního krytu dominují lesní porosty. Z přírodního hlediska má celé území velmi zachovalý ráz, který díky specifické flóře, především pak výskytu rašelinišť vrchovištního typu, patří k jednomu z nejcennějších míst nejen v rámci Šumavy, ale celého Česka (Kocum 2012).

Detailní fyzickogeografický popis zájmových území je zpracován v rámci jednotlivých článků disertační práce.



Obr. 5: Povodí řeky Vydry s vyznačením experimentálních povodí

3.2 Formování odtoku horských vrchovišť

Pro účely hodnocení formování odtoku v závislosti na půdním typu (organozem, podzol) v povodí řeky Rokytky byla využita data z pěti automatických srážkoměrných stanic (s intervalem měření 10 minut) v okolí povodí Rokytecké slati (v rozmezí vzdáleností 0,1 - 7,8 km od zájmového povodí). Rovněž byla využita data ze dvou totalizátorů, které provozuje v téže lokalitě Český hydrometeorologický ústav (Starostová 2012). Jednotlivé přístroje ale vykazovaly odchylky. Rozdíl hodnot srážek (automatický srážkoměr versus totalizátor) je pravděpodobně způsoben rozdílnou nadmořskou výškou povodí a srážkoměru. K rozdílům však také může dojít kvůli často sledovanému podhodnocení srážek v případě standardních automatických srážkoměrů, které je zapříčiněno existencí drobných větrných vírů nad stanicemi. V případě deště může dosahovat rozdíl asi 10 %, v případě sněhu však až 50 % (Dingman 2015).

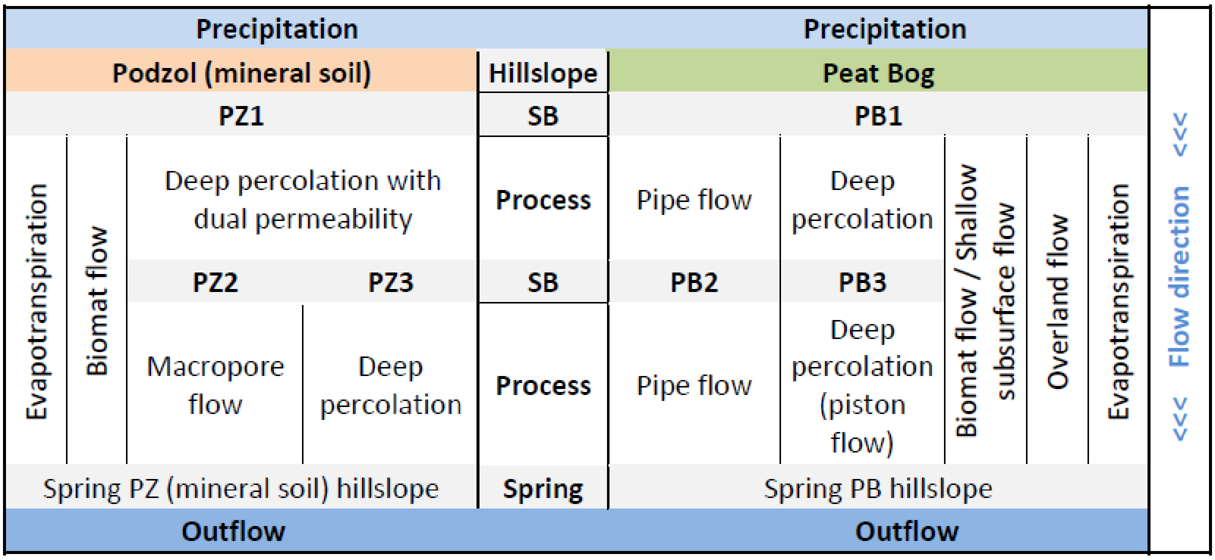
Dalším nezbytným parametrem byl výpočet potenciální evapotranspirace dle rovnice (1) Penman-Monteith (Monteith 1965).

Rovnice (1) $\lambda PET = \frac{\Delta(Rn-G)+\rho \cdot c(e_s-e_a)/r_a}{\Delta+\gamma\left(1+\frac{r_s}{r_a}\right)}$

kde **Δ** označuje derivaci tlaku nasycené vodní páry podle teploty vzduchu [kPa.°C^{-1}], **Rn** radiační bilanci povrchu [MJ.m^{-2}.d^{-1}], **G** tok tepla v půdě [MJ.m^{-2}.d^{-1}], **γ** psychrometrickou konstantu [kPa.°C^{-1}], $\mathbf{e_s}$–$\mathbf{e_a}$ deficit tlaku vodní páry [kPa] ($\mathbf{e_s}$ tlak nasycené vodní páry; $\mathbf{e_a}$ aktuální tlak vodní páry), **λ** měrné teplo vypařování [MJ.kg^{-1}], **ρ** hustotu vody [kg.l^{-1}], c specifické teplo vzduchu [kJ. kg^{-1}.°C^{-1}], $\mathbf{r_s}/\mathbf{r_a}$ poměr aerodynamického a povrchového odporu vzduchu [s.m^{-1}].

Pro účely práce byly také nezbytné výpočty akumulace a tání sněhové pokrývky, které byly získány na základě metody degree-day (Gupta 2001). Část práce je zaměřena na měření množství vody v půdě. Půdní vlhkost byla měřena na dvou místech na svahu s výskytem podzolů v hloubkách 20 cm a 60 cm pod povrchem pomocí tensiometrů (T8, UMS). Hladina podzemní vody v rašelinných částech byla měřena pomocí hydrostatických ponorných sond (TSH22, Fiedler) na třech místech. Pětileté měření zahrnovalo dva velmi suché roky (2015, 2017), ale srážkově nadprůměrný rok v rámci měření nebyl zaznamenán.

Veškeré měřené parametry byly vstupem do hydrologického modelu, který je založen na výpočtu dominantního preferenčního proudění (Boorman, 1995; Scherrer, Naef 2003). Využitý model vychází ze schématu HBV modelu (Bergström, 1992). Ten byl použit ve své standardní podobě pro subpovodí s minerálními půdami (40 % plochy povodí). Pro subpovodí s organogenními půdami (60 % plochy povodí) byl následně upraven dle konceptu akrotelm – katotelm, který byl navržen ve studii Ingram (1978). Obě subpovodí byla reprezentována vlastní strukturou modelu, která se lišila zejména v charakteru odtokové odezvy na srážkovou událost. Základní schéma modelu rozděleného do subpovodí ukazuje obr. 6.



Obr. 6: Schéma modelu rozděleného do subpovodí (SB) – minerálních půd podzolů (PZ) a horského vrchoviště (PB) (Vlček et al., 2020)

3.3 Hodnocení dynamiky výšky hladiny podzemní vody

Při hodnocení dynamiky výšky hladiny podzemní vody byla použita meteorologická data ze stanice Modrava. Ta byla využita nejen pro analýzu srážek, ale i výpočtu potenciální evapotranspirace dle metody Penman-Monteith (Penman 1948). Taktéž byla využita data z váhového sněhoměru na Rokytecké slati, z nichž byly použity hodnoty vodní hodnoty sněhu (SWE) pro hodnocení režimu podzemní vody během tání sněhové pokrývky. Nejvýznamnější část práce byla zaměřena na dynamiku hladiny podzemní vody v rašeliništi. Ta byla měřena na pěti místech (2x rašelinný les, 2x kleč, 1x centrální ostřicová část) v rašeliništi tak, aby reprezentovala daný vegetační typ. Hydrostatické ponorné senzory (TSH22, Fiedler) pro měření hladiny podzemní vody byly umístěny v trubkách v hloubce 70 cm. Veškeré přístroje měřily v desetiminutovém intervalu.

Práce navazuje na detailní pedologický průzkum a výpočty retenčního potenciálu horského vrchoviště předcházejícího výzkumu (Vlček et al., 2012). Na základě výsledků měření charakteristik vlhkosti půdy byly poté pro účely této studie stanoveny rovnice pro výpočet retenčního potenciálu během epizod intenzivního deště, tání sněhu (rovnice 2) a sucha (rovnice 3).

Rovnice (2) $RP = \frac{GWL(b)}{GWL(e)} * [(VWC_{max} - VWC_{mean}) * 1000]$

Rovnice (3) $RP = \frac{GWL(b)}{GWL(e)} * [(VWC_{mean} - VWC_{min}) * 1000]$

kde **RP** označuje retenční potenciál [l/m^2], **GWL(b)** hladinu podzemní vody na počátku sledované události [mm], **GWL(e)** hladinu podzemní vody na konci sledované události [mm], **VWC_{max}** maximální objemovou vlhkost [g/cm^3], **VWC_{mean}** průměrnou objemovou vlhkost [g/cm^3], **VWC_{min}** minimální objemovou vlhkost [g/cm^3].

Při výpočtech bylo předpokládáno, že na začátku dané epizody dosahovala objemová vlhkost v povodí průměrné hodnoty (0,7 g/cm^3) a z této hodnoty poté během srážkových epizod a tání sněhu rostla na maximum (0,95 g/cm^3). Naopak během suchých epizod z průměrné hodnoty klesala na minimum (0,5 g/cm^3). Dané hodnoty vychází z detailního pedologického průzkumu, který byl realizován na stejné zájmové lokalitě (Vlček et al., 2012).

3.4 Posouzení fyzikálních a chemických vlastností rašelinných vod

Základem studie bylo měření průtoků, hladin podzemní vody a fyzikálních a chemických vlastností povrchových i podpovrchových vod. Měření probíhala během jedné vegetační sezony (2018), kdy bylo provedeno celkem 21 terénních měření v týdenních intervalech. Během měření byla podchycena období vysoké saturace rašeliniště po tání sněhové pokrývky, vysokých srážkových úhrnů i suché a horké období v létě.

Pro realizaci terénních průzkumů byla vybrána tři různá rašelinná stanoviště (rašelinný les, těžená část rašeliniště, lagg) v rašelinném komplexu Mezilesní slati. V rámci každého stanoviště byl do závěrového profilu toku instalován Ponceletův měrný přeliv. Pro výpočet průtoku byla použita rovnice pro Basinův přeliv (rovnice 4) s využitím koeficientu pro Ponceletův přeliv (rovnice 5).

Rovnice (4) $\mathrm{Q} = \mathrm{m.b.}\sqrt{2.g}.h^{3/2}$

Rovnice (5) $m = \left[0.405 + \frac{0.003}{h} - 0.03\left(1 - \frac{b}{B}\right)\right][1 + 0.55)(\frac{b}{B})^2(\frac{S_0}{S})^2]$

kde **Q** reprezentuje průtok [m^3.s^{-1}], **m** součinitel přepadu pro Ponceletův přeliv, **b** délku přelivné hrany [m], **B** délku přelivu [m], **g** gravitační zrychlení [m.s^{-1}], **h** výška vody nad přepadem [m], $\mathbf{S_0}$ plochu průtočného profilu [m^2], **S** plochu přelivné části [m^2] (Šráček, Kuchovský 2003).

Každý experimentální vodní tok byl osazen 4 trubkami, v nichž byla ručně měřena hladina podzemní vody. Trubky byly zasazeny do rašeliny ve vzdálenosti 2 m a 5 m od vodního toku po obou jeho březích. Fyzikální a chemické vlastnosti byly měřeny kalibrovanými terénními měřicími systémy ve všech třech experimentálních vodních tocích i ve všech místech měření hladiny podzemní vody (celkem 12 míst měření). Mezi sledovanými parametry bylo pH, elektrická konduktivita, teplota vody a rozpuštěný kyslík. Veškerá meteorologická data, která byla využita ve studii, byla naměřena na meteorologické stanici Modrava (Fiedler, data v desetiminutových intervalech). V práci je také hodnocena dynamika hladiny podzemní vody v závislosti na denní potenciální evapotranspiraci. Pro výpočet potenciální evapotranspirace byla použita rovnice Penman-Monteith (Penman 1948). Pro zjištění vzájemných vztahů mezi sledovanými parametry byly použity základní statistické metody včetně výpočtů Spearmanových korelačních koeficientů na hladině spolehlivosti $p<0,05$.

3.5 Vliv revitalizačních opatření horských vrchovišť na hladinu podzemní vody

Pro hodnocení vlivu revitalizačních opatření horských vrchovišť na hladinu podzemní vody byl vybrán experimentální odvodňovací kanál v rámci Rokytecké slati. Tento kanál byl ve svém ústí přehrazen dvěma dřevěnými přehrádkami, zbytek kanálu zůstal bez opatření.

Podél odvodňovacího kanálu byly rozmístěny trubky, jež byly zasazeny do rašeliny v hloubce 1 m. Rozmístěny byly ve třech řadách souběžně s odvodňovacím kanálem (v každé řadě bylo 9 míst měření). První řada navazovala přímo na odvodňovací kanál, mezi dalšími řadami byl rozestup 3 m. Vznikla tak pravidelná síť 27 bodů, v nichž byla hladina podzemní vody měřena ručně. Celkem proběhlo 28 měření v období od 14. srpna do 31. října 2014. Jednotlivá místa měření byla geodeticky zaměřena. Pro mapové výstupy s ukázkou prostorového rozložení hladiny podzemní vody během sledovaných epizod byla využita interpolační metoda „natural neighbor“. Data hladiny podzemní vody byla analyzována pomocí základních statistických metod. Zároveň byla dávána do kontextu příčinných meteorologických faktorů. Pro tyto účely byla vypočítána potenciální evapotranspirace dle Penman-Monteith (Penman 1948). K vyjádření předchozího nasycení povodí byl využit index předchozích srážek API (antecedent precipitation index), uvažovaný pro pět předcházejících dní (rovnice 6).

Rovnice (6) $API = \sum 0{,}93^{i} \cdot P_{i}$

kde **i** vyjadřuje počet uvažovaných dní (počítáno zpětně), **P** denní úhrn srážek v i-tém dni sledovaného období [mm] (Mishra, Singh 2003).

4. Výsledky a jejich diskuze

Hydrologická funkce horských vrchovišť zahrnuje zejména problematiku retence vody, akumulaci vody a způsob formování odtoku. Oba procesy jsou silně ovlivněny specifickými fyzickogeografickými podmínkami rašelinných oblastí.

Obecně jsou horská vrchoviště charakteristická svou rychlou odtokovou odezvou během srážkových epizod a zpravidla nízkou dotací vodních toků v období sucha. Takové výsledky jsou známy v případě šumavských slatí (Janský, Kocum 2008; Čurda et al., 2011, Kocum 2012; Vlček 2017) i ze zahraniční literatury (Boorman et al., 1995; Evans et al., 1999, Bragg 2002; Binet et al., 2013). Většina výzkumů však probíhala na malých subpovodích, kde výrazně dominovaly organogenní půdy. Výsledky z experimentálních povodí taktéž naznačují, že příspěvek horských vrchovišť k celkové vodní bilanci je značný především během intenzivních srážek, kde převažuje odtok z organogenních půd. Naopak během sušších období převažoval odtok spíše z minerálních půd. V obou případech však záleželo zejména na aktuálním nasycení povodí a fyzikálních vlastnostech půdy. Tyto faktory rozhodují, za jakých okolností dojde k povrchovému odtoku, kolik vody se může infiltrovat do půdy během srážkové události nebo jak dlouho se dokáže voda v půdě zadržet během suchých period (Vlček 2017). Hlavním faktorem ovlivňujícím hydrologické procesy v rašeliništi je proto výška hladiny podzemní vody. Místa s vysokou hladinou podzemní vody zpravidla vykazovala vyšší rozkolísanost průtoků vodních toků. Důležitým faktem je také odlišný způsob podpovrchového odtoku ve sledovaných subpovodích. V horském vrchovišti převažuje mělký podpovrchový odtok. S ohledem na velké množství zadržované vody je nutné v rašeliništi uvažovat i pístové proudění (piston flow), proudění vody preferenčními cestami v rašelině (pipe flow), případně i přímý povrchový odtok.

Klíčovým prvkem ovlivňujícím hydrologické procesy, ale i celkový vývoj horských vrchovišť, je tedy výška a stabilita hladiny podzemní vody. Ta je řízena zejména výškou srážek a evapotranspirací. Její pohyb v akrotelmu ovlivňuje akumulaci organické hmoty, vegetační poměry a celkovou vodní bilanci v povodí (Allott et al., 2009; Lindsay et al., 2014). Výrazný vliv na výšku a kolísání hladiny podzemní vody má vegetační typ příslušného stanoviště, jelikož podmiňuje teplotní poměry a má také zásadní vliv na evapotranspiraci daného území (Bufková, Stíbal 2012; Kučerová et al., 2009; Holden et al., 2011). Nejvýše se zpravidla hladina vyskytuje v centrální části vrchoviště. Tak tomu bylo i v rámci sledovaného povodí (v průměru 10 cm pod povrchem), kde je tato část tvořena ostřicovými porosty. Avšak tato část byla zároveň nejnáchylnější k letním poklesům hladiny, stejně jako bylo pozorováno a popsáno v Binet et al., (2013). Současně se jedná o velmi citlivé části horského vrchoviště, jelikož jejich dlouhodobé vysušení může vést k sukcesním změnám (Kučerová et al., 2009; Kværner, Snilsberg 2011). Rašelinný les vykazoval výrazně nižší hadiny podzemní vody, ovšem v rámci tohoto biotopu se ukázaly značné rozdíly mezi místy měření. Rozdíl průměrných hladin podzemní vody během sledovaného období byl více než 25 cm. Les však vykazoval poměrně vysokou stabilitu průtoků. Část vrchoviště pokrytá klečovými porosty dosahovala průměrné hladiny podzemní vody přibližně 20 cm. Datové soubory v souhrnu disponovaly velmi podobnou variabilitou, lze tedy usuzovat, že hladina podzemní vody v různých vegetačních typech se sice nachází v jiné úrovni, avšak její kolísání osciluje ve velmi podobném rozpětí. Dané poznatky mají oporu i v literatuře, obdobné závěry přináší Kučerová et al., (2009); Bufková et al, 2010; Labadz et al., (2010). Nicméně existuje velké množství faktorů ovlivňujících hladinu podzemní vody daného mikrostanoviště, například topografie, fyzikální vlastnosti rašeliny, hydraulická vodivost a další (Allott et al., 2009).

Výrazný vliv na hladinu podzemní vody v případě Šumavy měly i antropogenní zásahy z minulosti. Odvodněné nebo těžené části rašeliniště vykazují nízkou hladinu podzemní vody a její vyšší rozkolísanost. Tato skutečnost je zdokumentována v řadě studií ze Šumavy i ze zahraniční (Bufková 2006; Bufková 2012; Bufková et al., 2010; Holden et al., 2004; Holden et al., 2011; Worrall et al., 2007). V experimentálním povodí byl prokázán pozitivní vliv revitalizačních opatření, jejichž výsledkem bylo zvýšení hladiny podzemní vody v průměru o 9 cm. Zároveň byl sledován i plošný vliv a dosah odvodnění i revitalizací. Jejich účinek je měřitelný i v dosahu přibližně 6 metrů, vzdálenost ovšem závisí na velkém množství parametrů. Byla pozorována i vyšší rozkolísanost průtoků narušených stanovišť. Nicméně v případě ovlivnění odtokového procesu není na danou problematiku jednotný názor (Janský, Kocum 2008; Čurda et al., 2011; Holden et al., 2011). Různé výsledky jsou pravděpodobně způsobeny specifickými fyzickogeografickými vlastnostmi sledovaných povodí, které společně formují odtokový proces.

Analýza retenčního potenciálu rašeliniště poukázala zejména na význam iniciální hladiny podzemní vody. Schopnost infiltrovat vodu je úměrná předchozímu nasycení rašeliniště. Podíl přijaté vody ze srážek se proto pohyboval ve velkém rozmezí. Po dosažení maximální saturace půdy daného vegetačního typu dochází k vyčerpání retenčního potenciálu a následnému rychlému nárůstu odtoku. Tento krátkodobý retenční potenciál dosáhl maximální hodnoty až 42 l/m^2. Horské vrchoviště tedy může za určitých podmínek tvořit plochu se značným retenčním potenciálem, nicméně v případě vysokého předchozího nasycení se tento retenční prostor neuplatňuje (Kocum et al., 2018). I v zimním období dochází k výrazné retenci vody. Maximální krátkodobý objem infiltrované vody z tajícího sněhu v součtu za celé sledované povodí dosahoval až k 10 000 m^3 vody, v závislosti na aktuálních podmínkách ve vrchovišti. Jedná se však o krátkodobou retenci, jelikož následně dochází k poklesům hladiny a tudíž i k povrchovému a podpovrchovému odtoku.

Posledním aspektem práce bylo hodnocení fyzikálně-chemických vlastností rašelinných vod. Ombrotrofní vrchoviště sice mají poměrně stálé hydrochemické vlastnosti, v rámci dílčích částí nebo mikrotopografických prvků ale mohou být zaznamenány rozdíly (Faubert 2004). Výrazně nízké pH a vysoké teploty u povrchové i podzemní vody byly naměřeny v narušené části rašeliniště. Nízké pH souvisí zejména s absencí vegetačního krytu, což při srážkách usnadňuje vymývání kyselých iontů (Wind-Mulder et al., 1996). Stejný proces pak pravděpodobně způsobuje vyšší hodnoty elektrické vodivosti, avšak ta v těžené části vykazuje velmi vysokou variabilitu. Zároveň je nutné upozornit, že hodnoty elektrické konduktivity jsou závislé i na dalších faktorech a naměřené hodnoty se mohou v rámci rašeliniště mírně lišit (Ponziani et al., 2011). Hodnoty rozpuštěného kyslíku nevykazovaly při porovnání sledovaných částí rašeliniště (těžená část, lagg, rašelinný les) výrazné rozdíly. Zároveň byly identifikovány některé vazby mezi sledovanými parametry. Výrazně se projevil vztah mezi pH a množstvím vody v povodí. Se snižující se hladinou podzemní vody a průtoky roste pH, podobně jako bylo popsáno v Seibert et al., (2009). Řada vazeb se projevila hlavně v narušené části rašeliniště. Silné korelace zde byly pozorovány mezi nízkým pH a vysokými hodnotami elektrické vodivosti, nízkou hladinou podzemní vody a nízkou teplotou vody i mezi nízkým pH a nízkou elektrickou vodivostí. Podobné vztahy v narušených částech rašeliniště byly popsány ve studiích Wind-Mulder et al., (1996); Ponziani et al., (2011).

5. Závěr

Práce hodnotí vodní bilanci a odtokové poměry horských vrchovišť, s přihlédnutím zejména k hladině podzemní vody, která je klíčovým faktorem ovlivňujícím stav a vývoj těchto cenných stanovišť.

Měření hladiny podzemní vody společně s analýzou hlavních faktorů ovlivňujících její výšku (srážky, potenciální evapotranspirace, nasycení povodí, vegetační typ) poukázalo na výraznou dynamiku, jejíž znalost je nezbytná pro hodnocení hydrologických procesů probíhajících v povodí. Obecné poznatky o rychlé odtokové odezvě během srážkových událostí jsou sice v rámci rašelinných komplexů známy, nicméně vše závisí na retenčním potenciálu, který je určen právě výškou hladiny podzemní vody. Stejně tak je nutné brát v úvahu specifika odtokového procesu, která se objevují zejména v oblastech s výskytem hydromorfních a organogenních půd (přímý povrchový odtok, pipe flow, piston flow).

Výrazný retenční prostor rašelinišť a schopnost akumulovat vodu byl zaznamenán i během tání sněhové pokrývky, naopak během suchých období dochází k poklesům hladiny podzemní vody a horské vrchoviště se podílí jen velmi málo na bazálním odtoku. Zjištěné poznatky je však nutno vnímat v kontextu daného půdního a vegetačního typu, či příslušné části rašelinného komplexu, jelikož při porovnání jednotlivých částí bylo poukázáno na výrazné rozdíly a specifika. To platí i pro popsané fyzikálně-chemické vlastnosti povrchových a podpovrchových vod horského vrchoviště a jejich vzájemné korelace, které byly identifikovány v rámci sledovaného povodí. Významným specifikem rašelinišť na Šumavě jsou historické antropogenní zásahy, které negativně ovlivnily nejdříve hydrologické, a poté i vegetační poměry stanovišť. V revitalizovaných místech experimentálního odvodňovacího kanálu došlo k výraznému zvýšení hladiny podzemní vody, zároveň byl pozorován i značný plošný dosah těchto opatření.

Významné plošné zastoupení organogenních a hydromorfních půd v oblasti centrální Šumavy je tedy specifikem, které výrazně ovlivňuje odtokový proces. Horská vrchoviště jsou sice místa s nejvyšší retenční kapacitou v krajině, nicméně uplatnění retenčního prostoru, množství a způsob odtoku či dynamika hladiny podzemní vody jsou značně závislé na konkrétních fyzickogeografických podmínkách daného stanoviště. Zjištěné poznatky tak přispívají k pochopení hydrologického režimu a k celkovému poznání horských vrchovišť.

6. Zdroje

Allott, T. E. H., Evans, M. G., Lindsay, J. B., Agnew, C. T., Freer, J. E., Jones, A., Parnell, M. (2009): Water tables in Peak District blanket peatlands. Moors for the future. Edale, Derbyshire, 51 s.

Armstrong, A., Holden, J., Kay, P., Francis, B., Foulger, M., Gledhill, S., McDonald, A. T., Walker, A. (2010): The impact of peatland drain-blocking on dissolved organic carbon loss and discolouration of water : results from a national survey. Journal of Hydrology, 381, 1-2, 112-120.

Baden, W., Eggelsmann, R. (1970): Hydrological budget of high bogs in the Atlantic region. Proceedings of the 3rd international peat congress 1968, Quebec. Ottawa: Department of Energy, Mines and Resources, 260–311.

Bergsma, B., Quinlan, C. (2009): Sifton Bog ESA – Conservation Master Plan 2009–2019. Upper Thames River Conservation Authority, Parks, London, 212 s.

Bergström, S. (1992): The HBV Model: Its Structure and Applications. Swedish Meteorological and Hydrological Institute (SMHI), Norrköping, 35 s.

Binet, S., Gogo, S., Laggoun-Défarge, F. (2013): A water-table dependent reservoir model to investigate the effect of drought and vascular plant invasion on peatland hydrology. Journal of Hydrology, 499, 30, 132-139.

Boorman, D. B., Hollis, J. M., Lilly, A. (1995): Hydrology of soil types: a hydrologically-based classification of the soils of the United Kingdom, Institute of Hydrology report no. 126, 146 s.

Bragg, O. M. (2002): Hydrology of peat-forming wetlands in Scotland. Science of The Total Environment, 294, 1–3, 111-129.

Brom, J., Procházka, J., Rejšková, A. (2009): Evaluation of functional properties of various types of vegetation cover using remotely sensed data analysis. Soil and Water Research, 4 (special issue 2), 49-58.

Bufková, I. (2006): Revitalizace šumavských rašelinišť. Zprávy České Botanické Společnosti, 41, 21, 181–191.

Bufková, I. (2012): Program revitalizace šumavských mokřadů a rašelinišť – Koncepce programu, aktualizovaná verze listopad 2012, Národní park Šumava, 33 s.

Bufková, I. (2013): Náprava narušeného vodního režimu rašelinišť v národním parku Šumava. Ochrana přírody, 2, 17-19.

Bufková, I., Stíbal, F. (2012): Restoration of drained mires in the Šumava National Park. In: Jongepierová, I., Pešout, P., Jongepier, J., W., Prach, K. (Eds.): Ecological restoration in the Czech Republic, Nature Conservation Agency of the Czech Republic, Praha, 147 s.

Bufková, I., Stíbal, F., Mikulášková, E. (2010): Restoration of drained mires (Šumava National Park, Czech republic). Proceedings 7th European Conference on Ecological Restoration Avignon, France, s. 23-27.

Burke, W. (1975): Aspects of the hydrology of blanket peat in Ireland. Hydrology of marsh-ridden areas. Proceedings of the Minsk symposium, IAHS Studies and Reports in Hydrology, 19, Unesco Press, Paris, 171–82.

Campbell, D. R., Rochefort, L. (2001): La végétation: gradients. In: Payette, S., Rochefort, L. (Eds.): Écologie des tourbières du Québec-Labrador. Québec, Canada: Les Presses de l'Université Laval, 129-140.

Čurda, J., Janský, B., Kocum, J. (2011): Vliv fyzicko-geografických faktorů na extremitu povodní v povodí Vydry. Geografie, 116, 3, 335-353.

Dingman, S., L. (2015): Physical Hydrology. Waveland Press, Inc., Long Grove, 643 s.

Doležal, T., Vlček, L., Kocum, J., Janský, B. (2020): Hydrological regime and physico-chemical water properties of various types of peat bog sites: case study of Mezilesní peat bog, Šumava Mts. Geografie, 125, 1, 21 - 46.

Estop-Aragonés, C., Klaus-Holger, K., Blodau, CH. (2012): Controls on in situ oxygen and dissolved inorganic carbon dynamics in peats of a temperate fen. Journal of Geophysical Research, 117, Issue G2, 1-14.

Evans, M. G., Burt, T. P., Holden, J., Adamson, J. K. (1999): Runoff generation and water table variations in blanket peat: evidence from UK data spanning the dry summer of 1995. Journal of Hydrology, 221, 3–4, 141–160.

Faubert, P. (2004): The effect of long-term water level drawdown on the vegetation composition and CO2 fluxes of a boreal peatland in central Finland. Universita Laval, Département de phytologie Faculté des Sciences de l'Agriculture et de l'Alimentation, M.Sc. thesis, 67 s.

Ferda, J. (1969): Hydrologická a klimatická funkce československých rašelinišť. Česká akademie zemědělská, Praha, 358 s.

Ferda, J., Hladný, J., Bubeníčková, L., Pešek, L. (1971): Odtokový režim a chemismus vod v povodí Horní Otavy se zaměřením na výskyt rašelinišť, Sborník prací HMÚ, 17, Praha, 22-126.

Gupta, R., S. (2001): Hydrology and hydraulic systems. Waveland Press 2nd edition, Long Grove, 867 s.

Haapalehto, T. O., Vasander, H., Jauhiainen, S., Tahavainen, T., Kotiaho, S. J. (2011): The Effects of Peatland Restoration on Water-Table Depth, Elemental Concentrations, and Vegetation: 10 Years of Changes. Society forEcological Restoration International. Restoration Ecology 19, 5, 587–589.

Holden, J., Chapman, J. P., Labadz, J. (2004): Artificial drainage of peatlands: Hydrological and hydrochemical process and wetland restoration. Progress in Physical Geography, 28, 1, 95-123.

Holden, J., Wallage, Z., Lane, S., McDonald, A. (2011): Water table dynamics in undisturbed, drained and restored blanket peat. Journal of Hydrology, 402, 1, 103–114.

Howie, S. A., Meerveld, I. T. (2011): The Essential Role of the Lagg in Raised Bog Function and Restoration: A Review. Wetlands, 31, 3, 613 – 622.

Hümann, M., Schüler, G., Müller, C., Schneider, R., Johst, M., Caspari T. (2011): Identification of runoff processes – The impact of different forest types and soil properties on runoff formation and floods. Journal of Hydrology, 409, 3-4, 637–649.

Ingram, H.A.P. (1978): Soil layers in mires: function and terminology. Journal of Soil Science, 29, 2, 224–227.

Janský, B. (2004): Retence vody v povodí. In: Langhammer, J. (Eds.): Sborník příspěvků semináře grantu GAČR 205/03/Z046 Hodnocení vlivu změn přírodního prostředí na vznik a vývoj povodní, 59 – 70.

Janský, B., Kocum, J. (2008): Peat bogs influence on runoff process: case study of the Vydra and Křemelná River basins in the Šumava Mountains, southwestern Czechia. Geografie – Sborník ČGS 113, 4, 383–399.

Ketcheson, S. J., Price, J. S. (2011): The Impact of Peatland Restoration on the Site Hydrology of an Abandoned Block Cut Bog. Wetlands 31, 6, 1263–1274.

Kocum, J. (2012): Tvorba odtoku a jeho dynamika v pramenné oblasti Šumavy. Univerzita Karlova v Praze, Přírodovědecká fakulta, katedra fyzické geografie a geoekologie. Disertační práce 206 s.

Kocum, J., Janský, B., Vlček, L., Doležal, T. (2018): Hydrological Function of a Midlatitude Headwater Peatland. Chapter 8, s. 141 – 164. In Topcuğlu, B., Turan, M. (Eds.): Peat. IntechOpen, Rijeka, Croatia, 164 s.

Kučerová, A., Kučera, T., Hájek, T. (2009): Mikroklima a kolísaní hladiny podzemní vody v centrální části Rokytecké slati. In: Černý, D., Dvořák, L. (Eds.): Weitfallerské slatě. Sborník z výzkumu na Šumavě, 2. Správa NP a CHKO Šumava, Vimperk, 103 s.

Kværner, J., Snilsberg, P. (2011): Groundwater hydrology of boreal peatlands above a bedrock tunnel – Drainage impacts and surface water groundwater interactions. Journal of Hydrology 403, 3, 278-291.

Labadz, J., Allott, T., Evans, M., Butcher, D., Billett, M., Stainer, S., Yallop, S., Jones, P., Innerdale, M., Harmon, N., Maher, K., Bradbury, R., Mount, D., O´Brien, H., Hart, R. (2010): Peatland hydrology. Draft scientific review, commissioned by the IUCN UK Peatland Programmes Commission of Inquiry on Peatlands, 52 s.

Lindsay, R., Birnie, R., Clough, J. (2014): Peat Bog Ecosystems: Structure, Form, State and Condition. IUCN UK Committee Peatland Programme Briefing Note No 2, s. 1-7.

Mishra, S., K., Singh, V., P. (2003): Soil Conservation Service Curve Numer (SCS-CN) Methodolgy. Dordrecht, Kluwer Academic Publisher, 511 s.

Moklyak, V. I., Kubyshkin, G. P., Karkutsiev, G. N. (1975): The effect of drainage works on streamflow. Hydrology of marsh-ridden areas, Proceedings of the Minsk symposium. IAHS Studies and Reports in Hydrology, 19, Unesco Press, Paris, 439-446.

Monteith, J., L. (1965): Evaporation and the environment. In: The state and movement of water in living organisms. Academic Press, for the Society for Experimental Biology, Cambridge, 205-234.

Penman, H. L. (1948): Natural evaporation from open water, bare soil and grass. Proceedings of the Royal Society of London. Series A, Mathematical and Physical Sciences, 193, 1032, 120–145.

Ponziani, M., Slob, E. C., Ngan-Tillard, D. J. M., Vanhala, H. (2011): Influence of water content on the electrical conductivity of peat. International Water Technology Journal, 1, 1, 14 -21.

Pošta, P., Janský, B., Šobr, M. (2003): Organogenní jezera v České republice in. Janský, B., Šobr., M. (Eds.) Jezera České republiky. Univerzita Karlova v Praze, Přírodovědecká fakulta, katedra fyzické geografie a geoekologie, 170 s.

Price, J. S., Heathwaite, A. L., Baird, A. J. (2003): Hydrological processes in abandoned and restored peatlands: An overview of management approaches. Wetlands Ecology and Management 11, 1, 65-83.

Rubec, C. (2005): The decade of the bog 1994-2004: Global progress on peatland wise use and conservation. Suo, 56, 1, 19-26.

Seibert, J., Grabs, T., Köhler, S., Laudon, H, Winterdahl, M., Bishop, K. (2009): Linking soil- and stream-water chemistry based on a Riparian Flow-Concentration Integration Model. Hydrology and Earth System Science, 13, 12, 2287–2297.

Scherrer, S., Naef, F. (2003): A decision scheme to indicate dominant hydrological flow processes on temperate grassland, Hydrol. Process., 17, 2, 391–401,

Starostová, M. (2012): Měření srážek totalizátory na Šumavě. Meteorologické zprávy, 65, 6, 180–183.

Šefrna, L. (2004): Pedologická charakteristika povodí Otavy ve vztahu k povodím. In: Langhammer, J. (Eds.): Sborník příspěvků semináře grantu GAČR 205/03/Z046 Hodnocení vlivu změn přírodního prostředí na vznik a vývoj povodní, 196-212.

Špirhanzl, J. (1951): Rašelina, její vznik, těžba a využití. Přírodovědecké nakladatelství, Praha, 356 s.

Šráček, O., Kuchovský, T. (2003): Základy hydrogeologie. Masarykova Univerzita v Brně, 1. Vydání, 177 s.

Vlček, L. (2017): Retence vody v půdách horských oblastí na příkladu Šumavy. Univerzita Karlova v Praze, Přírodovědecká fakulta, katedra fyzické geografie a geoekologie. Disertační práce, 127 s.

Vlček, L., Kocum, J., Janský, B., Šefrna, L., Blažková, Š. (2016): Influence of peat soils on runoff process: case study of Vydra River headwaters, Czechia. Geografie, 121, 2, 235-253.

Vlček, L., Kocum, J., Janský, B., Šefrna, L., Kučerová, A. (2012): Retenční potenciál a hydrologická bilance horského vrchoviště: případová studie Rokytecké slatě, povodí horní Otavy, JZ. Česko. Geografie 117, 4, 395–414.

Vlček, L., Šípek, V., Kofroňová, J., Kocum, J., Doležal, T**.**, Janský, B. (2020): Understanding runoff formation in a catchment of Peat bog and Podzol hillslopes. Journal of Hydrology. (in review)

Wilson, L., Wilson, J., Holden, J., Johnstone, I., Armstrong, A., Morris, M. (2011): The impact of drain blocking on an upland blanket bog during storm and drought events, and the importance of sampling-scale. Journal of Hydrology, 404, 3-4, 198-208.

Wind-Mulder, H. L., Rochefort, L., Vitt, D. H (1996): Water and peat chemistry comparisons of natural and post-harvested peatlands across Canada and their relevance to peatland restoration. Ecological Engineering, 7, 3, 161 – 168.

Worrall, F., Armstrong, A., Holden, J. (2007): Short-term impact of peat drain-blocking on water colour, dissolved organic carbon concentration, and water table depth. Journal of Hydrology, 337, 3, 315-325.

Worrall, F., Burt, T., Adamson, J. (2006): Long-term changes in hydrological pathways in an upland peat catchment—recovery from severe drought?, Journal of Hydrology, 321, 1, 5-20.

Yvon-Durocher, G., Jones, I., Trimmer, M., Woodward, G., Montoya, J.M., (2010): Warming alters the metabolic balance of ecosystems. Philosophical Transactions of the Royal Society B: Biological Sciences, 365, 1549, 2117-2126.

7. Články sepsané v rámci disertační práce

Disertační práce byla sepsána formou souboru článků. Soupis jednotlivých článků včetně uvedení pracovního podílu autora předkládané disertační práce je uveden v tab. 2.

Tab. 2: Soupis článků sepsaných v rámci disertační práce

Kategorie	**Citace článku**	**Pracovní podíl (%)**
Články v databázi SCOPUS	**Doležal, T**., Vlček, L., Kocum, J., Janský, B. (2017): Evaluation of the influence of mountain peat bogs restoration measures on the groundwater level: case study Rokytka peat bog, the Šumava Mts., Czech Republic. AUC Geographica, 52, 2, 1 - 10.	80
Články v databázi WoS	**Doležal, T.**, Vlček, L., Kocum, J., Janský, B. (2020): Hydrological regime and physico-chemical water properties of various types of peat bog sites: case study of Mezilesní peat bog, Šumava Mts. Geografie, 125, 1, 21 - 46.	80
	Vlček, L., Šípek, V., Kofroňová, J., Kocum, J., **Doležal, T.**, Janský, B. (2020): Understanding runoff formation in a catchment of Peat bog and Podzol hillslopes. Journal of Hydrology. (in review)	30
	Doležal, T., Vlček, L., Kocum, J., Janský, B. (2020): Influence of vegetation on groundwater level dynamics and calculation of retention ability of Rokytecká slať peat bog, Šumava Mountains, Journal Mires and Peat. (in review)	80
Články v recenzovaném periodiku	Kocum, J., Janský, B., Vlček, L., **Doležal, T.** (2018): Hydrological Function of a Midlatitude Headwater Peatland. Chapter 8, s. 141 – 164. In Topcuğlu, B., Turan, M. (Eds.): Peat. IntechOpen, Rijeka, Croatia, 164 s.	30

7.1 Formování odtoku v oblasti organogenních a minerálních půd

Vlček, L., Šípek, V., Kofroňová, J., Kocum, J., Doležal, T., Janský, B., (2020): Understanding outflow formation in basin with two dominated soil systems. Journal of Hydrology (in review).

Title: Understanding runoff formation in a catchment with Peat bog and Podzol hillslopes

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Abstract

Hydrological behavior of an ombrogenous raised peat bog has been observed and described in many studies; however, not in complexity with other soils. This research deals with the hydrological function of peat bog in a catchment where peat bog (formed by Histosol or other hydromorphic soils) covers only a part of the area (40–60%). Two soil types, creating two main hillslopes of the experimental catchment in this study, form the dominant soil types (Podzol and Histosol) in the Šumava Mountains, Czechia. A modified HBV model was used to estimate the contribution of each soil type to common outflow and for the estimation of the water balance. According to previous research and field observations, dominant hydrological processes were described for each hillslope (soil). The HBV model was used for the quantification of a ratio between fast and slow flow at Peat bog hillslope and Podzol hillslope. At Peat bog hillslope, the majority of outflow (67%) was formed from the upper soil layer (Acrotelm). In the mineral soil hillslope, a larger portion of runoff was generated from the lower soil layers or bedrock interface (61%). Peat bog contributes to a stream mainly during rainfall events; however, the model showed also significant deep percolation at the Peat bog hillslope and considerable contribution to baseflow during a year. Generally, more precipitation water was turned by the model into runoff at the Peat bog hillslope, which was also exhibited a lower rate of actual evapotranspiration (21% of precipitation), compared to 29% in the case of Podzol hillslope. If we consider land-use changes in this locality in terms of expanding or reducing peat areas (draining, drains damming, droughts, etc.), this model could sufficiently estimate the hydrological behavior of local streams and thus can be potentially used in hydrological planning by local authorities.

Key words: water balance, hydrological model, soil hydrology, Šumava Mts., Peat bog, runoff formation

1 Introduction

Theories on the effect of peat bogs on runoff formation under varying conditions differ widely and repeatedly among researchers. The problems of drainage, especially diking of former drainage channels, have become the field for broad debates within the literature (Baird, 1997, Conway & Millar, 1960; McDonald, 1973). Most of the studies have been carried out in catchments covered only by Histosols (peat) which are the dominant soil type creating peat bogs. However, catchments are mostly formed by other soil types which, together, comprise a complex system that influences the runoff formation. Streamflow in peaty catchments is characterised by its quick rise and fall, and high variability; that is, very low baseflow during dry periods and spiky storm hydrographs caused by heavy rainfall events. This behaviour has been described in detail in many hydrological studies (e.g. Bragg, 2002; Evans et al., 1999; Holden et al., 2001; Holden & Burt, 2003). Beside common mineral soils (Cambisols, Podzol, etc.), peat (Histosol) is known for frequent overland flow or near-surface flow which transfer quickly rainfall event to a stream. However, most of the above-mentioned studies have focused on pure peat areas only.

Nevertheless, catchments are predominantly comprised of several types of soil or vegetation which can lead to varied subsurface flow and runoff formation (Hümann et al., 2011). One of the most common locations in Czechia where peat bogs occur is the Šumava Mts. where they affect both water quality and hydrological regime (Čurda et al., 2011; Ferda et al., 1971; Janský & Kocum, 2008; Kocum et al., 2016; Vlček, et al., 2016; Vlček et al., 2017). Peat bogs cover approximately 20–35% of the catchment areas in this region (Bufková et al., 2010), but a larger proportion of these catchments is covered by mineral soils such as Podzol. Although peat bogs do not dominate the catchment area, their hydrological regime can determine runoff processes of the whole catchment (Vlček et al., 2012).

Many studies confirmed that in catchments formed by several soil types with their unique hydrological regimes, it is important to understand each system because each one could have its impact on water storage or water chemical properties (Kirchner, 2016; McDonnell et al. 2007; Robinson et al. 2013; Uhlenbrook et al. 2008). Several studies, therefore, focus mainly on waterlogged riparian zones which represent connectivity areas between hillslopes and streams (Seibert et al. 2009; Von Freyberg et al. 2014). On the other hand, riparian zones in the Šumava Mts., Ore Mts. or Jizerské Mts. form rather thin peaty buffers along streams with low water storage and flow delay; thus, we included riparian zones in this study into other morphological formations like hillslopes. Several experimental catchments were also found where peat (Histosol) creates only a part of a whole soil cover (Dick et al. 2018; Lessels et al. 2016; Scheliga et al. 2019; Šanda et al. 2014; Šanda et al. 2018; Tetzlaff et al. 2007). Compare to our catchment, peat covers concave parts of a valley. In this study, ombrogenous raised peat bog formed by Histosol and mineral soil Podzol create two separate hillslopes.

Rainfall-runoff relationships can be predicted by a wide range of hydrological models (Abbott et al. 1986; Arnold et al. 1998; Bergström, 1992; Beven & Kirkby, 1979). However, models tailored to the characterisation of the hydrological regime of peat bogs are scarce (Price et al. 2005). Some attempts have been made by Dunn and Mackay (1996) using the SHETRAN model to investigate the effect of drainage ditches. Lane et al. (2004) utilised a modified TOPMODEL, benefitting from the high-resolution Digital Elevation

Model, for the prediction of saturation of the blanket peat environment. Lane and Milledge (2012) used several TOPMODEL modifications for the estimation of the peat drains influence on runoff generation. Ballard et al. (2012) also used a simplified physically-based model to simulate the runoff reaction and water table fluctuations of drained peatland in the UK. However, the model performance was poorer in dry conditions. Lewis et al. (2013) used a physically-based GEOtop model for the identification of the hydrological response to afforestation in a small Ireland catchment. Nevertheless, the correct quantification of the runoff generation and particular water balance components is still encouraging in the peat bog environment because the models usually do not respect the Acrotelm-Catotelm scheme (Holden & Burt, 2003). The use of a box-model such as HBV (Bergström, 1992) is, therefore, an option because the box models (compared to physically based ones) seem to be favourable for the algorithmic convenience, but as pointed out by McDonnell (2003) they may also represent the way forward to match the appropriate level of understanding and behaviour of the hydrological systems.

The main objective of the present paper was to investigate the contribution of the Peat bog hillslope to the runoff formation in comparison with the mineral soil Podzol hillslope. This was conducted using a newly designed box model representing the hydrological behaviour of the Peat bog hillslope as well as the mineral soil hillslope (represented by an HBV-light model). The hydrological model is based on extensive measurements of the water regime in the peat bog environment during the four-year period (2013–17). Additionally, the Peat bog hillslope water balance was estimated and contrasted with the adjacent hillslope formed by Podzol soil. The interest in hydrological behaviour of the peat bog areas is given by the ongoing effort to recover these environmentally important places.

2 Site description

The experimental catchment is located in the central Šumava Mts. as a part of the Vydra River headwaters area (Fig. 1).

The climate in this area is variable, subject to both oceanic and continental influence. Using the Köppen climate classification, the site lies in the Dfc climate zone (Tolasz, 2007), which is characterised by a subarctic climate with an approximately uniform precipitation distribution. The annual amount of precipitation from the nearby meteorological station at a similar altitude equals 1695 mm yr^{-1} (1981–2010) (Starostová, 2012) and the average daily air temperature is 4.8 °C (ca 15 km).

The Rokytka catchment (RO) occupies 3.8 km^2 and about 30% of the area is covered by peat or other hydromorphic soils (Fig. 1a). For a better understanding of local hydrological processes, a small tributary of Rokytka (RP) has been selected. RP has an area of 0.65 km^2 where more than 60% is covered by peat (Fig. 1a, red border and 1b). The RP catchment was also selected for its special morphology. It is divided into two main hillslopes in east-west directions, each site with different vegetation and soil cover.

The soil cover of the RO catchment (3.8 km^2) is a typical example of Šumava Mts. soils where a vertical sequence of several soil types with Histosols is common. The area is mostly covered by entic Podzol and organic soils, mainly Histosol. In certain concave parts of the catchment, Gleysol can be found (Vlček et al. 2016). The soil cover of the small RP catchment (0.65 km^2) is more homogeneous and differs mainly with two dominant hillslopes. The soil profiles are similar throughout each hillslope without a clear gradient towards the stream. The soil type of western hillslope has been identified as an entic Podzol with a shallow organic top layer (<5 cm) and similar soil texture to a depth of 1 m. Some small parts of the Podzol hillslope are covered by haplic Podzol, but these areas are hardly identifiable without an excavation. Neither, there was a sharp transition between the mineral soil and the bedrock (well-weathered Gneiss or Granite) perceptible with electrical resistivity tomography (ERT) measurements, nor a persistent groundwater level could be detected (Vlček et al. 2017; supplements Fig. S1). The eastern hillslope is created by a well-developed raised ombrogenous peat bog with dominant soil type Histosol with depth varies from 0.5m (lower part) to ca 6m (top part).

Both hillslopes of the small RP catchment can be nicely visible by the vegetation cover while vegetation relates closely to soils (Fig. 2b). The western mineral soil Podzol hillslope is covered by beech stands at the upper hillslope zone; "dead" spruce stands (*Picea abies* L., Karst.) with healthy seedlings cover the lower hillslope zone and the addition of fir (*Abies alba* Mill.) and beech (*Fagus sylvatica* L.). Due to the bark beetle calamity outbreak, most of the spruce stands deceased. The forest is being filled slowly, mainly by spruce seedlings and grasses. The eastern hillslope is created by a well-developed raised ombrogenous peat bog, where three vegetation subsections can be found. The upper subsection of the peat bog is covered mostly by cotton-grass (*Eriophorum L.*) or moss (*Sphagnum L.*) with many small lakes. The middle subsection has the lowest water table fluctuation and the vegetation cover consists of pine (*Pinus mugo*), blueberry and moss. The lowest subsection occupies the bottom of the valley and is covered by waterlogged spruce forest with blueberry and moss. Some of the spruce trees are also affected by bark beetle (Bufková, 2009).

3 Methodology

3.1 Field measurements

The experimental site is equipped with an automatic system for measuring meteorological variables with wireless data transmission. The fundamental measurements necessary for this study composed of air temperature and relative humidity (Fiedler RV12/RK5, CZE), which were measured directly in the RO catchment. The global radiation (Kipp&Zonen CMP3, NL), precipitation and wind speed (Fiedler, CZE) were available from the Modrava meteorological station located 5.7 km from the catchment divide. All the above-mentioned data were measured at 10-minute intervals (Fig. 3) during years 2013–17. The snow pillow located in the RO catchment was used to measure snow water equivalent and snow depth in a one-hour time-step since the winter 2015/2016 (data available for the last two winters (2016-17).

Soil water regime was measured at the Podzol hillslope at two places (upper and lower part) at two depths (20 and 60 cm) by soil tensiometers (T8, UMS company). Their position was chosen according to previous soil survey and the vegetation cover described above. The volumetric soil water content was inferred from the pressure heads by means soil water retention curves that were determined by the pressure apparatus (Soilmoisture, USA) in the laboratory. The parameters of the soil water retention curves are documented in Table 1 and more thoroughly in the previous study (Vlček et al. 2017). Figure 3c shows the average values of volumetric soil water contents at 20 and 60 cm, respectively.

Groundwater level (Figure 3d) was measured by a TSH22 hydrostatic submersible level probe (Fiedler, CZE) and represents the average value of all three measured sites which were placed uniformly at the Peat bog hillslope. Groundwater level fluctuates between the surface (0 cm) and a depth of 40 cm below the surface. However, most of the

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time during the year peat bog is quite saturated by water. The average value of the groundwater level was 18 cm below the surface.

Catchment outlets were equipped with automatic measuring stations (Fiedler AMS company). At both catchments, water levels in 10-min step were measured (at RO outlet by ultrasonic level probe US3200; at RP outlet by pressure sensor with Hydro Logger H40). Water level data was then transferred to discharge by rating curves for each profile.

3.2 Hydrological processes used in the model

The structure of the model is based on an estimation of dominant preferential flow according to Boorman et al. (1995) and Scherrer and Naef (2003). Moreover, it is supplemented and modified according to the authors´ field experiences from the local study sites and other studies dealing with similar research topics. The model consists of two main systems which represent two hillslopes with dominant soil types (Podzol, Histosol respectively) (Table 2). Each hillslope is divided into several sub-basins which are connected by links corresponding to real hydrological processes or dominant preferential flows. In this case, we use the term "sub-basin" as a soil layer (soil part) with similar hydro-pedological processes or water regime. The structure of hydrological processes at each dominant hillslope of the RP catchment described below is visible in the Figure 2a.

3.2.1 Podzol hillslope (PZ)

Upper sub-basin PZ1 represents a soil layer (entic Podzol). The overland flow was not recognised at this hillslope even during heavy rainfall events (>40 mm day^{-1}). Near-surface biomat flow (Gerke et al. 2015) was proved by water sampling from different soil horizons in an excavated profile and by sprinkling experiment (Vlček et al., 2017). From sub-basin PZ1 water flows into two sub-basins (both representing regolith) with different flow delay.

Sub-basin PZ2 represents macropore flow proved at this locality by Vlček et al. (2017). Sub-basin PZ3 shows a deep percolation with a slow flow velocity, which could be estimated by hydraulic conductivity of the lower soil layer (~0.3 mm h^{-1}). The transition between soil and a regolith is gradual since ERT measurements did not find any visible change in electrical resistivity except solitaire rocks.

3.2.2 Peat bog hillslope (PB)

Peat bogs contain in general two main hydropedological layers — Acrotelm and Catotelm. Their hydrological behaviour was nicely described besides others by Holden and Burt (2003) and also at the RO catchment by Vlček et al. (2017). While Acrotelm can form a fast shallow subsurface flow (or biomat flow), it is hard to distinguish between biomat flow and saturated overland flow. Therefore, we joined these processes together in one fast shallow or surface flow. The main hydrological process of Peat bog hillslope PB is situated (by the model) in sub-basin PB1, which represents Acrotelm and a thin upper part (layer) of Catotelm. A shallow subsurface flow (biomat flow) was estimated as a dominant process not only in this locality (Vlček et al., 2017) but also in other peaty areas (Evans et al., 1999; Holden & Burt, 2003). Overland flow occurs when an Acrotelm gets fully saturated with water. The saturated overland flow was visible during heavy rainfall also at the Rokytka Peat bog hillslope.

In general, most of peat bog springs can dry up during summer while they are fed by a shallow layer of soil, especially by an Acrotelm and/or a thin upper layer of Catotelm. In the case of the non-drying spring, Catotelm caused a continuous flow. This hydrological process is enhanced at drained peat bogs where drain bottoms lay below groundwater level. The hydrological behaviour of springs (precisely water level fluctuation) was used in the model as another flow process from sub-basin PB1 (at the range between Acrotelm and

Catotelm; Fig. 2). The lower border of the PB1 sub-basin cannot be unfortunately exactly determined while it depends on a groundwater level, morphology or slope of a peat bog. Except for the two above-mentioned flows from the PB1 sub-basin (overland + biomat flow and flow from springs), the other two possible downward flows could be found. The first, sub-basin (PB2) means a fast flow (a quick response to a rainfall event), called a pipe flow, as discussed by Jones (1997), Holden and Burt (2003) and Uchida, Tromp-Van Meerveld, and McDonnell (2005). Pipes extend horizontally, while they are created mainly by partly decayed wood or by big pores originated from peat drying or erosion. The second, slow flow (PB3 sub-basin) corresponds to a deep percolation with a flow velocity similar to the hydraulic conductivity of the peat in Catotelm (> 1 cm day^{-1} (field in-situ measurements; Holden et al., 2001)). The downwards direction is driven by gravity forces and could be called a piston flow.

Water flows from Peat bog hillslope (PB) and Podzol hillslope (PZ) into the stream creating an outflow from the catchment. Both hillslopes are also affected by evapotranspiration from the upper sub-basins at each site (PZ1 and PB1).

3.3 Model description

The utilised model is based on the HBV scheme (Bergström, 1992), which was (for the peat bog domain) modified in order to be in line with the Acrotelm-Catotelm concept as proposed by Ingram (1978). The modelling domain was split into two different zones (representing two distinct parts of the catchment)—Peat bog hillslope (modified HBV) and Podzol hillslope (standard HBV scheme). Each domain was represented by its own model structure differing mainly in the character of the runoff reaction to rainfall. For the estimation of the total catchment runoff, the outflow of both zones was summed with respect to the areal distribution of both domains (peat covering 60% of the area). Before the HBV model modification, several restricting assumptions were defined:

- the flashiness of the hydrograph—episodic measurements of the spring discharge from the mineral soil and peat bog parts of the catchment. In the case of the peat bog, the difference between the maximum (2.1 L s^{-1} PZ hillslope; 19.1 L s^{-1} PB hillslope) and minimum discharge (0.01 L s^{-1} PZ hillslope; 0.8 L s^{-1} PB hillslope) was 10 times higher than from mineral soil.
- threshold relation between groundwater level in the peat bog and observed discharge
- small contribution of the peat bog water to the total runoff during low flow periods—Kocum et al. (2016) reported an approximately 10% contribution of the peat bog water to the total flow in the area based on isotope analyses.
- evapotranspiration takes places only from the Acrotelm and Catotelm and generally has invariable water content (Holden & Burt, 2003).
- the average specific yield from the peat bog should range between 0.2 and 0.3 [-] respecting the study of Bourgault et al. (2017) who quantified the specific yield of moss.

The mineral soil Podzol domain (PZ hillslope, covering 40% of the catchment) utilised the standard model in the HBV-light scheme (Seibert & Vis, 2012) with a soil box PZ1 and two groundwater storages (upper PZ2 and lower PZ3). The only modification was that the water was allowed to form runoff from the soil box following the relative permeability equation (Brooks & Corey, 1964; eq. 1). Hence, runoff is produced as a sum of outflows from the soil box PZ1 (Q_{PZ1}; eq. 1) and both groundwater storages - PZ2 and PZ3 (Q_{PZ2} and Q_{PZ3}; eq. 2–3). Actual evapotranspiration (AET) is allowed to take place only from the soil box PZ1 and is proportional to the saturation of the soil profile and PET (eq. 4). Rainfall is divided into the part that enters the soil box and the one immediately percolating to the groundwater (GW_{rchrg}) following equation 5. The partition of the percolation between both groundwater boxes (PZ2, PZ3) is based on the constant partitioning coefficient (*GWPART*).

$$Q_{PZ1_t} = K_s \left(\frac{\theta_{t-1}-\theta_r}{\theta_s-\theta_r}\right)^{2+\frac{3}{\lambda}} \tag{1}$$

$$Q_{PZ2_t} = \min(PZ2_t^{\alpha_{PZ}} K_1,\ PZ2_t) \tag{2}$$

$$Q_{PZ3_t} = PZ3_t K_2 \tag{3}$$

$$AET_t = PET_t \cdot min\left(\frac{\theta_{t-1}}{FC_{PZ1}LP_{PZ1}}, 1\right) \qquad \text{if } \Theta < FC_{PZ1} \cdot LP_{PZ1}$$

or
$$AET_t = PET_t \qquad \text{if } \Theta \geq FC_{PZ1} \cdot LP_{PZ1} \tag{4}$$

$$GWrchrg_t = P_t \left(\frac{\theta_{t-1}}{FC_{PZ1}}\right)^{\beta_{PZ/PB}} \tag{5}$$

where K_s is the saturated hydraulic conductivity (cm day^{-1}); Θ, Θ_r and Θ_s are the actual, residual and saturated water contents in the soil box (mm), respectively; λ is the pore size distribution index (-); $\alpha_{PZ/PB}$ is the outflow non-linearity coefficient for PZ or PB domain, respectively (-); $K_{1\text{-}5}$ are storage coefficients (-); *PZ2* and *PZ3* are water contents in upper and lower groundwater storage zones in PZ domain (mm); *AET/PET* stands for actual/potential evapotranspiration (mm day^{-1}); $FC_{PZ1/PB1}$ is the field capacity of the PZ or PB domain, respectively (mm); *LP* is the soil moisture value above which AET reaches PET (mm); $\beta_{PZ/PB}$ is a parameter that determines the relative contribution to runoff from rain or snowmelt in the PZ or PB domain, respectively (-); *P* stands for precipitation (mm); and GW_{rchrg} is the part of precipitation directly percolating to *PZ2* or *PB3*.

The PB domain (60% of the RP catchment) was based on two subsurface storages (representing Acrotelm and Catotelm) in which the rainfall was split into based on equation 5, as in the original HBV model. The upper Acrotelm box (PB1) was subject to evapotranspiration (eq. 6) and generated three forms of the runoff. First, overland flow and biomatflow (Q_{SURQ}) occur when soil is saturated and all water from the precipitation is immediately drained to the stream (eq. 7). Second, water from Acrotelm and an upper part of Catotelm feeds springs ($Q_{SPRINGS}$; eq. 8) during a year (except pipe flow). Above a certain threshold (below the saturation, *UZL*), fast storm flow can occur (Q_{PIPE} in eq. 9). The last runoff mechanism from PB1 sub-basin was a pipe flow (Q_{PIPE}) which can take place in the entire Catotelm zone (PB2). However, the water source for pipe flow is usually water from Acrotelm or rainwater. Therefore, the pipe flow process is connected to Acrotelm in the model.

From the lower Catotelm groundwater box (PB3), the runoff (Q_{PB2} in eq. 10) was produced as in the mineral soil domain and, hence, was represented by the HBV approach.

$$AET_t = PET_t \cdot min\left(\frac{PB1_{t-1}}{FC_{PB}LP_{PB}}, 1\right) \quad \text{if } PB1 < FC_{PB} \cdot LP_{PB}$$

or

$$AET_t = PET_t \quad \text{if } PB1 \geq FC_{PB} \cdot LP_{PB} \tag{6}$$

$$Q_{SURQ_t} = (P_t - GWrchrg_t + PB1_{t-1}) - PB1_{SURQ} \quad \text{if } Q_{SURQ_t} \geq 0 \tag{7}$$

$$Q_{SPRINGS_t} = K_3 \cdot \max(PB1_{t-1} - UZL, 0) \tag{8}$$

$$Q_{PIPE_t} = PB1_t^{\alpha_{PB}} K_4 \tag{9}$$

$$Q_{PB3_t} = PB3_t K_5 \tag{10}$$

where *PB1* (Acrotelm) and *PB3* (Catotelm) are water contents in upper and lower groundwater storage zones in PB domain (mm), and UZL (mm) is the threshold in *PB1* storage when Q_{SPRING} is activated.

Potential evapotranspiration (PET) was estimated for both hillslopes, in the same way, using the combined method of Penman-Monteith (Monteith, 1965) (eq. 11; Fig. 3b). Besides the meteorological variables such as air temperature, wind speed and vapour pressure, the net radiation represents one of the fundamental inputs in this approach. The potential evapotranspiration was calculated as follows:

$$\lambda PET = \frac{\Delta(Rn - G) + \rho \cdot c(e_s - e_a)/r_a}{\Delta + \gamma\left(1 + \frac{r_s}{r_a}\right)} \tag{11}$$

where Δ is the expression of the slope of the saturation vapour pressure versus air temperature curve (kPa °C^{-1}), Rn is the net radiation (MJ m^{-2} d^{-1}), G is the soil heat flux (MJ m^{-2} d^{-1}), γ is the psychrometric constant (kPa °C^{-1}), e_s–e_a describes the vapour pressure deficit (kPa) (e_s is the saturation vapour pressure and e_a is the actual vapour pressure), λ is the latent heat of vaporisation (MJ kg^{-1}), ρ is the water density (kg L^{-1}), r_s/r_a describes the ratio of surface and aerodynamic resistance (s m^{-1}). Soil heat flux was neglected in this study because it works with daily average sums of radiation. Net radiation consists of shortwave and longwave radiation balance. While shortwave radiation is available at the site (Kipp & Zonen CMP3, Netherlands), the FAO56 (Allen et al. 1998) approach was used to calculate longwave radiation. To avoid calculation errors due to using of default sets of coefficients in the FAO56 equation, we used calibrated coefficients that lead to more accurate results for longwave radiation balance and, subsequently, PET values (Kofroňová et al. 2019). The snow accumulation and snowmelt were modelled based on the degree-day method (Gupta, 2001) as the necessary data for the radiation balance approach were not available.

3.4 Model calibration

The model was calibrated in the RP catchment using the data from the entire period of hydrological years 2014–2017, hence the entire analyses were based on the calibration period. This was due to the relatively short length of available data, which (if split into calibration and validation periods) would not allow us to cover both wet and dry years sufficiently. The importance of longer time series for the model calibration was stressed e.g. by Larssen et al. (2007).

The model calibration was conducted in three steps as it was necessary to determine 22 model parameters. In particular steps, the selected parameters were calibrated with respect to observed values of snow water equivalent (STEP 1), soil moisture content/groundwater level height (STEP 2) and discharge (STEP 3), respectively. All the parameters were determined using the genetic algorithm with the RMSE as an objective function.

Before the calibration, the sensitivity analyses (SA) was conducted to investigate the influence of particular coefficients on the model performance. The SA procedure was based on changing one parameter at a time with the remaining ones being fixed (Pianosi et al. 2016). The sensitivity of the output to the changes in the input factors was observed by calculating the rate of change of the objective function (RMSE in our case; supplements Fig. S2). The results indicated that the model is sensitive especially to the changes of snow parameters (especially the temperature thresholds), field capacity and relative contribution of rainfall to runoff of the PB domain (*FCPB*, *βPB*), *K2* parameter governing the contribution of the lower groundwater box to the runoff in both domains and based on the SA the calibration of model coefficients was done using the genetic algorithm. The sensitivity and ranges (recommended in the HBV manual) of particular parameters are shown in the supplementary materials (Fig. S2 and Table S2).

First, four parameters were necessary to estimate snow accumulation and snowmelt. These were *TSNOW* (threshold snowfall temperature), *TMELT* (threshold snowmelt temperature), *CFMAX* (snowmelt rate) and *SFCF* (snowfall correction factor). These parameters were optimised to match the course of snow water equivalent obtained from the snow pillow. Second, the parameters influencing the water regime in the soil (*PZ1*) and Acrotelm (*PB1*) were calibrated with respect to measurements of tensiometers placed in the mineral soil Podzol and groundwater fluctuation in the peat bog. Three parameters were common for both domains but were calibrated separately: *FC* (field capacity), *LP* (soil moisture value above which AET reaches PET), *β* (a parameter that determines the relative contribution to runoff from rain or snowmelt). For the Podzol hillslope, three additional parameters were necessary: *Ks* (saturated hydraulic conductivity), *Θr* (residual soil water content) and *Θs* (saturated soil water). In the second step, the parameters from the first step were kept fixed at the optimised values. The same applies for the final third step with the parameters from the first two. For the estimation of runoff (STEP 3), three coefficients were necessary for the Podzol hillslope: *K1* and *K2* (storage coefficients) and the non-linearity coefficient αPZ. Three storage coefficients ($K2 - K5$) and non-linearity coefficient αPB were required for the outflow of the lower groundwater storage zones (one for PZ and one for PB domain). Additionally, two thresholds (*PB1SURQ* and *UZL*) governing the runoff formation from Acrotelm were needed. All the parameters in STEP 3 were optimised with respect to observed discharge.

4 Results

4.1 Climate and hydrological characteristics of the studied period

The studied period included two years with above-average (2014, 2016) and two years with below-average precipitation (2015, 2017). The precipitation records showed years 2014 and 2016 were wetter than average by 5 % and 13 %, respectively. Contrarily, the years 2015 and 2017 attained only 56 % and 80 % of the long-term precipitation amounts. All years exhibited above average air temperatures being 11–25% above the long-term annual value. All these long-term data originate from the Liz station (operating since 1976 in the distance of 20 km) as there is no climate station operating longer than 30 years insufficient distance.

Storm hydrographs at the catchments RP (0.65 km^2) and RO (3.8 km^2) were highly variable and could be characterised by quick and steep rising and falling limbs. The hydrological response to rainfall events was fast and the recession to antecedent base flow occurred rather quickly (Fig. 3e, f). In the study period, the mean daily runoff equalled to 3.6 mm day^{-1} (1329 mm y^{-1}) at the RO outlet. The RP attained an average runoff of 3.4 mm day^{-1} (1297 mm y^{-1}). Mean annual maximum flow (MHQ) of RO was 7.1 mm day^{-1} and mean annual minimum flow (MNQ) was 1.4 mm day^{-1}. MHQ of RP was 8.3 mm day^{-1} and MNQ equalled to 1.3 mm day^{-1}. The difference in maximum and minimum flow between RO (3.8 km^2) and RP (0.65 km^2) is possibly caused by a difference in peat coverage (RO ~ 30% peat bog coverage; RP ~ 60%) and it confirms the hydrologic behaviour of local streams (Čurda et al. 2011; Ferda et al. 1971; Kocum et al. 2016). The differences in annual discharges between RO and RO are discussed more in chapter 5.3.

4.1 Hydrological simulation

The Nash-Sutcliffe coefficient reached 0.64 in the model calibration and the RMSE was 3.1 mm day^{-1}. The Nash-Sutcliffe ranged between 0.56 and 0.74 in particular years. If the errors originating in the timing of the major snowmelt events were omitted (excluding 10 values out of 1558 simulated days), then the RMSE would decrease approximately to 2.5 mm day^{-1} and Nash-Sutcliffe would rise to the value of 0.70. Hence, the model performance statistics were strongly influenced by the estimation of the snow accumulation and melt during four winter periods (also proved by the SA). The model performance was checked by extending the simulated area to the entire RO catchment (covering 3.8 km^2) having the different percentage of catchment covered by peat bog (30% compared to 60% in the case of RP). The average Nash-Sutcliffe coefficient reached 0.67, which is also a satisfactory value. Hence, the model is able to predict runoff formation from the catchments with different coverage of the peat bog.

Figure 4 shows a comparison of observed and simulated data. Simulated values of discharge from RP and RO correlates with observed data in most measurements; however, there are some values which are over- or underestimated. RMSE equalled to 3.1 mm day^{-1} for RP, 2.2 mm day^{-1} for RO. Soil water content ranging from 200 to 350 mm well matched around the 1:1 line with RMSE equal to 13.5 mm day^{-1} (~1.9 cm^{-3} cm^{-3}). The highest deviation between observed and simulated data was found at high soil saturation (> 0.5). Groundwater level values make irregular clusters sorted around the 1:1 line with RMSE equalling 13.2 mm day^{-1} (~2.1 cm^{-3} cm^{-3}).

The comparison of outflow and groundwater level data from the Peat bog hillslope area usually shows that an increase in outflow appears at times of higher groundwater level (Fig. 5). For a specific threshold, which is formed by an interface between an Acrotelm and a Catotelm, groundwater level stops increasing while outflow (discharge) starts increasing. Some measurements are, however, quite far from the most common shape. The same behaviour was observed in simulated data (Fig. 5 right). Model HBV estimated the threshold between an Acrotelm and a Catotelm to be slightly higher than observed. Moreover, during the full saturation, the model estimated lower discharge events than were observed.

Simulated values correlate sufficiently with observed data of both modelled catchments (RP (Fig. 6, upper) and RO (Fig. 6, lower)). A weak point is represented by a simulation of peaks (high discharges) where some events were underestimated and some overestimated. Snowmelt periods represented another significant source of error in the runoff estimation. Even though Figure 6 contains comparisons at two catchments with different size of peat coverage, simulated values show similar results inaccuracy.

4.2 Contribution of Peat bog hillslope to the streamflow

The Peat bog hillslope generally contributes higher annual runoff volume to the total runoff than Podzol hillslope (by 175 mm, i.e. 17% in the period 2014–2017; fig. 7). However, this higher runoff was restricted to events with higher observed discharge. In these events, the contribution of Peat bog hillslope to the total flow was often over 80% with a maximum of 98%. On the other hand, the ratio of Peat bog hillslope contribution to total outflow was very low during low flow periods when Peat bog hillslope formed up to 15% of the total flow. The results document the propensity of peat bog to generate runoff when higher precipitation falls on well-saturated organic soil. In these conditions, a huge runoff formed which on average exceeded the contribution of runoff from mineral soil. Altogether, higher runoff from Peat bog hillslope was observed in 17% of the simulated days. On the other hand, in periods without significant precipitation, the runoff was dominated by the outflow from the mineral soil; albeit, forming a smaller portion of total flow volume on average. The higher annual runoff volume was generated from the peat bog area (compared to Podzol area) every year except 2015 (Table 3) due to a very low amount of precipitation which did not enable sufficient formation of runoff from the Peat bog hillslope.

4.3 Simulated water balance

The simulated runoff differed from the observed runoff on average by 6.7% in the period of hydrological years 2014–2017. The highest difference was observed in the year 2015 (simulated runoff was lower by 24%), which was the warmest year with the lowest amount of precipitation (Table 3). The simulated and observed runoff coefficients equalled 74.9% and 80.3%, respectively. This indicated the importance of the upper (mountain) regions in terms of runoff formation as the average runoff coefficient equalled 26.8% in the

Czech Republic (Tolasz, 2007). The estimated average annual actual evapotranspiration equalled 466 mm yr^{-1} for the mineral soil (81% of PET) and 337 mm yr^{-1} for the Peat Bog area (59% of PET). On average, the actual evapotranspiration of the whole RP catchment was 388 mm yr^{-1} and its annual course corresponded to the observed precipitation amounts (i.e. was highest in 2016 and lowest in 2015).

5 Discussion

Models can estimate answers to hydrological questions where standard field measurements, among reasons such as short time of measurements or any spatial restrictions, are not sufficient. In the case of the Rokytka catchment (RO), previous measurements could not evaluate sufficiently the shallow and deep subsurface flow contribution to the total runoff at both slopes, nor could they estimate the peat bog hillslope contribution to the outflow during low flow periods. The performance of the hydrological model is biased with several uncertainties. First of all, the amount of available precipitation is not sufficiently measured with standard rain gauges with sufficient time resolution.

First of all, the amount of available precipitation can be underestimated with standard rain gauges. The reason is that the measurements from the rain gauges may contain significant errors (especially in the winter period) (Dingman, 2015). Thus, even the corrected amount of winter precipitation is still uncertain because the estimation of precipitation in mountains is accompanied by several deficiencies (Sevruk, 2005). The major drawback in model performance is given by the snowmelt and the snow accumulation/melt routine based on the degree-day approach. In general, this approach is insufficient in the estimation of the timing of the snowmelt causing significant errors in the model's prediction of spring discharge. However, the degree-day approach is a standard method used in several hydrological models to estimate the evolution of snow cover. Water content in the upper soil layers seemed to be simulated reliably in both domains, especially in the mineral soil. The model's prediction of the water regime in Acrotelm was slightly less efficient than expected which is mainly due to the micro-topography of the peat bog. The soil water regime in Acrotelm was characterised using water level measurements obtained from three piezometers (each of them placed in a different part of the bog). Hence, we used their arithmetical average which might not accurately represent the behaviour of the entire peat bog. The different behaviour of the groundwater level measurements in a single peat bog probe was reported by Kellner and Halldin (2002). Finally, the determination of the runoff pathways was the result of the model design (using the runoff generation mechanisms obtained from the literature) without the use of tracer experiments or isotope analyses; the ratio among them is still uncertain and will be an object of our further research.

One of the major results of this study is represented by the estimation of the peat bog water balance and its contrast with adjacent mineral soil. Because a more accurate estimation of the rate of evapotranspiration was missing at the site, it was based on the observed soil water regime in both modelling domains. However, the results can be masked by the uncertain determination of the fluxes in different flow pathways. Several authors have focused on the estimation of AET in the peat bog environment but as far as the authors know, no study contrasted its value with the nearby mineral soil. The annual evapotranspiration of the Peat bog hillslope at the RP catchment ranged from 284 to 367 mm (18–28% of available precipitation), which is a lower amount than reported in the nearby (albeit located at lower altitude) forest environment (~50%, Šípek & Tesař, 2017). It is also lower than previously reported ratios of AET to precipitation by Lafleur et al. (2005). However, the annual rate and daily averages (1.1 mm day^{-1}) and maxima (~3.5 mm day^{-1}) correspond to the study of Sottocornola and Kiely (2010). The most pronounced reason probably lies in the reduced evapotranspiration of peat bog due to the formation of a crust when dry, likely isolating the peat top from lower wet layers as documented by Sottocornola and Kiely (2010). The formation of the surface crust was observed during field surveys in dry periods. Furthermore, cloud water as a significant amount of precipitation can also reduce the rate of AET (Heal et al. 2004). Eliáš et al. (1995) described the influence of cloud water on the water balance in the nearby lower-lying site to be around 7–10% of the total precipitation. Moreover, the differences may be also caused by the different occurrence of the vegetation and free water surface in the peat bog, which was stressed by Sottocornola and Kiely (2010). It must be noted that the ratio of peat bog and Podzol soil evapotranspiration may be affected by the number of dry or wet years in the study.

5.1 Podzol hillslope hydrological processes

Simulated outflow from Podzol hillslope (1083 mm per year, Fig. 7) is divided by the model in a fast preferential flow and slow percolation. Fast preferential flow includes shallow subsurface flow or biomat flow defined by Gerke et al. (2015).

As the surface flow was not recognised at this site, almost all water from rainfall infiltrates into the soil. However, a sprinkling experiment (Vlček et al., 2017) identified near-surface lateral flow. Fast runoff process at mineral soil hillslope was estimated to be 312 mm of the total (Fig. 7), thus not a dominant hydrological process. The sprinkling experiment showed a biomat flow as a dominant preferential flow in local mineral soil, except deep percolation. However, the experiment covers only a small area (2.25 m^2) and a near sub-surface flow can reduce by variable soil topography such as small depressions. These surface segmentations caused by fallen and uprooted trees could affect outflow during rain events by forming small storage spaces with a low infiltration capacity (field observation).

Slow flow nicely correlated with infiltration tests where the infiltration capacity was estimated in an order of less than mm h^{-1} (Jačka et al. 2014; Vlček et al., 2017). Geophysical surveys (ERT measurements) did not show any visible threshold or barrier to possible vertical flow. Together with surface topographic variability, available observed patterns and measured chosen features (section 3.2) indicated that the dominant hillslope runoff process should be a slow deep percolation into regolith or bedrock. This type of runoff process creates water supply for springs during rainless periods. Our model confirms a dominant slow percolation on the mineral hillslope (771 mm of a total 1083 mm yr^{-1}).

5.2 Peat bog hillslope hydrological processes

Compared to mineral soil, peat bog is known for its quick response of saturated overland flow events (Holden & Burt, 2003). Similar runoff formation is visible at the RO catchments peat bog side during a period of high groundwater level and heavy rainfall. This is also proven by the shape of the groundwater level runoff chart (Fig. 5) and in the study done by Evans et al. (1999).

A quick response of an outflow from a peat bog to a rainfall event could be caused by overland flow near-surface flow in an Acrotelm or pipe flow through a whole peat bog (Holden & Burt, 2002; Jones, 1997; Uchida et al., 2005). These mentioned runoff processes were joined to the fast flow and were estimated as the dominant runoff formation processes with the contribution of 773 mm to the total outflow (Fig. 7).

A slow runoff process at the Peat bog hillslope was formed by a deep percolation to bedrock and a small contribution to the stream from the upper layer (Acrotelm and thin upper part of the Catotelm). The first-mentioned process (percolation) is driven by the hydrological conductivity of peat. The mass of water in a peat bog can also cause a pressure height condition in a lower layer of a peat bog and could cause a piston flow into an underlying substrate. The second mentioned process (small contribution to a stream) is visible at the banks of a stream or banks of drainage channels or at places where sites with higher slope occur. Water is therefore forced to leak out from the peat. Similar processes can be found in drained peatlands (Schot et al. 2004). In some cases, this outflow leads to spring formation. However, this process is strictly dependent on the depth of the groundwater level (Worrall et al. 2007). These two above-mentioned hydrological processes form a minor but significant slow runoff formation at the Peat bog hillslope. Slow processes contribution to the total outflow from the RP catchment was estimated to be 439 mm.

5.3 Hillslopes contribution to the total runoff

A discharge from a peat bog is unstable with a high, quick response to rainfall events and visible spring drying during drought as compared to discharge from mineral soil (Evans et al., 1999; Ferda et al., 1971; Janský & Kocum, 2008). It was also shown in measured data from the RP catchment. In the comparison of measured and modelled annual discharges from this study, Peat bog hillslope at the RP catchment contributed to the outflow with an annual 1212 mm on average; more than outflow from the Podzol hillslope. However, the decision on the question from which soil flows more water is not so clear. It depends mainly on soil saturation (depth of groundwater level) or an amount of rainfall. Evans et al. (1999) showed that discharge in a peatland starts increasing at a specific depth of groundwater level. This hydrological behaviour was also confirmed in the RP catchment (Fig. 5). Some studies such as Blažková et al. (2002) work also with a spatial variability of the saturated area in time. In the case of RP and RO catchments, water table depth has influenced the hydrological regime more while the size of waterlogged areas varies very little.

In general, during wet years when a peat bog is saturated, higher outflow occurs from a peaty area then from mineral soils while dry periods can cause drying of spring from peat (Histosol). It is visible from the comparison of mean annual minimum flow (RP – 1.3 mm day^{-1}; RO – 1.4 mm day^{-1}) and mean annual maximum flow (RP – 16.0 mm day^{-1}; RO – 13.6 mm day^{-1}) from experimental catchments RP (~60 % of peat bog coverage) and RO (~30 % peat bog coverage). Mean annual minimum flow occurs at a catchment with the higher peat bog coverage and conversely with mean annual maximum flow. Moreover, based on the hydrological soil type and its dominant runoff formation process (Boorman et al., 1995; Scherrer & Naef, 2003), peat (Histosols) transfers faster rainwater during rainfall event into a stream compare to mineral soil such as Podzol. Therefore, a dry period causes reduced contribution of water from Peat bog hillslope and vice versa during wet periods. There is also another possible reason for different modelled estimation of annual outflow from hillslopes –the role of the snowpack. The uncertainty of input of snowmelt water into a stream can cause the amounts of water getting in an outflow during the study period. However, the main result of this study remains the ratio of slow and fast flows (hydrological processes).

6 Conclusion

This study estimated the hydrological contribution of two hillslopes with two dominant soil types having different hydrological regimes. Field observations, experiments and monitoring of soil moisture at the Podzol hillslope, groundwater level at the Peat bog hillslope and streamflow were analysed to describe the water regime of each slope.

Moreover, hydrological models (HBV) were modified and used in this research to distinguish fast and slow flow at each slope. The model confirms previous studies that peat bog is characterised by a quick response to a rainfall event; however, it also contributed to the stream via slow flow (36 % of the flow from the Peat bog hillslope). This slow flow was determined as a piston flow through peat. At the mineral soil Podzol hillslope, the dominant subsurface flow was estimated to be deep percolation reaching 71 % of the mineral soil outflow. In general, raised ombrogenous peat bog contributes slightly more than mineral soil Podzol to the outflow while its evapotranspiration is lower (337 mm y^{-1} for the peat bog and 466 mm y^{-1} for the Podzol soil). During years with an excess of precipitation, more water flows from the Peat bog hillslope, but during dry years, more water is drained from the mineral soil Podzol hillslope. This analysis, therefore, improves our knowledge about how soils contribute to a stream during rainfall events, base flow and drought periods.

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8 References

Abbott, M. B., Bathurst J. C., Cunge, J. A., O'Connell, P. E., Rasmussen J., 1986. An introduction to the European Hydrological System — Systeme Hydrologique Europeen, "SHE", 1: History and philosophy of a physically-based, distributed modelling system. Journal of Hydrology, 87 (1–2), 45-59, ISSN 0022-1694, doi: 10.1016/0022-1694(86)90114-9.

Allen, R. G., Pereira, L. S., Raes, D., Smith, M., 1998. Crop evapotranspiration – Guidelines for computing crop water requirements. FAO Irrigation and Drainage Paper, 56, Rome.

Arnold, J. G., Srinivasan, R., Muttiah, R. S., Williams, J. R., 1998. Large-area hydrologic modelling and assessment: part I. Model development. Journal of the American Water Resources Association 34(1), 73–89, doi: 10.1111/j.1752-1688.1998.tb05961.x.

Baird, A. J., 1997. Field estimation of macropore functioning and surface hydraulic conductivity in a fen peat. Hydrological Processes 11, 287–295, doi:10.1002/(SICI)1099-1085(19970315)11:3<287::AID-HYP443>3.0.CO;2-L.

Ballard, C. E., McIntyre, N., Wheater, H.S., 2012. Effects of peatland drainage management on peak flows. Hydrology and Earth System Sciences 16, 2299-2310, doi: 10.5194/hess-16-2299-2012.

Bergström, S., 1992. The HBV Model: Its Structure and Applications. Swedish Meteorological and Hydrological Institute (SMHI), Norrköping.

Beven, K. J., Kirkby, M. J. 1979. A physically based, variable contribution area model of basin hydrology. Hydrological Sciences Journal 24, 43–69, doi: 10.1080/02626667909491834.

Blažková, Š., Beven, K. J., Kulasová, A., 2002. On constraining TOPMODEL hydrograph simulations using partial saturated area information, Hydrological Processes, 16, 441–458, doi: 10.1002/hyp.331.

Bourgault, M. A., Larocque, M., Garneau, M., 2017. Quantification of peatland water storage capacity using the water table fluctuation method. Hydrological Processes 31(5), 1184–1195, doi: 10.1002/hyp.11116.

Boorman, D. B., Hollis, J. M., Lilly, A., 1995. Hydrology of soil types: a hydrologically-based classification of the soils of the United Kingdom. Institute of Hydrology report no. 126.

Bragg, O. M., 2002. Hydrology of peat-forming wetlands in Scotland, Science of The Total Environment 294 (1–3), 111–129, doi:10.1016/S0048-9697(02)00059-1.

Brooks, R.H., Corey, A.T., 1964. Hydraulic properties of porous media. Colorado State University, Fort Collins, 27p.

Bufková, I., 2009: Ochrana rašelinišť na Šumavě: aneb byly Rokytecké slatě první? In Černý, D. Dvořák, L. Weitfällerské slatě: Sborník výzkumu na Šumavě. 2. Vimperk: Správa NP a CHKO Šumava. 103.

Bufková, I., Stibal, F., Mikulašová, E., 2010. Restoration of drained mires in the Šumava National Park, Czech Republic. - In: Eiseltová M. (ed.), Restoration of lakes, streams, floodplains, and bogs in Europe: principles and case studies, Wetlands: Ecology, Conservation and Management, 3, 331–354, doi: 10.1007/978-90-481-9265-6.

Conway, V. M., Millar, A., 1960. The hydrology of some small peat-covered catchments in the northern Pennines, Journal of the Institute of Water Engineers, 14, 415–424.

Čurda, J., Janský, B., Kocum, J., 2011. Vliv fyzickogeografických faktorů na extremitu povodni v povodí Vydry. Geografie, 116 (3), 335–353.

Dick, J., Tetzlaff, D., Soulsby, C., 2018. Role of riparian wetlands and hydrological connectivity in the dynamics of stream thermal regimes. Hydrology Research, 49 (3). 634–647, doi: 10.2166/nh.2017.066.

Dingman, S. L., 2015. Physical Hydrology. Waveland Press, Inc., Long Grove, 643 p.

Dunn, S. M., Mackay, R., 1996. Modelling the hydrological impacts of open ditch drainage. Journal of Hydrology, 179. 37–66, doi: 10.1016/0022-1694(95)02871-4.

Eliáš, V., Tesař, M., Buchtele, J., 1995. Occult precipitation: sampling, chemical analysis and process modelling in the Sumava Mts., (Czech Republic) and in the Taunus Mts. (Germany). Journal of Hydrology, 166. 409–420, doi: 10.1016/0022-1694(94)05096-G.

Evans, M. G., Burt, T. P., Holden, J., Adamson, J. K., 1999. Runoff generation and water table fluctuations in blanket peat: evidence from UK data spanning the dry summer of 1995. Journal of Hydrology, 221 (3-4), 141-160, doi: 10.1016/S0022-1694(99)00085-2.

Ferda, J., Hladný, J., Bubeníčková, L., Pešek, L., 1971. Odtokový režim a chemismus vod v povodí Horní Otavy se zaměřením na výskyt rašelinišť. Sborník prací HMÚ, 17, HMÚ, Praha, 22–126.

Gerke, K. M., Sidle, R. C., Mallants, D., 2015. Preferential flow mechanisms identified from staining experiments in forested hillslopes. Hydrological Processes, 29, 4562–4578, doi:10.1002/hyp.10468.

Gupta, R. S., 2001. Hydrology and hydraulic systems. Waveland Press 2nd edition, Long Grove, 739 p.

Heal, K. V., Stidson, R. T., Dickey, C. A., Cape, J. N., Heal, M. R., 2004. New data for water losses from mature Sitka spruce plantations in temperate upland catchments. Hydrological Sciences Journal, 49 (3), 477–493, doi: 10.1623/hysj.49.3.477.54344.

Holden, J., Burt, T. P., Cox, N. J., 2001. Macroporosity and infiltration in blanket peat: The implicaions of tension disc infiltrometer measurements. Hydrological Processes 15, 289–303, doi: 10.1002/hyp.93.

Holden, J. & Burt, T., 2002. Piping and pipeflow in a deep peat catchment, Catena, 48 (3), 163–199, doi: 10.1016/S0341-8162(01)00189-8.

Holden, J. & Burt, T. P., 2003. Hydrological studies on blanket peat: The significance of the acrotelm-catotelm model. Journal of Ecology, 91(1), 86–102, doi: 10.1046/j.1365-2745.2003.00748.x.

Hümann, M., Schüler, G., Müller, C., Schneider, R., Johst, M., Caspari T., 2011. Identification of runoff processes – The impact of different forest types and soil properties on runoff formation and floods. Journal of Hydrology, 409, 637–649.

Ingram, H. A. P., 1978. Soil layers in mires: function and terminology. Journal of Soil Science, 29, 224–227, doi: 10.1016/j.jhydrol.2011.08.067.

Jačka, L., Pavlásek, J., Kuráž, J., Pech, P., 2014. A comparison of three measuring methods for estimating the saturated hydraulic conductivity in the shallow subsurface layer of mountain podzols. Geoderma, 219–220, 82–88, ISSN 0016-7061, doi: 10.1016/j.geoderma.2013.12.027.

Janský, B., Kocum, J., 2008. Peat Bogs influence on runoff process: Case study of the Vydra and Křemelná River basins in the Šumava Mountains, Southwestern Czechia. Geografie, 113(4), 383–399.

Jones, J. A. A., 1997. Pipeflow contribution areas and runoff response. Hydrological Processes, 11, 35–41, doi: 10.1002/(SICI)1099-1085(199701)11:1<35::AID-HYP401>3.0.CO;2-B.

Kellner, E., Halldin, S., 2002. Water budget and surface-layer water storage in a Sphagnum bog in central Sweden. Hydrological Processes, 16(1), 87–103, doi: 10.1002/hyp.286.

Kirchner, J. W., 2016. Aggregation in environmental systems – Part 1: Seasonal tracer cycles quantify young water fractions, but not mean transit times, in spatially heterogeneous catchments. Hydrology and Earth System Sciences, 20, 279–297, doi: 10.5194/hess-20-279-2016.

Kocum, J., Oulehle, F., Janský, B., Bůzek, F., Hruška, J., Vlček, L., 2016. Geochemical evidence for Peat Bog contribution to streamflow generation process: case study of Vltava River headwaters, Czech Republic. Hydrological Sciences Journal, 61(14), 2579–2589, doi: 10.1080/02626667.2016.1140173.

Kofroňová, J., Tesař, M., Šípek, V., 2019. The influence of observed and modelled net longwave radiation on the rate of estimated potential evapotranspiration. Journal of Hydrology and Hydromechanics, 67(3), 280–288, doi: 10.2478/johh-2019-0011.

Lafleur, P. M., Hember, R. A., Admiral, S. W., Roulet, N. T., 2005. Annual and seasonal variability in evapotranspiration and water table at a shrub-covered bog in southern Ontario, Canada. Hydrological Processes, 31(5), 3533–3550, doi: 10.1002/hyp.5842.

Lane, S. N., Brookes, C. J., Holden, J., Kirkby, M. J., 2004. A network index based version of TOPMODEL for use with high resolution digital topographic data. Hydrological Processes, 18(1), 191–201, doi: 10.1002/hyp.5208.

Lane, S.N., Milledge, D.G., 2012. Impacts of upland open drains upon runoff generation: a numerical assessment of catchment-scale impacts. Hydrological Processes, 27(12), 1701–1726, doi: 10.1002/hyp.5208.

Larssen, T., Høgåsen, T., Cosby, B.J., 2007. Impact of time series data on calibration and predictionuncertainty for a deterministic hydrogeochemical model. Ecological Modelling, 207, 22–33, doi: 10.1016/j.ecolmodel.2007.03.016.

Lessels, J.S., Tetzlaff, D., Birkel, C., Dick, J., Soulsby, C., 2016. Water sources and mixing in riparian wetlands revealed by tracers and geospatial analysis. Water resources research, 52(1), 456–470, doi: 10.1002/2015WR017519.

Lewis, C., Albertson, J., Zi, T., Xu, X., Kiely, G., 2013. How does afforestation affect the hydrology of a blanket peatland? A modelling study. Hydrological Processes, 27(25), 3577–3588, doi: 10.1002/hyp.9486.

McDonald, A., 1973. Some views on the effects of peat drainage. Scottish Forestry, 27, 315–27.

McDonnell, J. J., 2003. Where does water go when it rains? Moving beyond the variable source area concept of rainfall-runoff response. Hydrological Processes, 17(9), 1869–1875, doi: 10.1002/hyp.5132.

McDonnell, J. J., Sivapalan, M. M., Vache, K., Dunn S., Grant, G., Haggerty, R., Hinz, C., Hooper, R., Kirchner, J., Roderick, M.L., Selker, J., Weiler, M. 2007. Moving beyond heterogeneity and process complexity: A new vision for watershed hydrology. Water Resources Research, 43, 6p., doi: 10.1029/2006WR005467.

Monteith, J. L., 1965. Evaporation and the environment. In: The state and movement of water in living organisms, 205–234. Academic Press. for the Society for Experimental Biology, Cambridge.

Pianosi, F., Beven, K., Freer, J., Hall, J.W., Rougier, J., Stephenson, D.B., Wagener, T., 2016. Sensitivity analyses of environmental models: A systematic review with practical workflow. Environmental Modelling & Sofware, 79, 214–232, doi: 10.1016/j.envsoft.2016.02.008.

Price, J.S., Braunfireun, B.A., Waddington, J.M., Devito, K.J., 2005. Advances in Canadian wetland hydrology, 1999–2003. Hydrological Processes, 19(1), 201–214, doi: 10.1002/1099-1085(20000630)14:9<1579::AID-HYP76>3.0.CO;2-#.

Robinson, M., Rodda, J.C., Sutcliffe, J.V., 2013. Long-term environmental monitoring in the UK: Origins and achievements of the Plynlimon catchment study. Transactions of the Institute of British Geographers. 38(3), 451–463, doi: 10.1111/j.1475-5661.2012.00534.x.

Seibert, J., Grabs, T., Köhler, S., Laudon, H., Winterdahl, M., Bishop, K., 2009. Linking soil- and stream-water chemistry based on a riparian flow-concentration integration model, Hydrology and Earth System Sciences, 13, 2287–2297, doi: 10.5194/hess-13-2287-2009.

Seibert, J., Vis, M., 2012. Teaching hydrological modeling with a user-friendly catchment-runoff-model software package. Hydrology and Earth System Sciences, 16, 3315–3325, doi: 10.5194/hess-16-3315-2012, 2012.

Scheliga, B., Tetzlaff, D., Nuetzmann, G., Soulsby, C., 2019. Assessing runoff generation in riparian wetlands: monitoring groundwater-surface water dynamics at the micro-catchment scale. Environmental Monitoring and Assessment, 191 (2), 116, doi: 10.1007/s10661-019-7237-2.

Scherrer, S., Naef, F., 2003. A decision scheme to indicate dominant hydrological flow processes on temperate grassland, Hydrological Processes, 17, 391–401, doi: 10.1002/hyp.1131.

Schot, P.P., Dekker, S.C., Poot, A., 2004. The dynamic form of rainwater lenses in drained fens. Journal of Hydrology, 293 (1–4), 74-84, doi: 10.1016/j.jhydrol.2004.01.009.

Sevruk, B., 2005. Rainfall Measurement: Gauges. In: Anderson, M.G. & McDonnell, J.J. (eds), Encyclopedia of Hydrological Sciences, 529–535. John Wiley & Sons, Chichester.

Sottocornola, M., Kiely, G., 2010. Energy fluxes and evaporation mechanisms in an Atlantic blanket bog in southwestern Ireland, Water Resources Research, 46, W11524, doi: 10.1029/2010WR009078.

Starostová, M., 2012. Měření srážek totalizátory na Šumavě. Meteorologické zprávy, 65, 180–183.

Šanda, M., Vitvar, T., Jankovec, J., 2018. Seasonal Subsurface Water Contributions to Baseflow in the Mountainous Uhlířská Catchment (Czech Republic). Journal of Hydrology and Hydromechanics, 67, 41–48, doi: 10.2478/johh-2018-0018.

Šanda, M., Vitvar, T., Kulasová, A., Jankovec, J., Císlerová, M., 2014. Run-off formation in a humid, temperate headwater catchment using a combined hydrological, hydrochemical and isotopic approach (Jizera Mountains, Czech Republic). Hydrological Processes, 28 (8), 3217-3229, doi: 10.1002/hyp.9847.

Šípek, V., Tesař, M., 2017. Year-round estimation of soil moisture content using temporally variable soil hydraulic parameters. Hydrological Processes, 31 (6), 1438–1452, doi: 10.1002/hyp.11121.

Tetzlaff, D., Soulsby C., Waldron, S., Malcolm, I. A., Bacon, P.J., Dunn, S.M., Lilly A., Youngson A.F., 2007. Conceptualization of runoff processes using a geographical information system and tracers in a nested mesoscale catchment. Hydrological Processes, 21 (10): 1289–1307, doi: 10.1002/hyp.6309

Tolasz, R. (ed.)., 2007. Atlas podnebí Česka. = Climate atlas of Czechia. Tolasz, R et al. Český hydrometeorologický ústav, Univerzita Palackého v Olomouci, Praha-Olomouc. 255, bar. il., mapy, 1 CD-ROM. ISBN:978-80-86690-26-1; 978-80-244-1626-7.

Uhlenbrook, S., Didszun, J., Wenninger, J., 2008. Source areas and mixing of runoff components at the hillslope scale – a multi-technical approach, Hydrological Sciences Journal – Journal des Sciences Hydrologiques, 53, 4, 741–753, doi: 10.1623/hysj.53.4.741.

Uchida, T., Tromp-Van Meerveld, I., McDonnell, J.J., 2005. The role of lateral pipe flow in hillslope runoff response: An intercomparison of non-linear hillslope response, Journal of Hydrology, 311, 117–133, doi: 10.1016/j.jhydrol.2005.01.012.

Vlček, L., Kocum, J., Kučerová, A., Janský, B., Šefrna, L., 2012. Retenční potenciál a hydrologická bilance horského vrchoviště – případová studie Rokytecké slatě, povodí horní Otavy, JZ Česko. Geografie, 117(4), 1–17, doi: 10.37040/geografie2012117040395.

Vlček, L., Kocum, J., Janský, B., Šefrna, L., Blažková, Š., 2016. Influence of peat soils on runoff process: case study of Vydra River headwaters, Czechia. Geografie, 121(2), 235–253, doi: 10.37040/geografie2016121020235.

Vlček, L., Falátková, K., Schneider, P., 2017. Identification of runoff formation with two dyes in a mid-latitude mountain headwater, Hydrology and Earth System Sciences, 21, 3025–3040, doi: 10.5194/hess-21-3025-2017.

Von Freyberg, J., Radny, D., Gall, H. E., Schirmer, M., 2014. Implications of hydrologic connectivity between hillslopes and riparian zones on streamflow composition, Journal of Contaminant Hydrology, 169, 62–74, doi: 10.1016/j.jconhyd.2014.07.005.

Worrall, F., Burt, T. P., Adamson, J. K., 2007. Change in runoff initiation probability over a severe drought in a peat soil – Implications for flowpaths, Journal of Hydrology, 345(1–2), 16-26, doi: 10.1016/j.jhydrol.2007.07.008.

Data Availability Statement: Data will be provided by the author of this article on request.

Table 1. Parameters of retention curves used for Podzol hillslope soil. Locations correspond to Fig. 1. Number in brackets means a depth in the soil [cm].

Location	θs	θr	α [1/cm]
SM1 (20)	0.70	0.31	0.33446
SM1 (60)	0.48	0.16	0.50364
SM2 (20)	0.66	0.34	0.32719
SM2 (60)	0.44	0.17	0.45996

Table 2. Scheme of the model divided into sub-basins (SB) which are connected by an estimated process corresponding to a dominant preferential flow or other real hydrological process. In this case, we use the term "sub-basin" as a soil layer (soil part) with similar hydro-pedological processes or water regime.

<table>
<tr><td colspan="4">Precipitation</td><td></td><td colspan="5">Precipitation</td><td rowspan="8"><<< Flow direction <<<</td></tr>
<tr><td colspan="4">Podzol (mineral soil)</td><td>Hillslope</td><td colspan="5">Peat Bog</td></tr>
<tr><td colspan="4">PZ1</td><td>SB</td><td colspan="5">PB1</td></tr>
<tr><td rowspan="3">Evapotranspiration</td><td rowspan="3">Biomat flow</td><td colspan="2">Deep percolation with dual permeability</td><td>Process</td><td>Pipe flow</td><td>Deep percolation</td><td rowspan="3">Biomat flow / Shallow subsurface flow</td><td rowspan="3">Overland flow</td><td rowspan="3">Evapotranspiration</td></tr>
<tr><td>PZ2</td><td>PZ3</td><td>SB</td><td>PB2</td><td>PB3</td></tr>
<tr><td>Macropore flow</td><td>Deep percolation</td><td>Process</td><td>Pipe flow</td><td>Deep percolation (piston flow)</td></tr>
<tr><td colspan="4">Spring PZ (mineral soil) hillslope</td><td>Spring</td><td colspan="5">Spring PB hillslope</td></tr>
<tr><td colspan="4">Outflow</td><td></td><td colspan="5">Outflow</td></tr>
</table>

Table 3. Simulated water balance (all variables in mm) of the Rokytka tributary RP catchment (2014–2017).

	2014	2015	2016	2017	2014-2017
Precipitation	1617	990	2119	1636	1591
PET	552	601	562	599	579
AET_{PZ}	470	429	493	485	469
AET_{PB}	343	312	365	342	341
Q_{PZ}	1104	896	1388	1099	1122
Q_{PB}	1269	715	1728	1268	1244
AET_{total}	394	359	416	399	392
Q_{total}	1203	787	1592	1200	1196

Figure 1. Experimental catchment Rokytka tributary RP (0.65 km²) as a part of the catchment RO; (a) positions of divides and springs at RP; (b) Red dots are tensiometers at two sites (SM1, SM2) and orange dots are groundwater level probes at three sites (GW1–3).

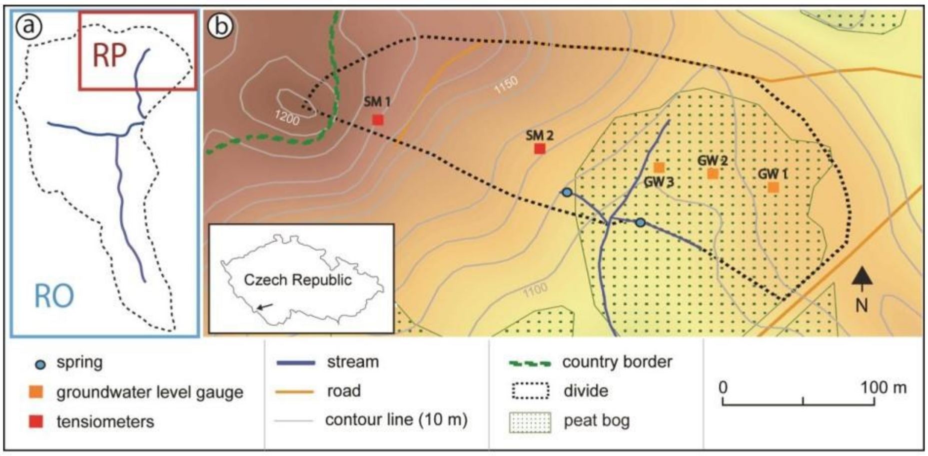


Figure 2. The model (a) and Rokytka tributary RP catchment profile (b) with two hillslopes; a) boxes (PZ/PB) represent each soil sub-basin (Podzol hillslope/Peat bog hillslope) of the model with direction of water flow (blue arrows); b) red dots are tensiometers at two sites (SM1, SM2), orange dots are groundwater level probes at three sites (GW1-3). Arrows show the estimated flow direction based on model (a). In this case, we use the term "sub-basin" as a soil layer (soil part) with similar hydro-pedological processes or regime.

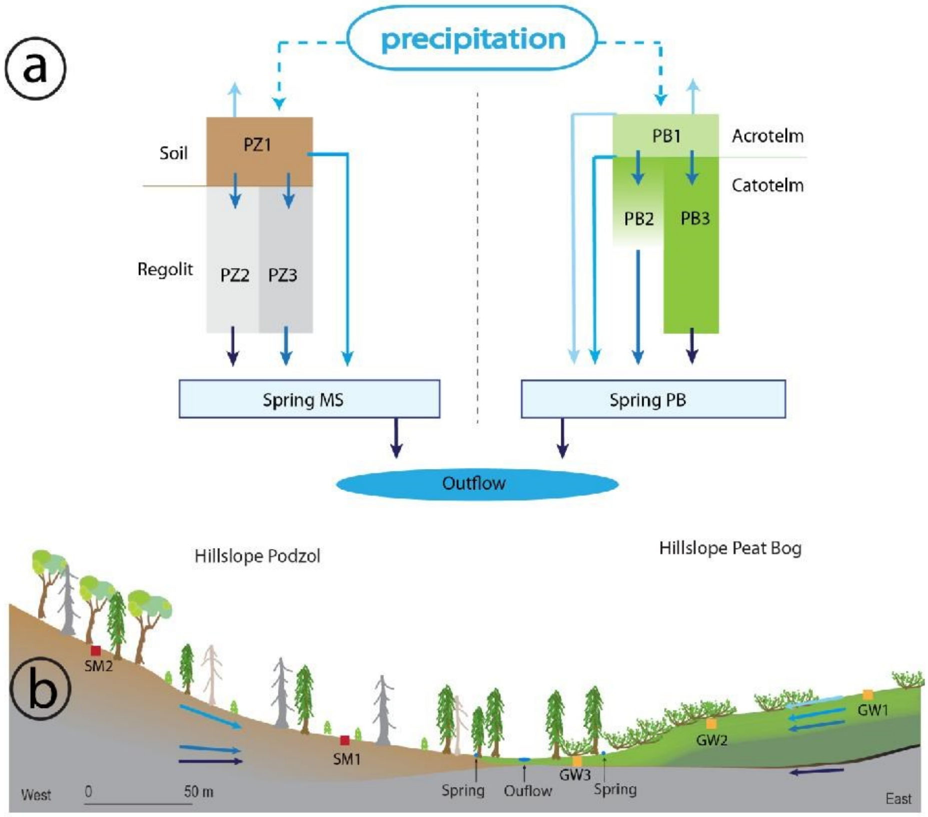


Figure 3. Meteorological and discharge input data from hydrological years 2013–2017 starting from 2012/11/01; a) precipitation (P) modified from measured data (described in chapter Meteo-hydrological input data), b) potential evapotranspiration (PET), c) mean water content of soil and Podzol hillslope from two depths (20, 60 cm), d) mean groundwater level from Peat Bog hillslope, e) specific discharge at the outflow from the Rokytka tributary RP catchment.

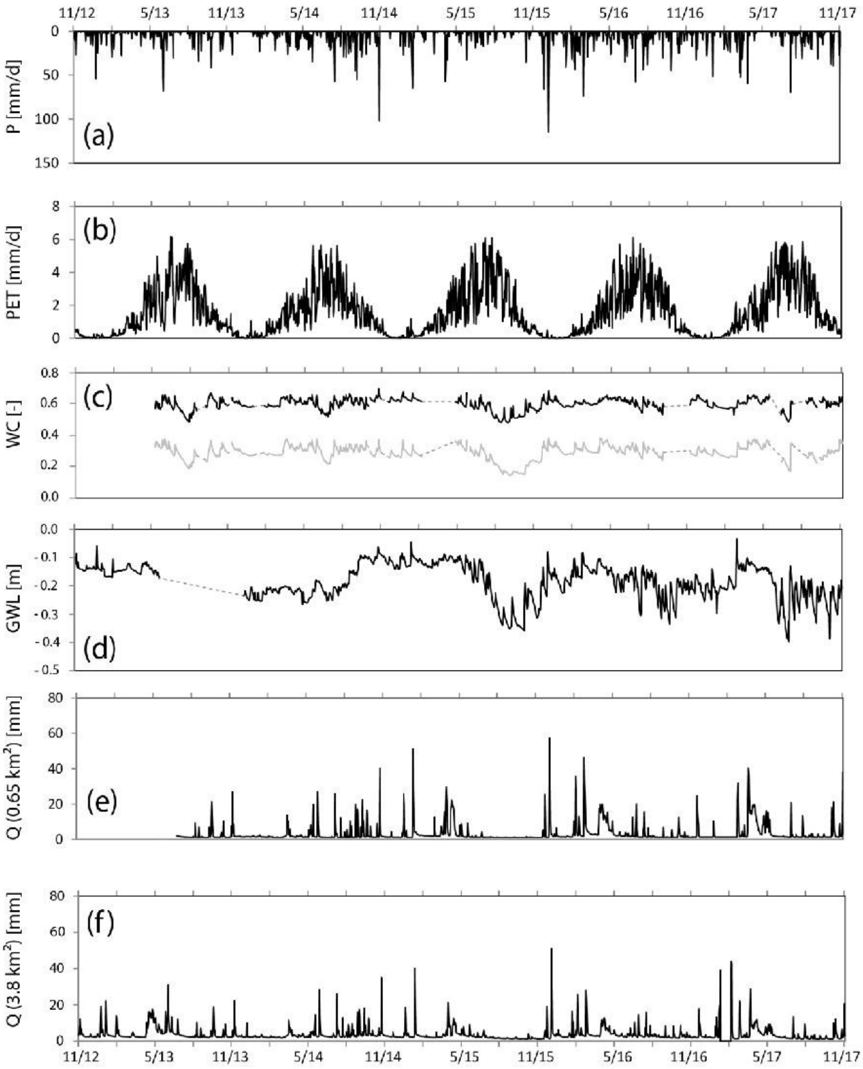


Figure 4. Comparison of observed and simulated values of stream Rokytka tributary RP outflow (upper left), Rokytka stream RO outflow (upper right), soil water (lower left) and groundwater level (lower right) from RP.

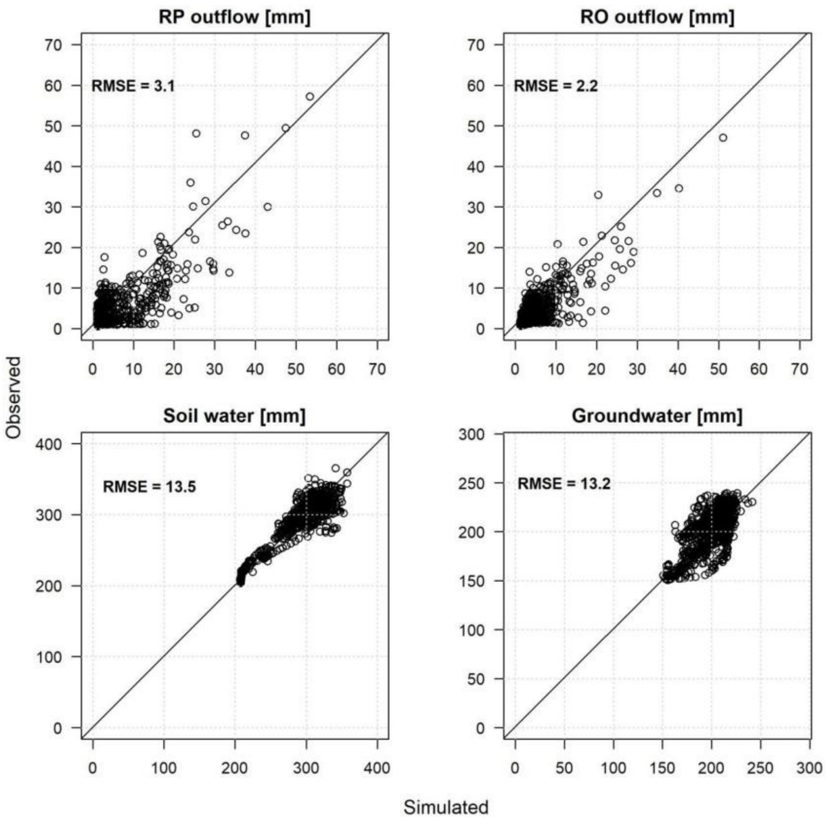


Figure 5. The relation between groundwater level in the Peat bog and outflow from the catchment Rokytka tributary RP using observed (left) and simulated (right) values.

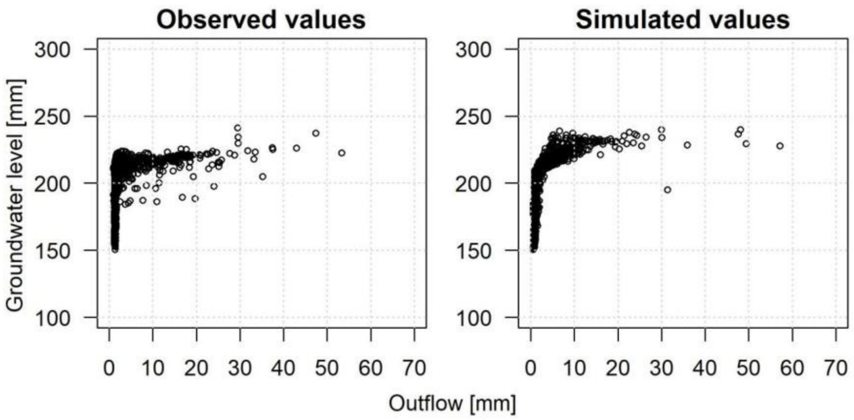


Figure 6. Comparison of observed and simulated values of outflow from catchment Rokytka tributary RP (upper) and Rokytka stream RO (lower) during hydrological years (2013–2017) starting from 2012/11/01. Measurement at RP started at day 239. Data of Rokytka (RO) was adapted to the same period.

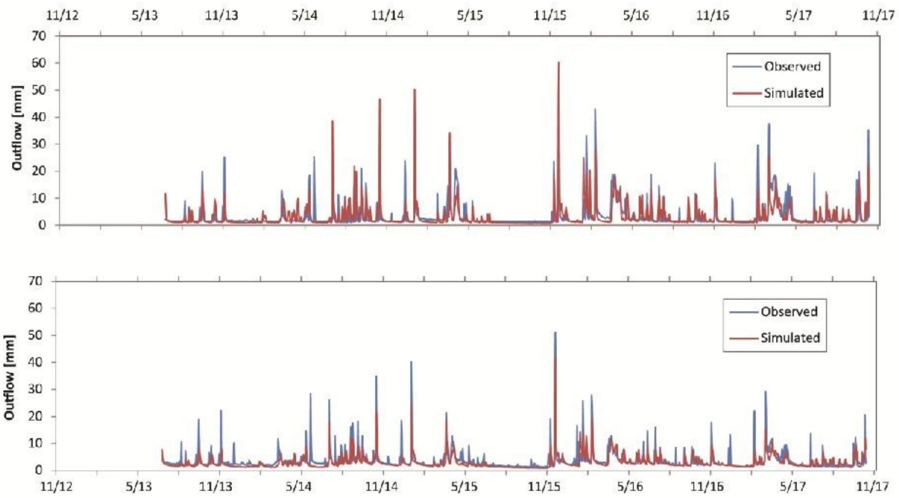
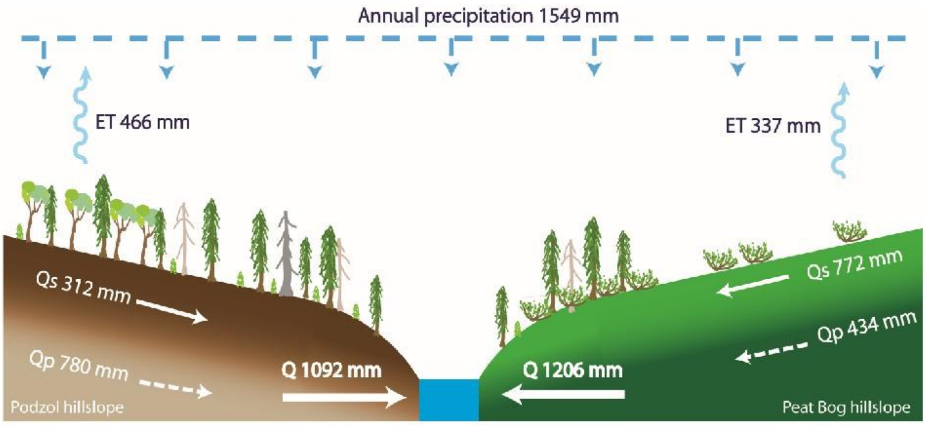


Figure 7. Model of flow distribution in an environmental system with two dominant soils (Podzol and Histosol); ET—evapotranspiration; Qs—stormflow or fast preferential (macropore) flow; Qp—percolation or slow flow; Q—total outflow from each slope.



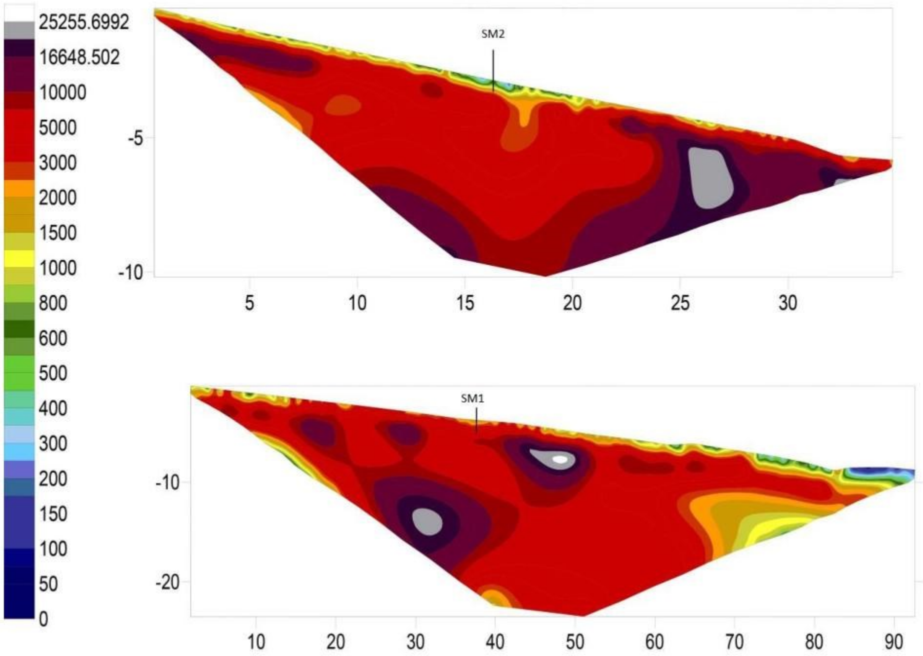


Figure S1. ERT (Electrical resistivity tomography) profile at Podzol hillslope at the location of SM1 and SM2.

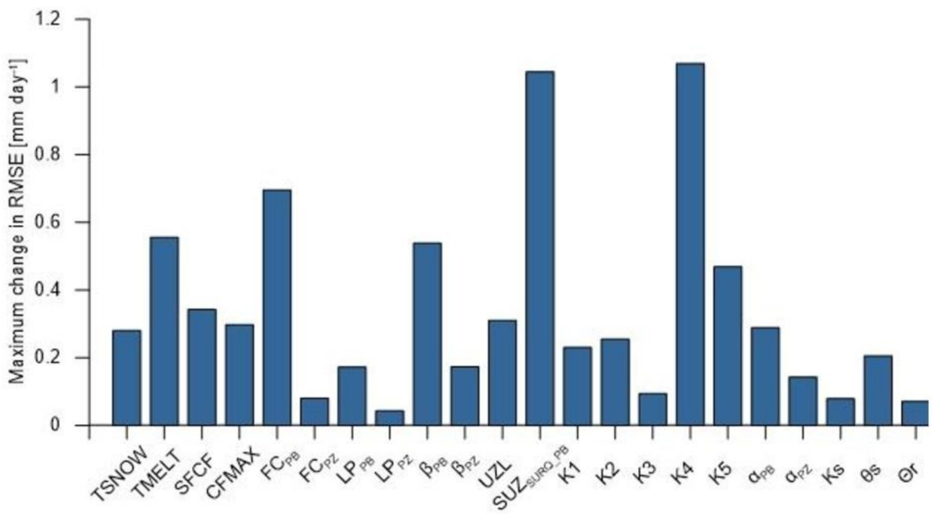


Fig. S2. Absolute maximum changes in RMSE of the discharge estimation at the RP catchment outlet when parameter values change in the entire parameter range.

Table S1. Model parameter ranges and their optimised values. Indexes indicate Peat Bog (PB) and Podzol (PZ) domains. Four snowmelt parameters divided by slashes represent four investigated winter seasons.

Parameter	Unit	Name	Min	Max	Optimised values
TSNOW	[°C]	threshold snowfall *T*	-3	3	2.9 / 0.6 / 1.9 / 0.9
TMELT	[°C]	threshold snowmelt *T*	-3	3	0.5 /-0.5 /-0.3 / 0.0
SFCF	[-]	snowfall correction	1	2	1.0 / 1.7 / 2.0 / 2.5
CFMAX	[mm/Δt°C]	snow melt rate	0	5	2.5 / 2.5 / 3.1 / 2.5
FC_{PB}	[mm]	Field capacity	200	500	308
FC_{PZ}	[mm]	Field capacity	200	500	361
LP_{PB}	[-]	soil moisture above	0	2	1.4
LP_{PZ}	[-]	which AET reaches PET	0	2	1.2
β_{PB}	[-]	relative contribution to	0	Inf	3.4
β_{PZ}	[-]	runoff from rain	0	Inf	4.5
UZL	[mm]	threshold parameter	0	350	221
SUZ_{SURQ_PB}	[mm]	threshold parameter	0	350	245
K1	[-]	storage coefficient	0	1	0.400
K2	[-]	storage coefficient	0	1	0.007
K3	[-]	storage coefficient	0	1	0.606
K4	[-]	storage coefficient	0	1	0.003
K5	[-]	storage coefficient	0	1	0.003
α_{PB}	[-]	non-linearity coefficient	0	1	0.640
α_{PZ}	[-]	non-linearity coefficient	0	1	1.272
K_S	[cm/day]	Saturated hydraulic conductivity	0	500	11.7
θ_s	[mm]	Saturation water content	200	500	341
Θ_r	[mm]	Residual water content	0	300	197

7.2 Hydrologická funkce horských vrchovišť v pramenných oblastech mírného pásu

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Provisional chapter

Hydrological Function of a Midlatitude Headwater Peatland

Jan Kocum, Bohumír Janský, Lukáš Vlček and
Tomáš Doležal

Additional information is available at the end of the chapter

Abstract

Peatland represents quite significant phenomenon in the headstream areas of Czech rivers. Considering the fact that these areas are crucial for streamflow generation process, it is very important to study the mechanism of runoff formation in a peatland and its hydrological function. Natural runoff process is affected by man already by its birth, thus in headwaters where numerous procedures related to runoff retardation and water retention increase in headstream areas could be realized. To understand and clarify the runoff generation process and the effect of various physicogeographic factors on its dynamics, the detailed analyses were carried out in the Vltava River headwaters (sw. Czechia) in recent years. It was necessary to consider the evaluation of peatland retention capacity, its hydraulic communication with draining watercourses and of runoff regime variability during various hydroclimatic conditions. The big attention was focused on findings of a runoff dynamics dependence on the groundwater table in the peatland and of the runoff chemistry and balance using isotopic hydrology methods. Natural tracers were applied at sprinkling plots to identify preferential flow and runoff formation at two opposite hillslopes in this peaty mountain headwater.

Keywords: headwater, peatland, peat bog hydrological function, hydrological extremes, runoff formation, retention potential, Vltava River, Šumava Mts., automatic stations, experimental catchment, oxygen isotopes, tracer experiment, dye

1. Introduction

Mountain peat bogs and peatland represent a significant phenomenon in headwaters of Czech rivers. They occupy a considerable part of the area where the outflow is formed. The study of the hydrological conditions of the most exposed parts of the Czechia therefore requires a very detailed field survey and study of the composition of peatland, its background, development, and hydrological function. These are the areas where the streamflow is generated and then transformed. These headwaters are crucial for the lower parts of river basins from the runoff point of view and in the sense of increasing extremity of climatic and hydrological features. Recently, these effects have been increasingly observed and their effects are mainly attributed to processes related to climate change, also in the mid-altitude part of the European continent.

In the context of catastrophic floods and extreme droughts that have occurred in recent years on the Czech territory, there is an urgent need of solving of issues dealing with protection against hydrological extremes, not using just classical engineering methods. There is a new protection strategy focusing on gradual increase of river catchment retention capacity including its headwater regions where numerous procedures related to runoff retardation could be realized. However, the realization of such measures must be preceded by a thorough research of these areas, not only in terms of hydrological, but also soil or vegetation point of view. It calls for an interdisciplinary concept of research and a comprehensive understanding of the existence of this phenomenon from many perspectives.

Suitable conditions for the research realization at present are related to the mid-latitude Vltava R. headwaters (sw. Czechia) representing the core zone of frequent extreme runoff events with high heterogeneity in terms of physicogeographic and socio-economic aspects. Due to the significant existence of peatland phenomenon in this area, detailed assessment of peat bogs hydrological function, its retention capacity and hydraulic communication have been done in order to evaluate its retention potential. Both classical hydrology approaches and modern methods were used to answer actual questions.

2. State of the art

A number of foreign and domestic projects have solved the matter of peat bog hydrological function but no one has been fully comprehensive. Opinions on their function, already appeared in the second half of the twentieth century, vary a lot. Ferda [1] made the detailed analysis of various approaches to tackle these questions in the Šumava Mts. On the base of "theory of sponge," that occurred in the late 1960s, peatland was distinctive for its significant water retention and discharge regulating ability, and for its discharge heightening ability in dry periods. Other studies from the late 1970s then confirmed the peat bogs retention capacity and show that the only possible way to increase the retention capacity is to lower groundwater level (GWL) by means of drainage. Since that time, the issue of hydraulic communication between peat bog complexes and draining streams (incl. procedures of drainage) has become a field of broad debates among experts (e.g., [2–7]). An interesting and detailed study of the literature covering opinions on both sides can be found in the paper of Holden et al. [8]. Conflicting results presented in the abovementioned papers depend on the different physicogeographical conditions. However, in general, acquired findings proved significant runoff variability of watercourses draining peatland areas. It can be said that the peatland influence on hydrological regime balance had been quite overestimated in the past.

The same result was acquired in the study area of the Vltava R. headwaters in Šumava Mts. [9–13]. Papers show a significantly negative influence of unaffected peatland on a runoff process from its variability point of view. This mountain range has the largest peat bog areas in the Czechia as well as in Central Europe. The existence of large amounts of peat bogs in this area is caused by a humid climate and by optimal relief configuration [14]. The influence of peat land on water quality in watercourses is assessed as unambiguously negative, while intensity of the effect is related to its area and volume in a catchment. Waterlogged areas in Central Europe are formed mostly in flat areas or shallow valleys (e.g., in Biebrza, Poland [15], or in western Slovakia [16]) but climatic and hydrological conditions are different from those of mountainous peat bogs. Quite similar conditions for upland peat bog development can be found in Scandinavia and Scotland. Therefore, it is better to compare hydrological processes within the Šumava Mts. peat bogs to those in Scottish or Scandinavian waterlogged areas.

The influence of peat bogs on hydrological processes has also been discussed with respect to the effect on water quality, especially the ionic structure of water in periods of high or low discharges [17–21]. In dry periods, runoff from peat bogs decreases or becomes almost intermittent. This results in improvement in the quality of the water in the streams draining the peat bog. This was confirmed by studies carried out by Ferda et al. [22] and others [23–25]. However, during spring snowmelt and summer rainfall totals, decline in water quality is observed as peat bog complexes are fully saturated. In case of water release during dry periods, this would be expected to result in decreased quality.

Defining the environment in which hydrological processes take place is quite complicated. Determination of basic hydrological processes using information about the qualitative composition of water is inconvenient and the concept of surface runoff is not sufficient. Hydrogeochemical approaches are suitable to explain the streamflow generation process and to understand the mechanism of water retention in a catchment. Since the theory of so-called "effective precipitation "[26] was accepted, the hydrological response of runoff to causal rainfall has been extensively studied. Despite this, the real mechanism of water behavior underground has not been so clearly described [27]. The absence of such detailed data results in simplified assumptions and insufficient description of complicated processes such as causal aspects of runoff generation. Rainfall-runoff transformation requires additional data that can be obtained using a natural indicator. This information can be provided by a combination of isotope and geochemical approaches [28, 29]. This new dimension to hydrological studies has proven extremely simple and superior to previous theories [27, 30]. Using information about isotopic structure within the soil, subsurface water and causal precipitation amount, proportion of these phases in extreme runoff episode based on isotope concentration in the outflow can be determined. However, mechanism causing this exchange is not completely known [29, 31]. Water can often move apart through isotopically and geochemically specified spaces, channels, or be retained [32]. These spaces are not space-homogenous, and their contribution over time to the proportion of runoff is not necessarily constant [33].

The main anthropogenic changes in the Šumava Mts. peat bog complexes have been caused by efforts of draining and drying. Peat bogs have been traditionally drained for the purpose of peat exploitation, agricultural land cultivation, or increase in wood exploitation in waterlogged forest areas. Nevertheless, the extent of surface drains was already considerable at the turn of the nineteenth and the twentieth century. However, the major period of drainage digging was in the 1970s and 1980s of the twentieth century. Nowadays, the drainage systems are still visible. Stocktaking researches have displayed that drainage has affected almost 70% of peat bogs in the Šumava Mts. [34]. The open system of drains causes especially: fast surface flow, steeper culmination, and higher fluctuations of GWL [35]. Performed restorations can improve these aspects and consequently increase the GWL by several centimeters in a year [36]. A research from Schachtenfilz in the Bavarian Forest has confirmed that restoration measures increased GWL and decreased its fluctuation [34]. Since 1998, a complex restoration program has been implemented in the area of the Šumava National Park. The program is primarily aimed at a general improvement of disturbed water regime in the peat bog area [37]. A concept of so-called "target water level" has been exercised during the restoration in the Šumava Mts. The method is based on determination of necessary water level, which is particular for each peat bog and which is desirable to be achieved by restoration measures. The necessary water level can be described as a maximal tolerated decline of water in a ditch under the dam head, which is bearable for a given type of a peat bog [38].

Peat bogs are physically and ecologically adapted on the depth of GWL. The depth has a great significance for ecological niches of vegetative species and hence even for peat development [39]. The response of GWL on an exercised restoration is usually very fast; nevertheless, the changes in water chemism and consequent reactions of peat bog species are very slow. Peat bog vegetative species are vulnerable and sudden changes of pH factor or changes in the amount of nutrients after exercising restoration can also have negative effects. Peat bog restoration consequently includes stabilization and increase of GWL and a repeated habitation of the standpoint by peat bog species. It is thus important to limit the amount of water drain [40].

3. Study area

The subject area is located within the upper Vltava (Moldau) R. basin, the left tributary of Elbe River, in Central Europe (see **Figure 1**). Headstream part of this basin, where experimental research was undertaken, represents an area with the significant existence of a phenomenon of a peatland that is of mountainous type, mainly fed by atmospheric precipitation. Although the studied area is mountainous, its exposure in the planed and highly exposed part of Šumava Mts. gives it a flat watershed character favorable for the existence of high moor. The catchment is formed by a typical old-aligned surface with an altitude varying between 1.100 and 1.300 m a.s.l. From the geological point of view, according to the tectonic zoning, the basin belongs to the area of Moldau-Danube elevation. Within the various parts of this area, a number of specific experimental catchments were chosen. Their area and slope are similar with the exception of the Rokytka Brook basin, which is slightly flatter. They also have similar soil and vegetative conditions, and most of the area was influenced by a bark beetle infestation. The biggest difference is the extent of peat soils which represents the main reason that why these comparable experimental basins were chosen. All catchments have been monitored several years by installed water level gauges in their closing profiles.

In the Rokytka B. basin, our "field laboratory," the peatland complex comprises several large and many small mountain peat bogs, which are surrounded by forest peat bogs, waterlogged

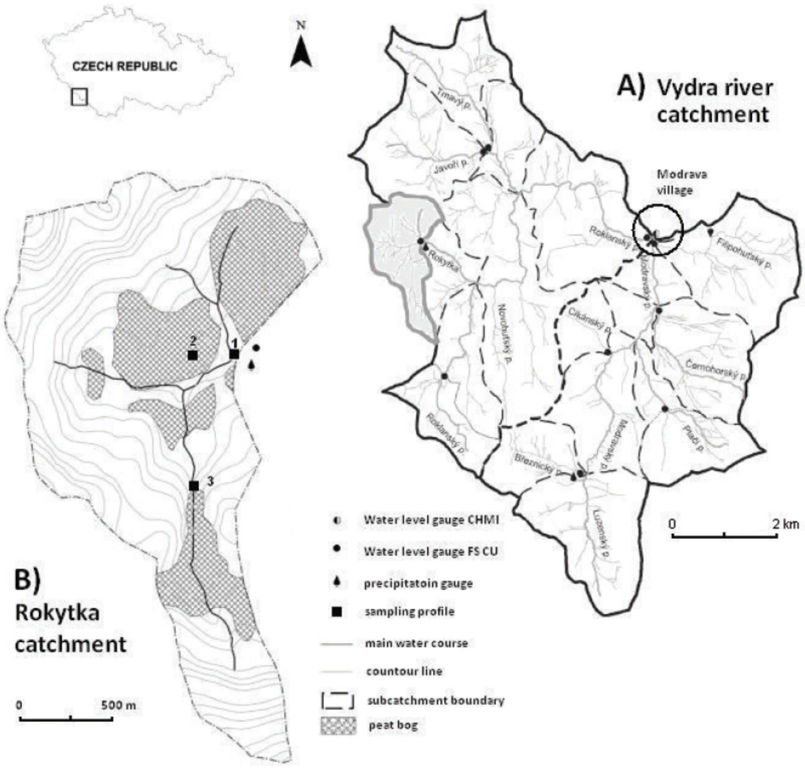

Figure 1. Localization of the study area incl. the CHMI (Czech Hydrometeorological Institute) and FS CU (Faculty of Science, Charles University in Prague) water stage recorders and automatic precipitation gauges within the Vltava R. headwaters. (a) Rokytka B. Experimental catchment within the Vydra River headstream area; (b) sampling profiles and the main peat bog complexes. Sampling profiles: (1) outflow, (2) peat bog lake, and (3) tributary.

pine stands and minerotrophic sedge peat bogs. According to the ZABAGED digital terrain model (the basic platform for geographical data of the Czech Republic) and to the TGM Water Research Institute DIBAVOD (digital basis for water management data), the experimental catchment of Rokytka B., down to the closing profile with installed water level gauge, has an area of 3.86 km^2. The total area of the main studied peat bog within the Rokytka B. catchment is almost 250 ha and its depth reaches up to 7 m. Maximum depth of the peat bog was measured in its central part. It represents historically the deepest analyzed profile in the whole Šumava Mts. with the oldest dating. The research of the Rokytka peat bog was also focused on a selected experimental drainage ditch as the anthropogenic impact, which is located in the northern part of the catchment, at 1.100 m a.s.l. It drains an area of 0.14 km^2. The drainage ditch was partially dammed by small restoration dams; partially it was left functional, with a depth of 1 m.

The bedrock is composed of weathered rocks, mainly granite. Soil conditions in the study area include the features of on-site Organosols, as described by Šefrna [41]. Local soils are typical for the area of Šumava Mts. with characteristic vertical sequence of several types of soil, with Histosols on the ridges and in basins. The largest area of the basin is covered by Entic Podzol, the second most common type of local soil is Histosol (about 26%). Lower part of the basin is filled with a relatively broad peat bog complex with quite significant cubic capacity up to 7.2 m depth. Number of peat bog lakes can be found here as well

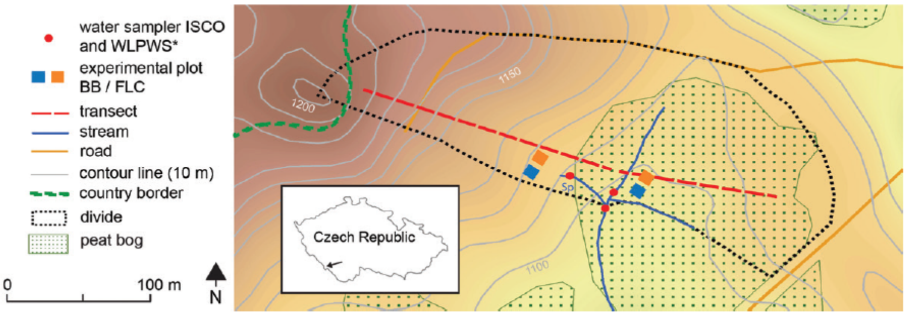

Figure 2. Overview of the Rokytka B. headwater test site (0.6 km²); SpDspring; * water-level proportional water sampler [44].

(see **Figure 1 (b)**). In certain lower parts of the basin, Gleysols are spread out. To consider runoff conditions, water-saturated Organosols can be considered as extreme runoff accelerators. Their retention effect is not approved in the status of full water saturation, even if Organosols have a broad capacity for retention of water. Local vegetation is linked to peat bogs themselves, and forest. Peat bogs are surrounded by waterlogged spruce forest and minerotrophic sedge peat soils [42]. The rest of the forest vegetation is mainly composed of spruce with the addition of fir and beech, and is present predominantly on the south-facing slope. The forest has been influenced by the spruce bark beetle calamity.

To identify the runoff formation in detail using dye tracer experiments, the study site in the northern part of the Rokytka B. catchment was marked out (**Figure 2**). This second-order stream drains the area of 0.6 km² in the altitude between 1.100 and 1.260 m a.s.l. The test site can be divided into two parts represented by two opposite hillslopes with different soil types and vegetation cover. The mineral soil hillslope composed of a Podzol (PZ hillslope) is covered by beech stands at the upper hillslope zone and by dead spruce stands with healthy seedlings at the lower part. The soil profiles do not show a clear gradient toward the stream and are similar throughout the slope. Entic Podzol has been identified, with quite shallow organic top layer (<5 cm) and similar soil texture to a depth of about 1 m. Small parts of the PZ hillslope are covered by Haplic Podzol, but excavation is needed for proper identification. Neither there was a sharp transition between the mineral soil and the bedrock (well-weathered Gneiss or Granite) perceptible with electrical resistance tomography (ERT) measurements nor could a persistent GWL be detected. The organic soil hillslope is covered by a well-developed mountain peat bog (PB hillslope). The entire area consists of a mixture of various stages of decomposed peat. However, Acrotelm and lower Catotelm can be distinguished at depths ranging from 8 to 25 cm [43].

4. Materials and methods

To assess the hydrological balance and runoff formation in a peaty mountain headwater several methodical approaches and various data were used. Automatic stations for the variability monitoring of hydro-meteorological features and physiochemical parameters of surface water were installed in closing profiles of studied experimental catchments. Modern experimental hydrology also uses hydrochemical and geochemical approaches to explain the mechanisms which are related to water retention and runoff formation in headstream areas. Geochemical approach using stable oxygen isotope principle was applied to understand and clarify the streamflow generation processes in the highly peaty catchment. Contribution of water from peat bog areas to the total surface runoff has been assessed for unit hydrogram separation by means of anion deficiency. Tracers such as Brilliant Blue and Fluorescein-Sodium were used and applied at sprinkling plots to identify preferential flow and runoff formation at two opposite hillslopes in this peaty mountain headwater.

4.1. Monitoring of hydroclimatic conditions

The crucial means of obtaining high-quality data for consecutive analyses is represented by the functional system of automatic ultrasound or hydrostatic pressure water-stage recorders, climatic stations and shuttle precipitation gauges (**Figure 1**). Monitoring stations are provided by GSM module that can transmit data through GPRS network. Other modern equipment and methods were used in chosen experimental locations to determine rainfall-runoff relations. A number of experimental profiles also contained sensors for the observation of physiochemical parameters. This network, complemented by the Czech Hydrometeorological Institute (CHMI) state profiles, represents a crucial basis for precise analyses of a local runoff regime. In the profiles given, needful instantaneous discharge measurements using a hydrometric propeller or flow tracker were performed in order to construct accurate consumption curves with high confidence coefficients. Primarily, the influence of peat bog complexes on hydrological conditions was assessed by detailed comparison of runoff regimes in a number of chosen subcatchments with respect to diverse peatland extent and to other relevant physicogeographical parameters. Mechanism of a runoff formation (incl. recent peat bog revitalization processes) was studied primarily using basic hydrological statistics with particular attention to periods of high or low discharge rates. This approach was afterward complemented by much more predicative ion, carbon and oxygen isotope balance analyses (see Chapter 4.5).

4.2. Runoff variability assessment

To assess the runoff variability in chosen profiles, classic hydrological statistics were used at the first step. To assess the degree of extremity in the ascending phase of a flood wave, the method of extremity indices was used [11]. In its first phase, it consists of the determination of the mean discharge of individual streams in the period before the flood wave (D-8 to D-2). The assumption is that this discharge would be reached in the following days if there were no causal situation. For the same period (D-8 to D-2), coefficient of variation (Cv1) from the mean hourly discharges was calculated. The calculated values give us a picture of the degree of fluctuation of individual streams in the period before the flood wave. In the second phase, the variation coefficient for the D-1 to DD period was calculated for each stream, referring to detected theoretical mean discharge of the stream in the period before the causal situation (D-8 to D-2) obtained by the above procedure. D-1 to DD period is the range in which the flood wave increased, culminated and decreased in this case. Calculated values of the coefficient of variation (Cv2) thus represent the rate of flood flow variability from their normal course, which would be theoretically reached without the flood situation. Mutual evaluation therefore provides a good picture of the extent of the flood wave extremity of individual streams in relation to their mean discharge. The use of this method is only applicable to certain flood situations, assuming similar causal conditions for all monitored streams. The following procedure was used to control and eliminate the possible distortion of values of the variability coefficient, depending on the duration of the peak flow and the wavelength on the individual streams. This consists in expressing the value of the mutual share of the maximum reached value of the 10-min discharge in the period D-1 to DD (hereinafter referred to as KP) and the mean discharge in the period before flood wave (hereinafter referred to as PP), in this case D-8 to D-2. The value obtained is referred to as the I_{EKP} peak flow extremity index (I_{EKP} = KP/PP).

4.3. Hydropedological survey

Detailed description of soil profiles and soil sampling for laboratory analyses was carried out. In general, soil retention capacity is measured using a number of methods. One of the most widely used is measurement by the neutron method, the method of retention curves [45], measurement of water isotopes change after passing through the soil [46], and other techniques. Gravimetric method, used in our research, still has many advantages. The most important thing is the simplicity of this method, little time-consuming, and it can be used to evaluate multiple factors at once (soil type, vegetation, etc.). Moreover, in many cases, this method provides results that are more accurate. The retention capacity of the individual parts of the bog was compared with the GWL. Between GWL and surface runoff from the bog, its relation with respect to other factors such as precipitation amount was assessed.

4.4. Groundwater level observation

Groundwater level measurements were implemented during the period from August to October 2014 [47]. This period was crucial for the evolution of GWL within the year. The GWL was measured manually in tubes which were inserted into the peat to a depth of 1–1.5 m. The water level was measured in lines which were copying parts of the drainage ditch. Thus, a regular net with 27 GWL measurement points, placed in regular distances, was created. The GWL was measured from the surface. For this purpose, particular segments were created from the measuring areas, and the GWLs were then compared with each other within the scope of the individual sections and lines (see **Figure 3**). The line 1 was divided into part A and part B for better accuracy. Part A is located directly to restoration dams, and part B is placed in area which is not affected of restoration measures. At each point, 28 values of GWL were measured. Further, particular level changes were statistically evaluated in the scope of individual sections and lines to better demonstrate the dependence of GWL fluctuation on the distance from a drainage ditch, or from restoration dams. Data of GWL from an automatic station in Rokytka peat bog were also used. At first, the whole dataset was analyzed by basic statistical characteristics and data testing. For distribution of measured values of GWL in various intervals, box plots were used. Statistical characteristics variance, correlation coefficient and directive deviance were calculated in software Stat-Soft Statistica. GWL fluctuation was put into context with particular significant factors of rainfall-runoff process, such as potential evapotranspiration. In this research, Penman-Monteith equation was used for the determination of daily potential evapotranspiration [48]. The antecedent precipitation index API [49] was also applied and calculated for five previous days. The index is used for determination of catchment saturation and it expresses the influence of precipitation which occurred in previous days to the given date. It thus demonstrates the ability of a catchment to absorb more precipitation.

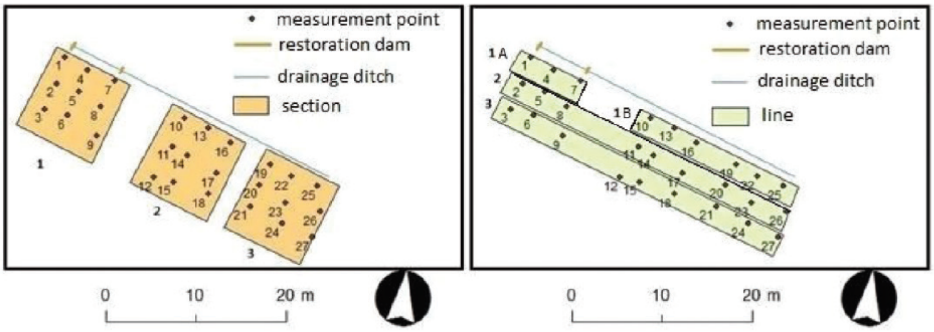

Figure 3. The scheme of particular measurements of GWL and of the segments where the GWL was measured.

4.5. Geochemical analyses

Precipitation and surface water sampling for chemical and isotope analyses was carried out in monthly and two-weekly time steps, respectively, with respect to the whole discharge range, in order to obtain data from extreme episodes such as thaw, snowmelt, rainfall and drought. Precipitation amount and its isotopic composition (δ^{18} O-H_2O) were measured in the adjacent catchments of Roh and Doupě, which have very similar characteristics and are close to the study area. Surface water sampling was carried out in three different sampling profiles: outflow profile (water level gauge), bog profile (organogenous lake) and inflow profile (tributary). The study catchment was closed by the automatic ultrasound hydrological gauge for continual discharge monitoring. The principle of $^{18}O/^{16}O$ fractionation was used for runoff formation modeling. It can be applied due to the uniqueness of the $^{18}O/^{16}O$ isotope ratio of each source—precipitation, subsurface water, surface water—at a particular time. The symbol "delta," used to express the $^{18}O/^{16}O$ isotope ratio, represents the relative proportion of measured $^{18}O/^{16}O$ to a standardized $^{18}O/^{16}O$ proportion (Standard Mean Ocean Water) [28, 30]. Simple model (incl. the inputs from the bog and tributary) was applied to calculate the contribution of the bog to the Rokytka B. outlet. Due to similar signals of δ^{18} O-H_2O in the bog and precipitation total, it was not possible to assess the input of direct precipitation separately. Water balance of the Rokytka B. experimental catchment stems from a mass balance [50]. The contribution of the bog to the Rokytka B. runoff was therefore calculated on the basis of the following equations:

$$Q_O \, \delta^{18} O_O = \circledcirc Q_i \, \delta_i = Q_B \, \delta^{18} O_B + Q_T \, \delta^{18} O_T \tag{1}$$

$$p = Q_B / Q_O \tag{2}$$

$$p = (\delta^{18} O_O - \delta^{18} O_T)/(\delta^{18} O_B - \delta^{18} O_T) \times 100 \tag{3}$$

where δ^{18} O_O is the outflow isotopic composition, δ^{18} O_T is the tributary isotopic composition, δ^{18} O_B is the bog isotopic composition, p is the relative contribution of bog water (%) and Q is the discharge in observed profiles.

4.6. Dye tracer experiments

The dye tracer experiments were carried out at the mineral soil hillslope and organic soil slope of the Rokytka B. headwater during baseflow conditions. At each hillslope, two 1.5 m x 1.5 m plots were sprinkled with both dyes (Brilliant Blue (BB), CAS#3844–45-9, concentration 5 g L^{-1}; Sodium-Fluorescein (FLC), CAS#518–47-8, concentration 2 g L^{-1}). All sprinkling plots were located at the transition between the concave, lower part of the hillslope and riparian zone in the vicinity of the stream [43]. The overall sprinkling time at each plot was ~ 2 h in order to simulate a rainfall intensity of 20 mm h^{-1}. These amounts and intensities represent a heavy rainfall storm in the Šumava Mts. Excavation of the FLC sprinkling plots followed out. After about 4 h sprinkling, exposing of soil profiles and the photography of FLC-stained soil structures were performed under short-time UV illumination (410 nm). As FLC is strongly light sensitive, it was carried out at night [51]. Pictures of the soil profiles were taken during the excavation with a digital Micro Four Third camera with a crop factor of 2.0 under daylight conditions beneath a shading tarp to avoid direct sunlight and shadow effects in case of the BB plots. Pictures at the FLC plot were taken at night with the same camera. Each FLC soil profile was illuminated separately with two light sources (500 W Halogen lamp, 27 W UV LED lamp) to visualize fluorescent FLC-stained soil structures similar to Gerke et al. [52].

The dye-stained flow patterns for both dyes BB and FLC at all soil profiles were analyzed according to a method described by Weiler and Flühler [53]. This method was originally developed for analyzing BB. Therefore, the color space of photographs is converted from the Red-Green-Blue (RGB) color space taken by the camera sensor into the Hue-Saturation-Value (HSV) color space. It was afterward classified and spatially analyzed with an algorithm written in IDL code [54]. This procedure was applied for both dyes (BB and FLC), thus for two different groups of photographs. To detect and analyze FLC in the soil profile photographs similarly to the BB photographs, the dye detection routine in the original IDL code was adapted for optimal FLC identification [43].

5. Results

5.1. Hydroclimatic conditions

In order to assess characteristics of runoff regime and hydroclimatic conditions, hydrological year 2008 was chosen. This year was very average in the sense of hydrometeorological features in recent years. Year 2008 was chosen also because of the fact that cooperation with the Czech Geological Survey (CGS) on geochemical analyses started this year ([55], see Chapter 5.5). The total amount of precipitation in the Rokytka B. catchment in this year was 1485 mm. The seasonal course of $\delta^{18}O$-H_2O in precipitation was very consistent. Rokytka B. represents typical hydrological behavior of streams in the central Šumava Mts., with peak flows occurring in April and May during snowmelt (**Figure 4**). The annual discharge was 0.18 m^3 s^{-1}, so the studied year, 2008, showed an average value. Potential evapotranspiration was calculated using the Penman-Monteith Equation [48] from the set of 2007–2014 data. Evapotranspiration data varied little within the year, with a maximum movement of around 100 mm $month^{-1}$, see **Figure 4**. Observed data were homogenized and deemed representative for consecutive analyses. To evaluate general features of rainfall-runoff regime, mean daily and monthly discharges were calculated.

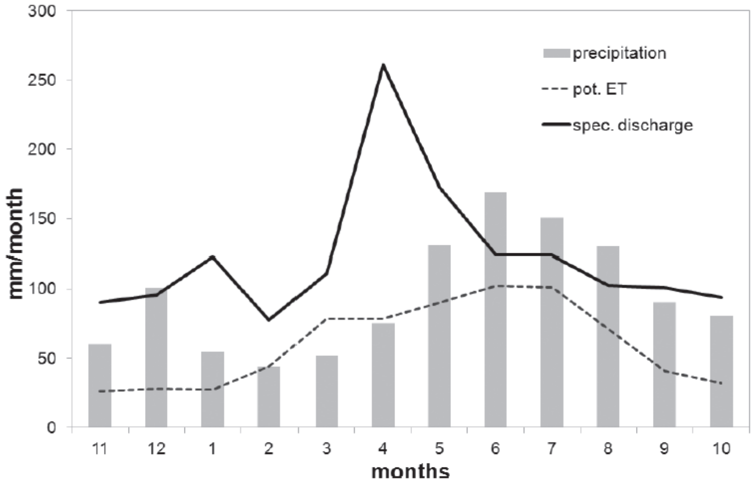

Figure 4. Mean monthly precipitation, specific discharge and potential evapotranspiration (pot. ET) in the study catchment of Rokytka B.

Studied year 2008 was from this point of view determined as an average year (**Figure 4**). The time series show a typical course every year with occasional exception related to thaws. Total runoff (1437 mm) was comparable to the measured amount of precipitation (1485 mm). The precipitation amount did not include water from snow during winter, so it seems quite low compared to the discharge. Higher rate of total precipitation was probably caused also by horizontal precipitation such as fog or frost. In general, the contribution of horizontal deposition in the area of Šumava Mts. is estimated at a minimum of 10%. Most elevated locations, incl. the Rokytka B. catchment, should have a higher horizontal deposition of around 15% [56–58].

5.2. Runoff regime variability

Based on hydrological time series analysis carried out within the upper Vltava R. basin, Kocum [12] determined the significant dependency of runoff variability on a peatland extent in a catchment. Continual records of instantaneous discharge offer an extraordinary database that is unique. Homogenized data can serve as an input for comprehensive analyses of ascending and descending phases of flood waves, and of minimum runoff episodes during dry periods. Detailed statistical analysis of daily, monthly, and yearly time series identified significantly higher runoff variability in the Vydra R. basin. This part of upper Vltava R. basin represents quite peaty area, compared to the nonpeaty Křemelná R. basin. Runoff variability in experimental subcatchments was assessed using the peak flow frequency analysis with respect to the different rates of discharge (**Figure 5**). Analysis of runoff reaction to causal rainfall amount during several rainfall events was also used. These analyses of extreme runoff phases (peak flow frequency method, e.g., [59] or [60]) showed much higher frequency of peak flows and their shorter reaction to causal precipitation total (i.e. lower water retention potency) in the case of highly peaty areas (Rokytka B.). Therefore, it can be said that there is more distinct runoff variability of streams draining peatlands and peat forming soils [61, 12].

Extremity of a hydrologically significant runoff event and specific p-g conditions in individual catchments were subjected to correlation analysis which was based on the method of extremity

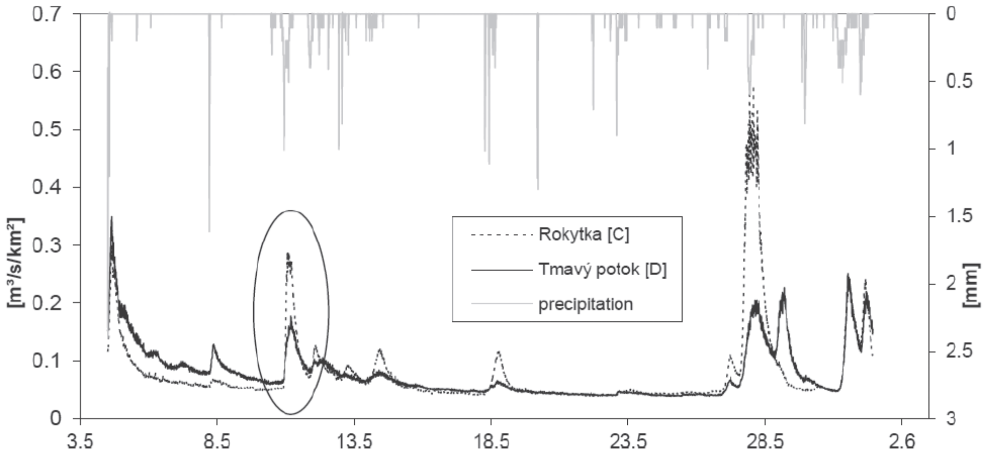

Figure 5. Specific discharge of Rokytka B. (C; 23.1% peat bog extent) and Tmavý B. (D; 2.3%) in May 2013.

indices and on the p-g parameters of the studied catchment. Similar index was used for estimating 100-year flood flows in unobserved catchments [62, 63]. The analysis shows that the extremity of the flood flow is affected mainly by a peat bog extent and by a catchment shape.

5.3. Retention capacity of peatland

Literature suggests that the landscape in the Czech conditions is able to accommodate up to 400 mm of water, an average of 40–90 mm [64, 65]. When considering the average groundwater table (GWL) bogs in the experimental catchment represent areas with the smallest retention capabilities. Retention values are similar to those found in shallow soils (about 140 mm excluding the actual humidity). Considering the lowest GWL bogs represent a significant retention areas within the catchment (230 and 267 mm). Since GWL is higher than its average value for three quarters of a vegetation period, peatland represents within the catchment the area with the smallest retention capacity. However, it is questionable whether the actual moisture measurement was sufficient. In terms of hydrological features, peatland therefore has crucial influence on the retention potential in the landscape [66].

5.4. Evaluation of the influence of peat bog restoration measures on the groundwater level

The variability of GWL represents an important factor of the evaluation the peat bog retention potential. Two different episodes were selected for the evaluation. The first one, the **episode of an intensive precipitation** (55.4 mm), was analyzed between the September 11, 2014 and September 15, 2014 at the Rokytka catchment. It is obvious that GWL along the drainage ditch shows a high amplitude (see **Figure 6**). With longer distance from the drainage ditch, the GWL increases and its change during an episode decreases. The level is the highest in the section close to restoration dams. Their influence is perceived as positive, as they raise GWL. They also have a stabilizing effect. However, the results also imply that in a certain distance from restoration dams, their effects can no longer be seen and GWL fluctuates naturally as in the

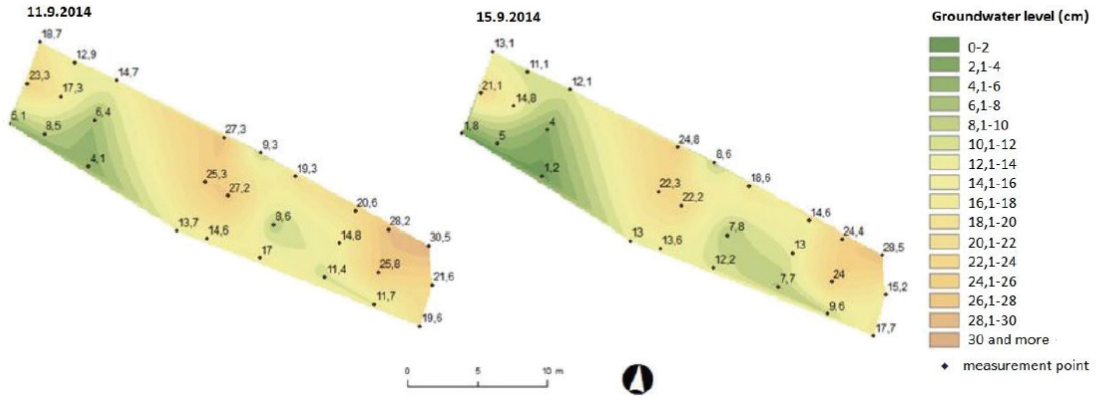

Figure 6. Changes of GWL during a selected episode of intensive precipitation between the September 11, 2014 and September 15, 2014. The given numbers in the graph represent measured GWL in centimeters on a given day.

peat bogs, which are not influenced by a drainage. It is also evident that the decreases or increases of GWL are very variable, and there are noticeable differences between individual points (up to 6.4 cm), in spite of the fact that it is a small homogenous area. On the contrary, in areas near restoration dams, the GWL was increasing very gradually and a similar increase was reached at all the measurement points. Another observed episode was during a **dry period**, when there was only 1.4 mm of precipitation from the September 2, 2014 to September 7, 2014 (see **Figure 7**). The smallest changes of GWL in a period with low precipitation were reached in the middle line of the observed area (3 m from the drainage ditch). It is interesting that in this episode, rather big amplitudes can be found, even in the area of restoration. It can be caused by the fact that before the period of drought, the GWL was very high, precisely right under the surface; hence, following decreases could have progressed faster there. The biggest difference between water levels is significant again and it is even up to 9.2 cm during the monitored 5-day range. It has been confirmed repeatedly that in the areas located further from restoration, the GWL is distinctly lower, and, moreover, there is a remarkable and fast fluctuation of GWL, which is not beneficial for the evolution of mountain peat bogs [47].

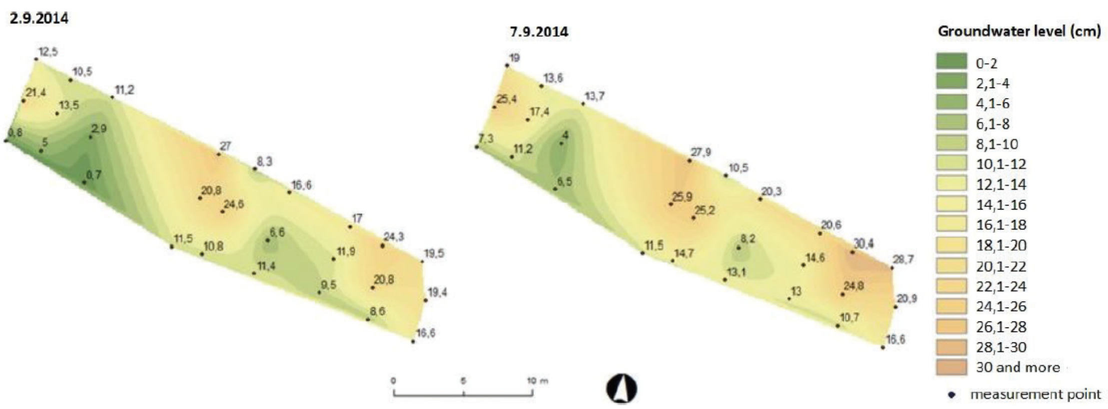

Figure 7. Changes of GWL during a selected episode of drought between the September 2, 2014 and September 7, 2014. Given numbers in the graph represent measured GWLs in centimeters on a given day.

5.5. Runoff chemistry and balance

Peat bog: Water in the Rokytka peat bog had low dissolved solids concentrations. Seasonal profile of δ^{18} O-H_2O (see **Figure 8**) was similar to that for precipitation, as it represents the main source of water in the bog. The hydrogen ion concentration (pH) in bog water is predominantly regulated by total organic carbon (TOC). This concentration shows quite strong seasonal profile related to evaporation and organic matter production (high TOC in summer and low TOC in winter period). Naturally higher content of organic acids along with a low total mineralization results in low pH and low alkalinity of water. Nitrates can be observed in the bog only in winter, while their source is represented by winter precipitation.

Tributary: Study catchment of Rokytka B. is supplied with a number of tributaries. However, two of them are the most significant. Since they show very similar chemistry, due to the fact that both affluents showed very similar chemistry, data from that with higher discharge were analyzed. Total mineralization of Rokytka B. was higher than in the bog. Its δ^{18} O-H_2O profile was more balanced as shown in **Figure 8**. The δ^{18} O-H_2O balance is a result of the prevailing supply of groundwater. Only in periods of higher precipitation, Rokytka B. can contain water from shallow soil horizons with a higher TOC content. Hydrogen ion concentration of Rokytka B. was significantly dependent on discharge and the profile of affluent discharge was very similar to that of brook itself. Increased concentration of TOC was probably related to the production of organic substances during the summer period. There was no significant correlation between TOC and pH.

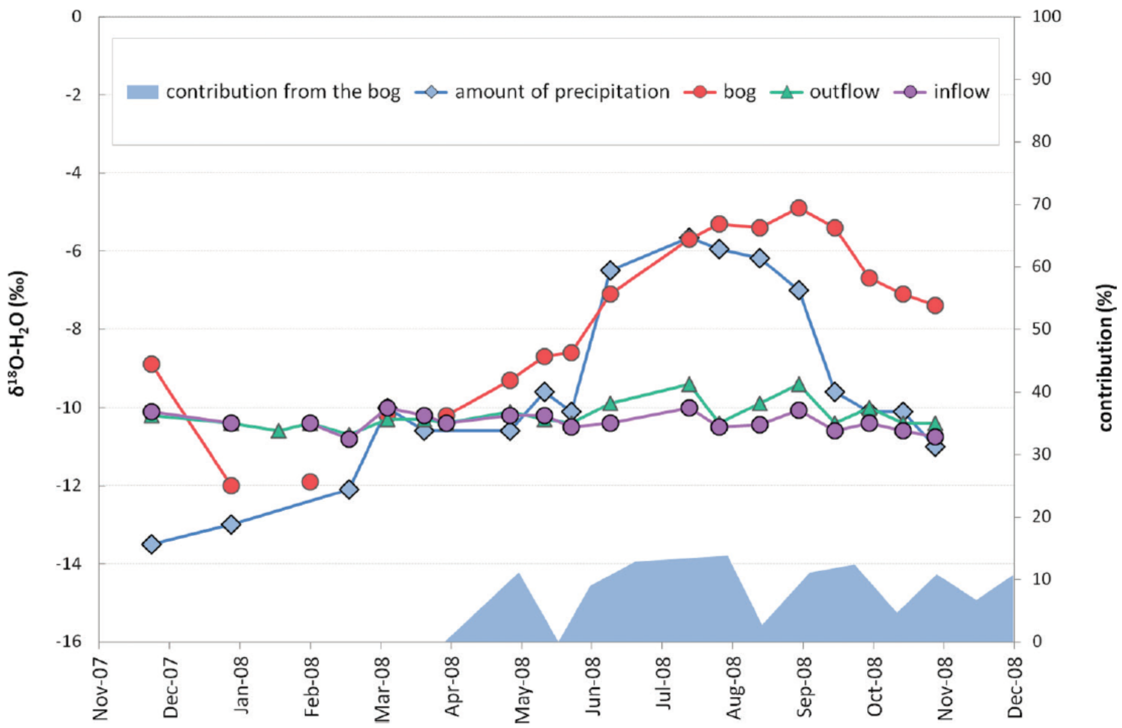

Figure 8. Profile of δ^{18} O-H_2O in surface water and precipitation in the Rokytka B. catchment for the hydrological year 2008; the y-axis shows the relative balance contribution of bog water to the total runoff from the catchment.

Outflow: Chemistry of Rokytka B. in the closing profile looks very similar to the chemistry of the main affluent. On the base of the results (see **Figure 8**), it is clear that the contribution of bog water to the outflow of Rokytka B. was negligible, ranged not more than about 10% of total runoff outside the winter period. During winter, the bog contribution was insignificant and the total runoff was formed only by tributaries, that is, underground water. General character of chemistry of Rokytka B. comes mainly from water sources that have been in contact with mineral soils, even during the period of increased runoff (see stable $\delta^{18}O$-H_2O, **Figure 8**). A strong argument for claiming that the main sources of Rokytka B. runoff are represented by its tributaries, which are mainly supplied by groundwater, is that, compared to the bog, there was also a high concentration of cations in the brook. Regularly increasing TOC concentrations are most likely from the riparian zone, where TOC is washed off during the increased runoff period. Production of seasonal organic matter would also have some influence [55].

5.6. Identification of runoff formation using dye tracers

Near-surface flow in the NW direction toward the stream was revealed by the visual survey of the soil surface in the vicinity of the BB sprinkling plot. Brilliant Blue was detected in a small, water-filled depression 10.5 m downslope from the sprinkling plot. The BB stained flowpaths went from the NW side of the sprinkling plot and followed mostly lateral preferential flow structures formed by decomposed trees or roots. They did not strictly follow the terrain gradient. This lateral preferential flowpath was identified as the main direction of subsurface flow.

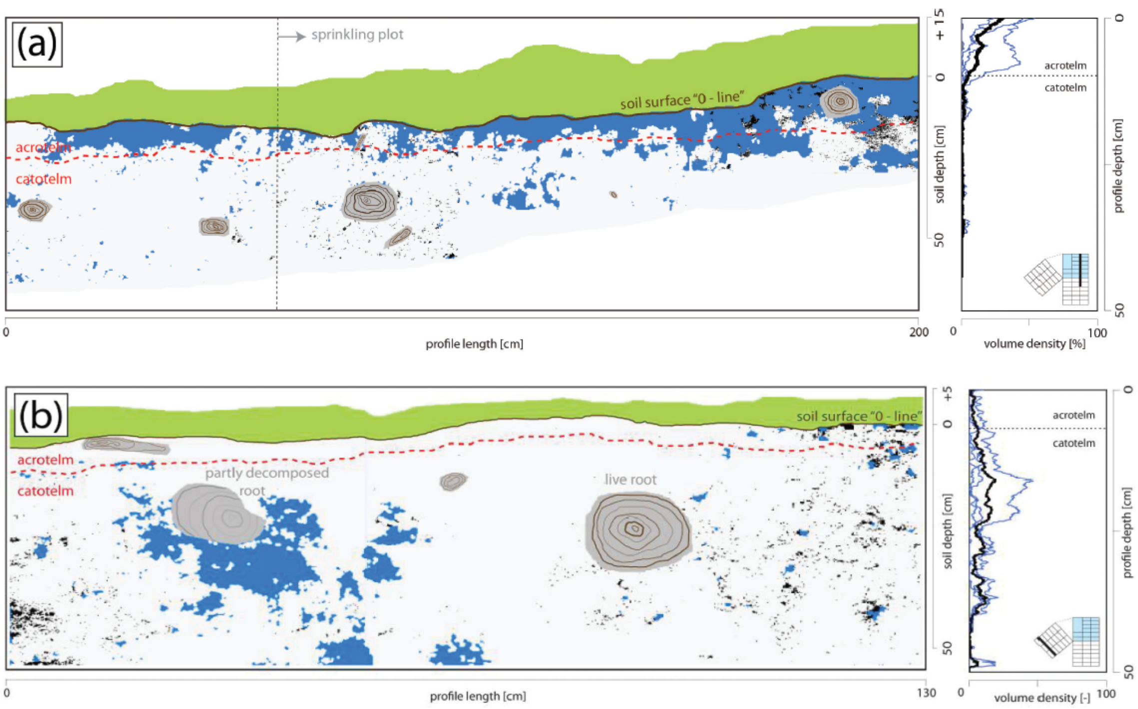

Figure 9. (a) Scheme of lateral soil profile (IL0.5) and (b) frontal soil profile (FD0.25) at the BB sprinkling plot PB3 at the organic soil hillslope (i.e. peat bog). The position of the profile is visualized in bottom right corner. Blue = BB dye, gray = roots, green = vegetation, black = unclassified shadows, red-dotted line = soil horizon dividing line. Charts on the right represent the vertical distribution of the volume density of the BB.

Smaller and less stained flowpaths were detected downslope from the sprinkling plot. BB was disappearing 2 m from the sprinkling plot. BB followed lateral soil pipes that were formed by decomposed roots or fallen trees. Undecomposed timber and healthy trees did not create such effective lateral preferential flowpaths. Accordingly, they had no significant impact on dye-stained patterns (see **Figure 9**). Major flowpaths of BB could be detected even several days after the dye application because BB created clearly detectable dye-stained patterns on the dark peat particles as well. The excavation of BB stained soil patterns at the organic soil hillslope (PB3) proceeded from two directions, NW and SW, following the stained flowpaths in the soil. Near the sprinkling plot, most of the dye was detected at the surface and in near-surface soil horizons, which correlates with Acrotelm (**Figure 9**). About 2.0 m downslope from the BB sprinkling plot at hillslope PB the dye-stained patterns diminished in the Acrotelm and were observed mainly in and around macropores in the Catotelm [43].

6. Discussions

Within the long-term project, *various approaches for the evaluation of hydrological balance* of mid-latitude mountain peatland and peat bogs were used. Classic statistical methods and modern research approaches were implemented in order to understand the real mechanism of the streamflow generation process in areas with significant peat bog phenomenon. The 12-year duration of the project entails the crucial findings that were used in this paper and complement the long-term time series of data from the state profiles. However, different approaches were not used throughout the whole period but in chosen terms. Application of all used methods in the whole period was not possible because of financial and personal resources, as well as the ongoing technology development. However, what was supervised very much in detail was always the choice of correct and relevant data base of needed parameters and suitable time periods. Combination of such corresponding analyses was crucial for complex outcomes that were presented. It has to be stated that every each methodology approach and acquired result casually supports and supplements one another. Such a broad and detailed study has never been carried out in this area and brings completely new findings that are minimally comparable with different types of peat bog complexes.

Thus, general solution of the issue of a *peat bog impact on the runoff process* is not possible. It depends on many factors, mainly on the type of a peat, on its condition and on the extent of anthropogenic influence. Opinions on the peat bogs hydrological function have undergone considerable development and are often contradictory. Generally, the hydrological importance of peat bogs has been overestimated in the past and cannot be regarded as flow regulators because draining streams show extremely high volatility. More controversial discussions within the foreign and domestic literature (e.g., [2, 5, 6]) can be found within the question of drainage of former ameliorative channels or its torrent control respectively. Based on research in the upper Vltava R. basin, it could be stated that it is crucial to take into account the specific characteristics of peat deposits and its surrounding natural conditions while evaluating the revitalization measures effect on runoff dynamics.

Within the literature, a number of *positive and negative examples* of the peat land influence on hydrological regime can be found. These contradictory claims can be paradoxically united. When the bog is drained, runoff variability decreases, but it leads to destruction in time by the bog succession. If GWL would be regulated and reduced in time of need, bog retention potential could be used without the threat of its existence. Periodical fluctuations of GWL in the bog are natural constituents of its development. Minimum time lag between the monitored GWL and surface outflow points to a negligible ability to absorb significant rainfall totals by the bog complex and to a minimum hydraulic communication between the bog complex and its draining stream.

Detection of natural tracers is a useful method to provide the key information in hydrological observation studies of catchment runoff formation. These methods use the different behavior of a small quantity of water molecules. Study of water dynamics by means of natural tracers is typically oriented on usage of oxygen (^{18}O) and hydrogen (^{2}H) isotopes [31]. Stable oxygen and hydrogen isotopes are elements that occur naturally, in variable concentrations, in the hydrological cycle. It provides the unique information about the water that enters a catchment in the form of precipitation, that retains in the catchment and that passes out in the form of runoff. Hypotheses and knowledge of runoff regime dynamics of studied areas gained on the basis of classical hydrological approaches were therefore confirmed by detailed hydrochemical and geochemical analyses. The application of this modern approach in such an optimal model catchment, such as the Rokytka B. catchment, appears as a legitimate shift in research. According to above stated fact, geochemical data show no significant hydraulic connection of the studied bog with the Rokytka B. bed. Moving at a maximum of around 10% out of winter period, as a consequence, the contribution of surface runoff by water from the bog is very insignificant. The predominant portion of underground water (forced out due to the pressure gradient) in total runoff was also confirmed by separation of each runoff component according to geochemical parameters. The problem of hydraulic communication between peat bog complexes and draining streams needs to be solved strictly with respect to local p-g conditions! As it was already said, these findings represent the first knowledge of such a focus in conditions of the Vltava R. headwaters. A similar study describing the use of stable oxygen and hydrogen isotopes was carried out on Uhlířská catchment in the upper part of Černá Nisa River basin in Jizerské Mts. [29, 67]. The prevailing share of subsurface water in the total runoff was confirmed, as in the case of the Rokytka B. study, by the separation of runoff components according to geochemical parameters. During the accelerated runoff, the proportion of water from the causal precipitation episode is gradually increasing, thus contributing to dilution of the draining water. The study of Šanda and Císlerová [67] shows that the drainage of this water is accelerated by the system of partial drainage bases of underground and groundwater in the form of artificial and natural forest gutters, chasms and saturated areas with an ongoing return flow. This course can also be observed in the case of selected catchments in the Šumava Mts. with the existence of nonrevitalized peat bog areas with melioration channels.

If we assess abovementioned outcomes from a hydrological point of view, we have to state following: In physicogeographical conditions of Vltava R. headwaters, peatland acts as a negative element for runoff transformation. Hydrological features of local waterlogged areas are disfavorable. Our primary hydrological assumption of insignificant impact of peatland on runoff dynamics, especially during extreme episodes (floods, droughts), was confirmed by acquired findings from geochemical analyses performed. Considerably weak impact of a peat bog on runoff was also supported by a high concentration of cations in the surface runoff compared to the bog. Much more significant contribution to surface runoff of Rokytka B.

has a groundwater from the basin. In general, very close correlations between pH and actual discharge in experimental profiles were found regularly. A reasonably close relationship was also observed in the closing profile of Rokytka B. catchment. Our research findings strongly support the fact that peatland areas within the studied catchment do not significantly communicate hydraulically with surface streams and their hydrological function is, in the concrete area of Vltava R. headwaters, insignificant [9, 11].

Within the research, the question of impact of ongoing *revitalization measures of the local peat bogs* (made by Šumava National Park management) on the runoff dynamics was opened. Its wholly satisfactory solution, although it should be decisive in the selection of measures to improve the runoff conditions in the area, does not yet exist. Significantly, higher extremity of flood situations was found out in cases of revitalized streams. Local revitalization process consists in damming of former ameliorative channels draining peat bogs. Detailed analyses approved that these revitalization measures stabilized runoff conditions in yearly course and had balancing effect during average runoff situations. In a number of experimental catchments, the presence of revitalization measures can also impact negatively on given flood event. Studies confirmed that revitalization adjustments in selected subcatchments had balancing effect on runoff conditions only to the certain level of its extremity. In most cases, runoff extremity was intensified as soon as the certain water-level stage (respectively discharge) was exceeded. To confirm the correctness of these statements and to correctly understand the functioning of this mechanism, broader data base is needed.

In peaty catchments, the retention ability depends mainly on the shallow depth of the phreatic zone in the peat bog, whereas the deep phreatic zone in the Podzol plays a minor role [13]. Peat bog areas are hypothesized to control storm runoff formation in these headwaters. Peat bogs can significantly contribute to stormflow when the peat is fully saturated, that is, storm events exceeding a threshold of 10–15 mm [68]. As mentioned above, according to a geochemical study based on 2 years of monthly stream water sampling [55], peat bogs contribute only 10% to baseflow at the outlet of the entire Rokytka B. catchment. However, some zones of a peat bog area, such as springs or soil pipe systems connected to the stream, exhibit high fluctuations in discharge [69]. This fact could explain the observed spiky storm hydrographs at the entire Rokytka B. catchment outlet (area of 3.8 km^2) and at the Rokytka headwater test site (0.6 km^2). Presented runoff fluctuations from peaty areas could be caused by surface flow (as observed within a field survey at the Rokytka peat bog), near-surface flow [7, 40] or subsurface stormflow in soil pipes [70, 40, 71]. Outcomes of Holden and Burt [72] at a blanket Peat site showed that near-surface flow (i.e., Biomat flow, BMF) up to the depth of about 10 cm can contribute more than 90% to the plot's outflow. Biomat flow can be defined as a lateral stormflow in the organic litter layer which has quite high porosity and high hydraulic conductivity in the topsoil [71]. Storm hydrographs at the Rokytka B. headwater are highly volatile and are characterized by quick and steep rising and falling limbs. The hydrologic response to rainfall events is fast and the recession to antecedent baseflow occurs rather quickly [43].

7. Conclusions

Based on acquired outcomes from time series statistical analyses, much more distinct runoff variability of streams draining highly peaty catchments in the Vltava R. headwaters (sw. Czechia), especially during extreme hydrological situations, was observed. This fact was confirmed by hydropedological, hydrochemical and geochemical approaches. Geochemical data show no significant hydraulic connection of the studied bog with its draining stream. The predominant portion of underground water in total runoff was also confirmed by separation of each runoff component according to geochemical parameters. However, this subject needs to be solved strictly with respect to local physicogeographic conditions. These conclusions correspond to the typical mid-latitude peat bog area in conditions of Czech mountainous areas. Their restoration measures carried out in recent years have a positive effect on GWL. It was proven that restoration decreases fluctuation and increases GWL, which is essential for a natural evolution of a mountain peat bog. Tracer experiments detected biomat flow, shallow lateral subsurface flow and mostly deep percolation at the Podzol hillslope. At the organic peat bog biomat flow at short distances and mostly lateral pipe flow following decayed tree-root systems with long lateral subsurface flow distances were recognized. It can be stated that bogs in the studied basin represent separate hydrological units with their own typical runoff regime, which does not contribute to the discharge curve balancing (during both floods and droughts), and that their hydrological function in this mountainous area is insignificant.

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References

[1] Ferda J. Hydrologický význam horských vrchovištních rašelinišť. Sborník ČSAZV – Lesnictví. 1960;**10**:835-856

[2] Conway VM, Millar A. The hydrology of some small peat-covered catchments in the Northern Pennines. Journal of the Institute of Water Engineers. 1960;**14**:415-424

[3] Burke W. Principles of drainage with special reference to peat. Irish Forestry. 1967;**24**:1-7

[4] McDonald A. Some views on the effects of peat drainage. Scottish Forestry. 1973;**27**:315-327

[5] Moklyak VI, Kubyshkin GP, Karkutsiev GN. The effect of drainage works on streamflow. In: Hydrology of marsh-ridden areas, Proceedings of the Minsk Symposium, June 1972. IAHS Studies and Reports in Hydrology. Vol. 19. Paris: Unesco Press; 1975. pp. 439-446

[6] Baird AJ. Field estimation of macropore functioning and surface hydraulic conductivity in a fen peat. Hydrological Processes. 1997;**11**:287-295

[7] Holden J, Burt TP, Cox NJ. Macroporosity and infiltration in blanket peat: The implications of tension disc infiltrometer measurements. Hydrological Processes. 2001;**15**:289-303

[8] Holden J, Chapman PJ, Labadz JC. Artificial drainage of peatlands: Hydrological and hydrochemical process and wetland restoration. Progress in Physical Geography. 2004;**28**(1): 95-123

[9] Janský B, Kocum J. Peat Bogs Influence on Runoff Process: Case Study of the Vydra and Křemelná R. Basins in the Šumava Mountains, Southwestern Czechia. Geografie - Sborník ČGS. 2008;**113**(4):383-399. ISSN 1212-0014

[10] Kocum J, Janský B, Čurda J, Česák J. Extremita odtoku v pramenné oblasti Otavy (řešení negativních dopadů klimatické změny). In: Vrabec M, Durčanský I, Hladný J, editors. Hydrologické dny 2010 - Voda v měnícím se prostředí. Sborník příspěvků ze 7. národní konference českých a slovenských hydrologů a vodohospodářů. Praha: Nakladatelství ČHMÚ; 2010. pp. 597-604. ISBN 978-80-86690-84-1

[11] Čurda J, Janský B, Kocum J. Vliv fyzickogeografických faktorů na extremitu povodní v povodí Vydry. Geografie – Sborník ČGS. 2011;**116**(3):335-353. ISSN 1212-0014

[12] Kocum J. Tvorba odtoku a jeho dynamika v pramenné oblasti Otavy. Praha. Dizertační práce: Přírodovědecká fakulta Univerzita Karlova v Praze; 2012. p. 206

[13] Vlček L, Kocum J, Janský B, Šefrna L, Kučerová A. Retenční potenciál a hydrologická bilance horského vrchoviště: případová studie Rokytecké slatě, povodí horní Otavy, JZ. Geografie. 2012;**4**:371-395

[14] Pošta P. Organogenní jezera v České republice. Magisterská práce. PřF UK, Praha; 2004. p. 119

[15] Wassen MJ. Hydrology, water chemistry and nutrient accumulation in the Biebrza fens and floodplains (Poland). Wetlands Ecology and Management. 1995;**2**:125-137

[16] Juráková M, Klementová E. Status and trends of the wetlands in West Slovakia. Slovak Journal of Civil Engineering. 2003;**3**:1-8

[17] Novák M. Huminové vody ve vodách údolních nádrží. Vodní hospodářství. 1955;**5**(4): 127-128

[18] Onderíková V, Štěrbová A. Prispevok k biologii a chemizmu Oravskej nádrže. Vodní hospodářství. 1956;**6**(2):46-51

[19] Novák M. Výzkum kvality vody v údolní nádrži Lipno. Vodní hospodářství. 1959;**9**(9): 378-383

[20] Freeman C, Evans CD, Montieth DT, Reynolds B, Fenner N. Export of organic carbon from peat soils. Nature. 2001;**412**:785-786

[21] Worrall F, Burt TP, Adamson JK. Controls on the chemistry of runoff from an upland peat catchment. Hydrological Processes. 2003;**17**:2063-2083

[22] Ferda J, Hladný J, Bubeníčková L, Pešek J. Odtokový režim a chemismus vod v povodí Horní Otavy se zaměřením na výskyt rašelinišť. Sborník prací HMÚ. 1971;**17**:22-126

[23] Hruška J et al. Role of organic solutes in the chemistry of acid-impacted bog waters of the western Czech Republic. Water Resources Reasearch. 1996;**32**(9):2841-2851

[24] Hruška J, Kohler S, Bishop K. Buffering processes in a boreal dissolved organic carbon - rich stream during experimental acidification. Environmental Pollution. 1999;**106** (1):55-65

[25] Oulehle F, Janský B. Limnologie a hydrochemismus v NPR Rejvíz. In: Jezera České republiky (Současný stav geografického výzkumu). Prague: Charles University in Prague; 2003. pp. 93-108

[26] Horton RE. The role of infiltration in the hydrologic cycle. Transactions - American Geophysical Union. 1933;**14**:446-460

[27] McDonnell JJ. Where does water go when it rains? Moving beyond the variable source area concept of rainfall-runoff response. Hydrological Processes. 2003;**17**:1869-1875

[28] Hoeg S, Uhlen BS, Leibundgut C. Hydrograph separation in a mountainous catchment - combining hydrochemical and isotopic tracers. Hydrological Processes. 2000;**14**: 1199-1216

[29] Šanda M. In: Vrabec M, Durčanský I, Hladný J, editors. Přirozené stabilní izotopy kyslíku a vodíku v hydrologii experimentálního povodí, Hydrologické dny 2010 - Voda v měnícím se prostředí. Sborník příspěvků ze 7. národní konference českých a slovenských hydrologů a vodohospodářů. Praha: Nakladatelství ČHMÚ; 2010. pp. 255-261. ISBN 978-80-86690-84-1

[30] Klaus J, McDonnell JJ. Hydrograph separation using stable isotopes: Review and evaluation. Journal of Hydrology. 2013;**505**:47-64

[31] Kendall C, McDonnell JJ, editors. Isotope tracers in catchment hydrology. Amsterdam: Elsevier Science B.V.; 1999. ISBN 13: 978-0-444-50155-4

[32] Kirchner JW. A double paradox in catchment hydrology and geochemistry. Hydrological Processes. 2003;**17**:871-874

[33] Šanda M, Kulasová A, Císlerová M. Hydrological processes in the subsurface investigated by water isotopes and silica. Soil and Water Research. 2009;**4**:83-92

[34] Bufková I. Naprava narušeneho vodniho režimu rašelinišť v národnim parku Šumava. Ochrana přirody. 2013;**2**:17-19

[35] Ballard CE, McIntyre N, Wheater HS, Holden J, Wallage Z. Hydrological modelling of drained blanket peatland. Journal of Hydrology. 2011;**407**:81-93. DOI: 10.1016/j.jhydrol.2011.07.005

[36] Worrall F, Armstrong A, Holden J. Short-term impact of peat drain-blocking on water colour, dissolved organic carbon concentration, and water table depth. Journal of Hydrology. 2007;**337**:315-325. DOI: https://doi.org/10.1016/j.jhydrol.2007.01.046

[37] Bufková I, Stíbal F, Mikulášková E. Restoration of drained mires (Šumava National Park, Czech republic). Proceedings 7th European conference on ecological restoration Avignon, France; 2010, pp. 23-27. DOI: 10.1007/978-90-481-9265-6_16

[38] Bufková I. Revitalizace šumavskych rašelinišť. Zprávy České Botanické Společnosti, Praha. 2006;**41**(21):181-191

[39] Kvarner J, Snilsberg P. Groundwater hydrology of boreal peatlands above a bedrock tunnel–drainage impacts and surface water groundwater interactions. Journal of Hydrology. 2011;**403**:278-291. DOI: 10.1016/j.jhydrol.2011.04.006

[40] Holden J. Peatland hydrology and carbon release: Why small-scale process matters. Philosophical Transaction, The Royal Society. 2005;**363**:2891-2913. DOI: https://doi.org/10.1098/rsta.2005.1671

[41] Šefrna L. Pedologická charakteristika povodí Otavy ve vztahu k povodním [online]. Charles University in Prague; 2004. Available from: http://hydro.natur.cuni.cz/zmeny_povodni/pdf/sefrna.pdf [Accessed 8 July 2009]

[42] Bufková I. Ochrana rašelinišť na Šumavě : aneb byly Rokytecké slatě první. In: Černý D, Dvořák L, editors. Weitfällerské slatě, Sborník výzkumu na Šumavě, 2, Vimperk, Správa NP a CHKO Šumava. 2009. p. 103

[43] Vlček L, Falátková K, Schneider P. Identification of runoff formation with two dyes in a mid-latitude mountain headwater. Hydrology and Earth System Sciences. 2017;**21**:3025-3040. DOI: 10.5194/hess-21-3025-2017

[44] Schneider P, Lange A, Doppler T. Development and evaluation of a water level proportional water sampler, American Geophysical Union, Fall Meeting 2013, abstract #H43H-1579; 2013

[45] Šanda M. Proudění vody v půdním profilu na svahu horského povodí. Doktorské disertační minimum. Praha: ČVUT; 1998. p. 12

[46] Zhang W et al. The impact of vegetation and soil on runoff regulation in headwater streams. Catena. 2011;**87**:182-189

[47] Doležal T, Vlček L, Kocum J, Janský B. Evaluation of the influence of mountain peat bogs restoration measures on the groundwater level: Case study Rokytka peat bog, the Šumava Mts., Czech Republic. AUC Geographica. 2017, [S.l.], july 2017. ISSN 2336-1980, doi: 10.14712/23361980.2017.11

[48] Penman HL. Natural evaporation from open water, bare soil and grass. Proceedings of the Royal Society. 1948;**194**:120-145

[49] Mishra SK, Singh VP. Soil Conservation Service Curve Number (SCS-CN) Methodology. Vol. 511. Dordrecht: Kluwer Academic Publisher; 2003

[50] Bůzek F. A rapid procedure for preparing Oxygen-18 determination in water samples. Isotopenpraxis. 1984;**19**:70-72

[51] Käss W. Tracing Technique in Geohydrology. Rotterdam: A. A. Balkema; 1998. p. 581

[52] Gerke KM, Sidle RC, Mallants D. Criteria for selecting fluorescent dye tracers for soil hydrological applications using Uranine as an example. Journal of Hydrology and Hydromechanics. 2013;**61**:313-325. DOI: 10.2478/johh-2013-0040

[53] Weiler M, Flühler H. Inferring flow types from dye patterns in macroporous soils. Geoderma. 2004;**120**:137-153. DOI: https://doi.org/10.1016/j.geoderma.2003.08.014

[54] Weiler M, Naef F. An experimental study of the role of macropores in infiltration in grassland soils. Hydrological Processes. 2003;**17**:477-493

[55] Kocum J, Oulehle F, Janský B, Buzek F, Hruška J, Vlček L. Geochemical evidence for peat bog contribution to streamflow generation process: Case study of Vltava R. Headwaters, Czech Republic. Hydrological Sciences Journal. 2016;**61**:2579-2589. DOI: 10.1080/02626667.2016.1140173

[56] Eliáš V, Tesař M, Buchtele J. Occult precipitation: Sampling, chemical analysis and process modelling in the Šumava Mts. (Czech Republic) and in the Taunus Mts. (Germany). Journal of Hydrology. 1995;**166**:409-420

[57] Tesař M, Eliáš V, Šír M. Preliminary results of characterization of cloud and fog water in the mountains of southern and northern bohemia. Journal of Hydrology and Hydromechanics. 1995;**43**(6):412-426

[58] Pavlásek J, Tesař M, Máca P, Hanková R, Hudečková K, Jačka L, Klose Z, Ředinová J. Ten years of hydrological monitoring in upland microcatchments in the bohemian Forest, Czech Republic. IAHS-AISH Publication. 2010;**336**:213-218

[59] Ahearn EA. Peak-flow frequency estimates for U.S. Geological Survey streamflow-gaging stations in Connecticut. Water-Resources investigations report, 03-4196. U.S. Geological Survey, East Hartford, Connecticut; 2003. 36 p

[60] Feldman AD. Flood Hydrograph and Peak Flow Frequency Analysis. Technical Paper. Davis, California: HEC; 1979. p. 28

[61] Kocum J, Janský B. Retence vody v pramenných oblastech Vydry a Křemelné – případová studie povodí Rokytky. In: Černý D, Dvořák L, editors. Weitfällerské slatě. Sborník referátů ze semináře 21.1.2009. Vimperk: Správa NP a CHKO Šumava; 2009. pp. 26-48. ISBN 978-80-87257-00-5, ISSN 1803-4470

[62] Šercl P. Vliv fyzicko-geografických faktorů na charakteristiky teoretických návrhových povodňových vln. Sborník prací ČHMÚ, sv. 54, ČHMÚ, Praha; 2009. 88 p

[63] Štěpánková R. Vliv fyzicko-geografického prostředí na vznik a vývoj povodní v povodí Berounky. . Disertační práce. Praha: PřF UK; 2004. p. 178

[64] Tesař M, Šír M, Zelenková E. Vliv vegetace na vodní a teplotní režim tří povodí ve vrcholové části Šumavy. Aktuality šumavského výzkumu, s. 2004;**2**:84-88

[65] Lichner L, Šír M, Tesař M. Testování retenční schopnosti půd. Aktuality šumavského výzkumu. 2004;**2**:63-67

[66] Vlček L. Retence vody půdou v experimentálních povodích se zaměřením na organozemě. Magisterská práce. Praha: PřF UK; 2011. p. 67

[67] Šanda M, Císlerová M. Transforming hydrographs in the hillslope subsurface. Journal of Hydrology and Hydromechanics. 2009;**57**(4):264-275

[68] Vlček L, Kocum J, Janský B, Šefrna L, Blažková Š. Influence of peat soils on runoff process: Case study of Vydra R. headwaters, Czechia. Geografie. 2016;**121**:235-253

[69] Holden J, Burt TP. Piping and pipeflow in a deep peat catchment. Catena. 2002;**48**:163-199. DOI: https://doi.org/10.1016/S0341-8162(01)00189-8

[70] Rizzuti AM, Cohen AD, Stack EM. Using hydraulic conductivity and micropetrography to assess water flow through peat-containing wetlands. International Journal of Coal Geology. 2004;**60**:1-16. DOI: 10.1016/j.coal.2004.03.003

[71] Gerke KM, Sidle RC, Mallants D. Preferential flow mechanisms identified from staining experiments in forested hillslopes. Hydrological Processes. 2015;**29**:4562-4578. DOI: https://doi.org/10.1002/hyp.10468

[72] Holden J, Burt TP. Hydrological studies on blanket peat: The significance of the acrotelm-catotelm model. Journal of Ecology. 2003;**91**:86-102. DOI: 10.1046/j.1365-2745.2003.00748.x

7.3 Dynamika hladiny podzemní vody a retenční schopnost horského vrchoviště

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Influence of vegetation on groundwater level dynamics and calculation of retention ability of Rokytecká slať peat bog, Šumava Mountains

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SUMMARY

The hydrological functioning of peat bogs and their retention potential as a factor in the water balance of the landscape are important topics in the context of the increasing frequency of extreme droughts. This article evaluates water table dynamics and retention ability in a mountain bog, in relation to the vegetation cover type (spruce forest, shrub, sedges). The study draws on continuous observations of the groundwater level in various parts of Rokytecká slať bog over a period of six years, with a detailed evaluation of causative meteorological factors in the experimental catchment. The amount of water retained at specific locations with different vegetation was studied during extreme episodes of rapid snowmelt, intensive rainfall and long-lasting drought. It is shown that, although the groundwater level varies between different vegetation types, its fluctuations span similar ranges. The retention ability of spruce forest is high but tends to decrease rapidly during intensive rainfall. In contrast, during drought events, the most rapid drop in groundwater level is observed in sedge-dominated vegetation. During periods of snowmelt, long-lasting water retention is enabled if the bog surface is not totally frozen. Initial groundwater level is the factor that ultimately determines the amount of water infiltrated or evaporated during extreme episodes.

KEY WORDS: Czech Republic, hydrology, retention potential, water table

1. INTRODUCTION

Peat bogs are important for hydrological processes in the headwater catchments of rivers. Because of their unique water regimes, they can regulate runoff processes, especially during extreme meteorological events (Kocum 2012). Due to their high retention capacity, spring areas containing peat (soil type Histosols) are important subjects in the study of hydrological processes (Evans *et al.* 1999, Holden *et al.* 2001, Worrall *et al*. 2006).

Peat bogs cover 20–35 % of most catchment areas in the Šumava Mountains of the Czech Republic (Bufková *et al*. 2010). Their importance, vegetation, hydrological and pedological characteristics have been explored in previous studies (Bufková 2009, Kučerová *et al*. 2009, Kocum 2012, Vlčet *et al*. 2012), with particular focus on the context of the current trend towards increasing hydrological extremes (see also Vlček *et al*. 2016, Doležal *et al*. 2017). However, their retention potential and behaviour during extreme meteorological events have not yet been examined, and this remains an impediment to determining their ability to influence runoff processes (Janský & Kocum 2008).

In mountain bogs, the stability and dynamics of the groundwater level (*GWL*) are key factors not only for the hydrological situation of the catchment area, but also for the future development of the bog itself (Holden *et al*. 2001, Janský & Kocum 2008, Holden *et al*. 2011, Vlček *et al*. 2012). The main factors influencing *GWL* in ombrogenous bogs are precipitation, evapotranspiration, topography and, on a local scale, peat porosity and hydraulic conductivity (Allott *et al*. 2009). Land cover and land use create an essential building block of processes that fundamentally affect the soil water regime; for example interception, evapotranspiration or the water retention characteristics of the soil.

GWL determines the species composition of bog vegetation (Labadz *et al*. 2010, Bufková *et al*. 2012), which is physically and ecologically adapted to high water table and very responsive to changes in the water regime. For example, changes in *GWL* can affect the thermal conditions of peat bogs and cause variations in evapotranspiration (Kučerová *et al*. 2009); and a long-term change in *GWL* can cause succession. Thus, *GWL* has substantial implications for the ecological niches of plant species, and even for peat development (Kværner & Snilsberg 2011). The vegetation pattern, in turn, has important effects on hydrological processes and *GWL* dynamics in bogs. Strong correlation has been observed between the vegetation type and *GWL* stability. Even in small catchments, substantial differences in evapotranspiration of various vegetation types were observed. A denser vegetation cover also results in improvement of the self-regulating process of peat, as evaporation is reduced due to the higher level of air moisture maintained (Binet *et al*. 2013). Nevertheless, it is important to mention that the actual hydroclimatic conditions in the catchment must be taken into consideration as the various vegetation types show high variability of soil moisture during extreme episodes (Šípek *et al*. 2020). Brom *et al*. (2009) mention considerable differences in the microclimate and soil water regime of various vegetation types. The energy balance of such vegetation types is the principal cause of the soil moisture dynamics and the water regime dynamics in the landscape.

GWL in bogs responds rapidly to meteorological factors and can vary by several centimetres per day (Vlček *et al*. 2012). In many types of bogs, the dynamics of *GWL* show considerable seasonal and interannual variations, especially during extreme hydrological episodes. In particular, raised bogs and waterlogged forests whose sole water supply is precipitation show considerable lowering of the water table during summer drought whereas, in winter, water that originates from snowfall is retained on the bog surface although runoff is still possible (Kučerová *et al*. 2009). During long-lasting drought episodes with low water table, irreversible changes in the physical properties of peat may occur and thus the hydrological processes in the landscape may also be affected in a negative way (Evans *et al*. 1999).

The retention potential of peat bogs is very variable and depends greatly on soil properties. When considering the average groundwater level in peat bogs (a few centimetres below the surface), catchments represent areas with small retention values similar to those found in shallow soils. Under conditions of low groundwater level, however, bogs represent significant retention areas within the catchment (Kocum *et al*. 2018).

The main aims of the present study are:

a) to describe the *GWL* dynamics in various types of peat vegetation;
b) to determine the *GWL* dynamics of various vegetation types during extreme episodes (intensive rainfall, long-lasting drought and rapid snowmelt);
c) to calculate the retention potential of a peat bog during monitored episodes.

2. STUDY AREA

The Rokytka experimental catchment is situated in the central part of the Šumava Mountains and constitutes the headwater area of the Vydra river. The experimental catchment covers a total area of 4 km^2 and is situated at an altitude of approximately 1,100 metres.

The catchment area is divided into three parts, according to their dominant vegetation cover: a peat bog with sedges, a peat bog covered in dense ericaceous shrubs and a waterlogged forest (Figure 1). In the central part of the catchment area, there is a mountain raised bog, covered with non-forest vegetation, particularly with various species of sedges. Among others, the most frequent sedge is the bog sedge (*Carex limosa*), accompanied by typical mire vegetation, e.g. deer grass (*Trichophorum caespitosum*), marsh cranberries (*Vaccinium oxycoccos*) and cottongrass (*Eriophorum vaginatum*).

There are also various sphagnum mosses, e.g. feathery bogmoss (*Sphagnum cuspidatum*), magellanic bogmoss (*Sphagnum magellanicum*) or olive bogmoss (*Sphagnum majus*). The mountain raised bog is densely surrounded by bog pine (*Pinus mugo*). The last part of the catchment, which is also the largest one, is covered by waterlogged forest, dominated by Norway spruce (*Picea abies*). The forest used to be hit by the bark beetle in the past and up to now, there is lot of organic matter in gradual decomposition. The prevailing species of the herbaceous layer are European blueberry (*vaccinium myrtillus*), bog blueberry (*vaccinium uliginosum*) and different sphagnum and moss species.

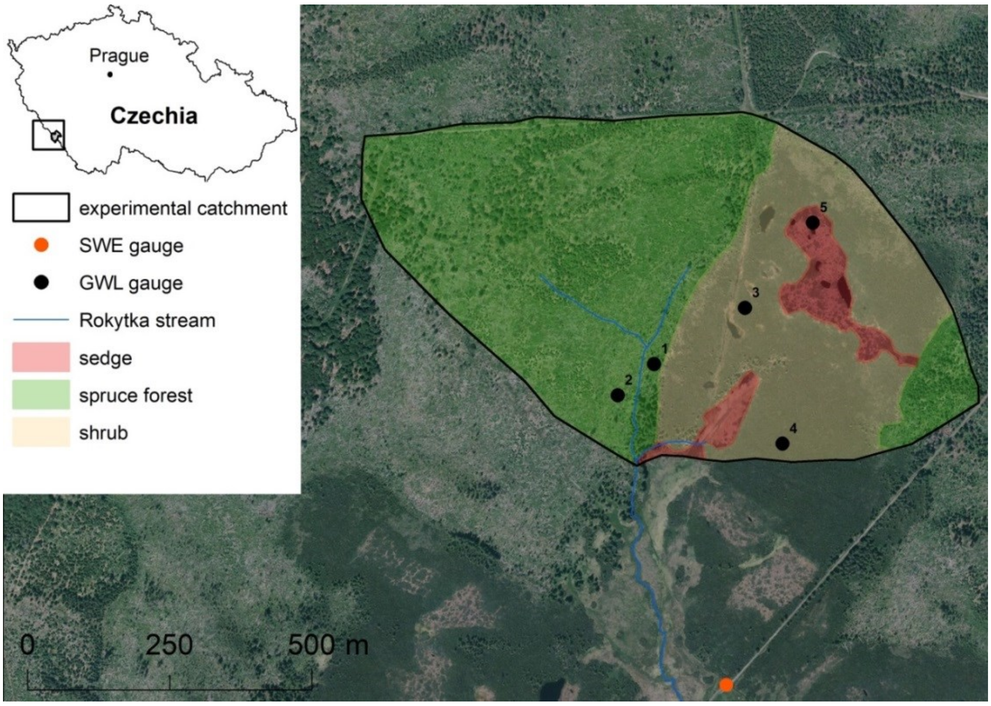


Figure 1. Experimental catchment with designation of different vegetation parts and positions of groundwater level (*GWL* 1-2 spruce forest, 3-4 shrub, 5 sedge) and snow water equivalent (*SWE*) gauge.

The long-term average annual precipitation alters between 1100 and 1300 mm and the average temperature is approximately 5°C. The annual discharge minimum is usually reached at the end of February (before the snow melting) or in September (at the end of summer). The maximum discharge is generally in spring, when the snow melts (Kocum 2012).

The soil of Rokytka catchment is a typical example of the Šumava Mountains soils, where a vertical sequence of several soil types with Histosols is common. The catchment is partly covered by entic Podzol. In certain concave parts of the catchment, Gleysol can be found (Vlček *et al*. 2016). The largest part of the catchment is covered by Histosols, where the depth of peat can reach up to 7 m in some parts (Bufková 2009).

3. MATERIALS AND METHODS

Precipitation and all the meteorological data used in this paper were measured at the nearest meteorological station with a long-term dataset at Modrava village (5 km from the experimental catchment, at the same altitude). Snow water equivalent data (*SWE*) was measured on an open space area at the Rokytecká slať. Both variables were measured at 10 minute intervals. The most valuable part of the study focused on the *GWL* dynamics in a peat bog. The measurements were taken in 5 tubes (2 in waterlogged forest, 2 in shrub, 1 in sedge) to represent the monitored vegetation types in the peat bog (Figure 2). Hydrostatic level probes (TSH22, Fiedler company) for the *GWL* measurements were placed 70 cm deep in the tubes. Wireless transmission was provided for the entire duration of the monitoring process at 10 minute intervals.

Figure 2. Groundwater level gauge in experimental sites – shrub, spruce forest, sedge.

The study also evaluates the dynamics of changes in the groundwater level during dry periods in relation to daily potential evapotranspiration (*PET*). The potential evapotranspiration was calculated by the Penman–Monteith equation (Equation 1).

$$(1)\ PET_0 = \frac{0{,}408.\Delta.(Rn-G)+\gamma.\frac{900}{T+273{,}16}.u.(e_s-e_a)}{\Delta+\gamma.(1+0{,}34.u)}$$

where Δ represents the inclination of the water vapour saturation curve in connection with the temperature [kPa.°C^{-1}], **Rn** the radiation balance [MJ.m^{-2}.day^{-1}], **G** the flow of heat into the soil [MJ.m^{-2}.day^{-1}], $\boldsymbol{\gamma}$ a psychometric constant [kPa.°C^{-1}], **u** the wind speed [m.s^{-1}], **(e_s-e_a)** the saturation deficit of air at elevation *z* [kPa], and **T** the average air temperature [°C] (Penman 1948).

The *GWL* dynamics evaluation and the retention potential calculation of a peat bog were based on the soil moisture data, collected in the same area of interest during the previous pedological research led by Vlček *et al.* (2012). It was detected that the immediate average volumetric water content in the area of interest at Rokytka catchment was 0.7 g/cm^3, the maximum volumetric water content was 0.95 g/cm^3 and the minimum volumetric water content was 0.5 g/cm^3. Based on these measurements, two equations for the calculation of the retention potential during extreme episodes of intensive rainfall or rapid snowmelt (Equation 2) and long-lasting drought (Equation 3) were determined for the purpose of this study:

$$(2)\ RP = \frac{GWL(b)}{GWL(e)} * [(VWC_{max} - VWC_{mean}) * 1000]$$

$$(3)\ RP = \frac{GWL(b)}{GWL(e)} * [(VWC_{mean} - VWC_{min}) * 1000]$$

where **RP** is the retention potential [l/m^2], **GWL(b)** represents the value of the groundwater level at the beginning of the observed episode [mm], **GWL(e)** stands for the groundwater level at the end of the observed episode [mm], $\mathbf{VWC_{max}}$ is the maximum volumetric water content [g/cm^3], $\mathbf{VWC_{mean}}$ is the average volumetric water content [g/cm^3], and $\mathbf{VWC_{min}}$ is the minimum volumetric water content [g/cm^3].

It was assumed for the calculations that the initial value at the beginning of each episode was the average one (0.7 g/cm^3). Then only at extreme episodes of intensive rainfall and rapid snowmelt the values grew to the maximum value (0.95 g/cm^3). By contrast, during the long-lasting drought episodes, the initial value decreased from the average to the minimum value

($0.5\ g/cm^3$). These values were measured during previous detailed pedological research in the area of interest (Vlček *et al.* 2012).

The examined episodes were chosen appropriately to represent equally and homogeneously all the monitored vegetation types, while occasional failures in recordings from individual gauges had to be taken into consideration. In the case of two values representing the same monitored vegetation type (spruce forest, shrub) available, the final value of the infiltrated or evaporated water was obtained as their average value, related to the total area of the respective vegetation type.

4. RESULTS

Comparison of the *GWL* dynamics of the respective vegetation types revealed significant differences (Figure 3). The highest *GWL* was observed in the sedge site, which is situated in the central part of the peat bog. Several times during the periods of intensive precipitation, the peat became fully saturated and the water table rose above the ground surface, infilling with water all the minor depressions of the peat mass. This extraordinary behaviour was also registered during the field research. In addition, it was equally detected twice at the shrub 1 site. On the other hand, the lowest *GWL* was registered at the spruce forest 2 site, which showed intensive diminution of *GWL* during long-lasting summer drought events, particularly in the very dry summer of 2015.

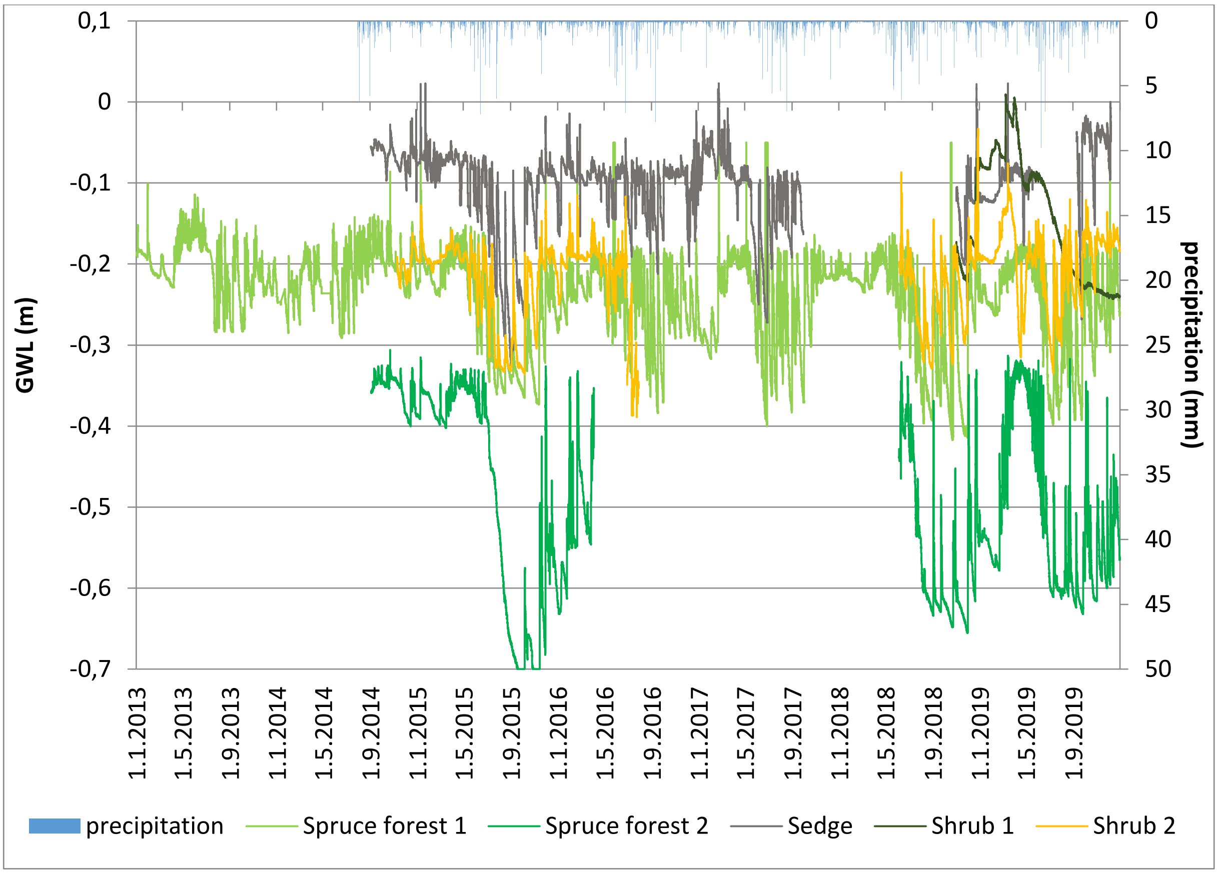


Figure 3. Groundwater dynamics in observed vegetation types of peat bog.

The basic statistical characteristics of the *GWL* records for the various vegetation types prove an extremely low *GWL* at the spruce forest 2 site, which reached the lowest possible measurable extent (the tube was placed at a depth of 70 cm below the surface), so its real level could be situated even lower (Table 1). The results obtained show that forest cover may show high variability of *GWL*, related to the physico-geographical features of the microsite.

By contrast, relatively high values of the average *GWL* were measured in the sedge sites, which corresponds to the natural state of the central part of an intact mountain peat bog. The respective average values of *GWL* determine the presence of a particular vegetation type. After examining the *GWL* range, comparable values of *GWL* were discovered (except for the lower range detected in the shrub 1 site, compared to other vegetation types, which could be caused by a shorter period of observation). Similar variability of all sets of values is confirmed by the range of standard deviations. Thus it can be deduced that, although *GWL* differs vertically between vegetation types by units of centimetres, it fluctuates over a very similar range.

Table 1. Basic information of subcatchments and statistics evaluation of the groundwater level.

	Spruce forest 1	Spruce forest 2	Sedge	Shrub 1	Shrub 2
Beginning of measurement	1.1.2013	4.9.2014	3.9.2014	2.11.2018	10.11.2014
Area (m^2)	291 257		29 499	169 530	
Average GWL (m)	-0.24	-0.49	-0.10	-0.15	-0.22
Maximum GWL (m)	-0.05	-0.31	0.02	0.01	-0.03
Minimum GWL (m)	-0.42	-0.70	-0.33	-0.24	-0.39
Range of GWL (m)	0.,37	0.39	0.35	0.25	0.36
Standard deviation	0.06	0.12	0.05	0.07	0.05
Coefficient of variation	23.63	23.38	48.01	50.74	23.39

4.1 Intensive rainfall episodes

Graphical representations of the monitored events (Figure 4) indicate that each vegetation type behaves in a different manner and has specific water retention and interception abilities. In the first case, there was only a moderate elevation of the water table for the sedge and the shrub 1 sites because the *GWL* values were already high before the pertinent rainfall event, meaning that the soil became saturated very quickly and could not receive any more water. In the other two other cases, *GWL* was relatively low at the beginning of the measurement and thus a larger amount of water could be absorbed. Concerning the spruce forest 2, its very low initial *GWL* was the principal cause of high water retention during the first monitored event; however, the same situation was not reproduced during the rainfall 3 event. In spite of intensive precipitation, *GWL* did not rise above -30 cm at the spruce forest 2 site.

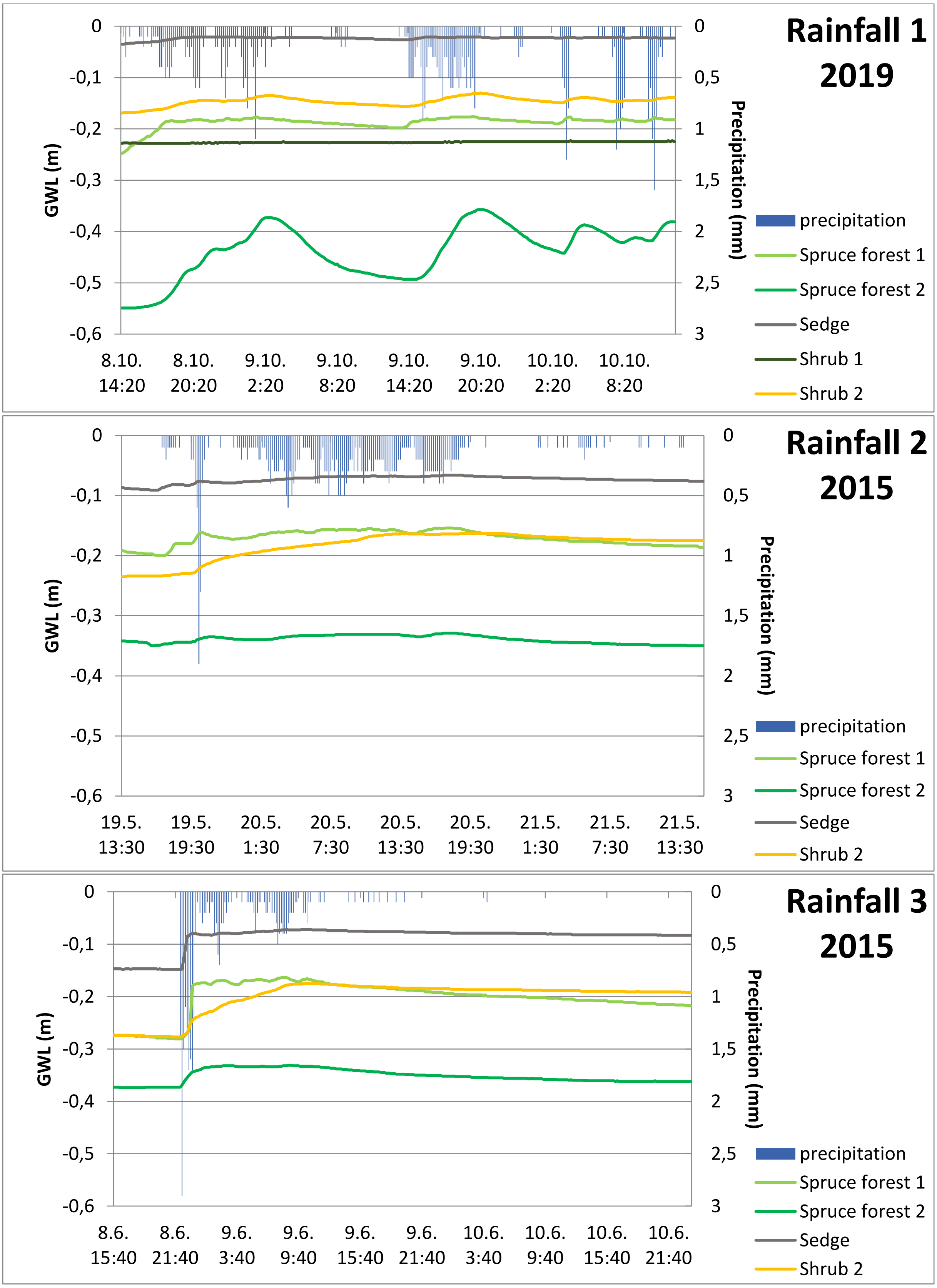


Figure 4. Groundwater level dynamics during the rainfall episodes 1–3.

The respective vegetation types proved to have very different retention potential values during the three monitored episodes (Table 2). The most impressive amount of retained water was detected at the spruce forest 2 site during the first precipitation event (83%).

It should be noted that after a rapid elevation of the *GWL*, a rapid decline followed, and so the result implies a short-term water retention, with a quick runoff of infiltrated water. An important retention potential of a peat bog was revealed during the most significant rainfall event, when the maximum water storage capacity reached over 9,000 m^3; with regard to the two remaining episodes, the total water storage capacity of the peat bog remained high. Considering the wide range in the amount of infiltrated water at the monitored vegetation types, we may conclude that the principal factor affecting the progress of the event is the actual saturation of the headwater.

Table 2. Changes of groundwater level and calculation of volume of infiltrated water during rainfall episodes.

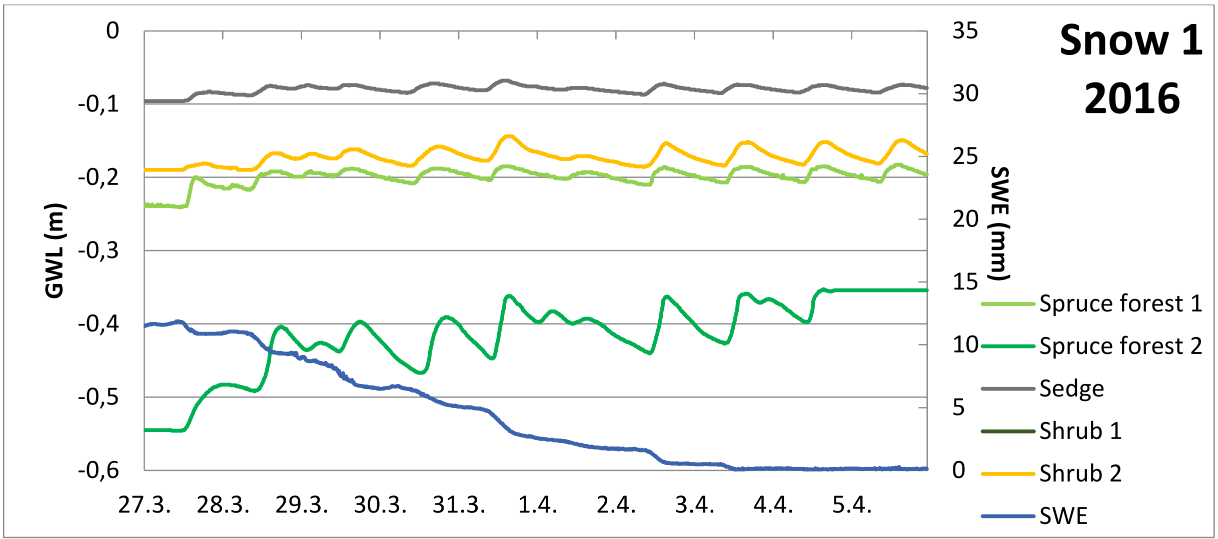
Rainfall 1 (50.6 mm)	**Spruce forest 1**	**Spruce forest 2**	**Shrub 1**	**Shrub 2**	**Sedge**
Increase in GWL (mm)	71	168	3	26	13
Amount of infiltrated precipitation (l/m^2)	17.8	42.0	0.7	6.5	3.3
Proportion of infiltrated precipitation (%)	35.0	83.0	3.0	12.8	6.4
Volume of infiltrated water (m^3)	8701.3		614.5		95.9
Total volume of water (m^3)	9411.7				

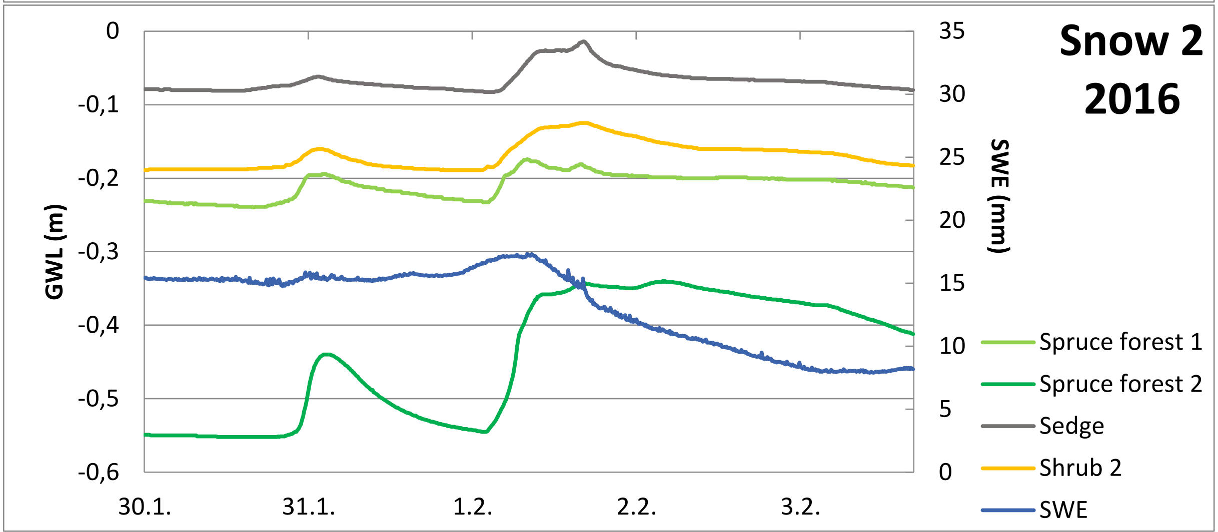
Rainfall 2 (37.4 mm)	**Spruce forest 1**	**Spruce forest 2**	**Shrub 1**	**Shrub 2**	**Sedge**
Increase in GWL (mm)	42	18		71	24
Amount of infiltrated precipitation (l/m^2)	10.5	4.5		17.8	6.0
Proportion of infiltrated precipitation (%)	28.0	12.0		47.5	16.0
Volume of infiltrated water (m^3)	2184.4		3009.2		176.9
Total volume of water (m^3)	5370.5				

Rainfall 3 (28.3 mm)	**Spruce forest 1**	**Spruce forest 2**	**Shrub 1**	**Shrub 2**	**Sedge**
Increase in GWL (mm)	55	11		83	64
Amount of infiltrated precipitation (l/m^2)	13.7	2.8		20.7	16
Proportion of infiltrated precipitation (%)	48.6	9.8		73.3	56.5
Volume of infiltrated water (m^3)	2402.9		3517.7		471.9
Total volume of water (m^3)	6392.5				

4.2 Snow melt episodes

Despite the possibility of a partially frozen surface, which would hinder the water infiltration, a significant volume of infiltrated water was stored in the majority of registered cases. In one particular situation (snow 3 – shrub 1), the water table was initially almost coincident with the ground surface and thus no infiltrated water was detected. The largest volume of infiltrated water was observed at the spruce forest 2 site, most probably because *GWL* was low at the beginning of the monitored event (Figure 5). At this point it should be noted that the measurements of SWE were taken in an open space, so the progress of snowmelt under forest or shrubs may be slightly different.





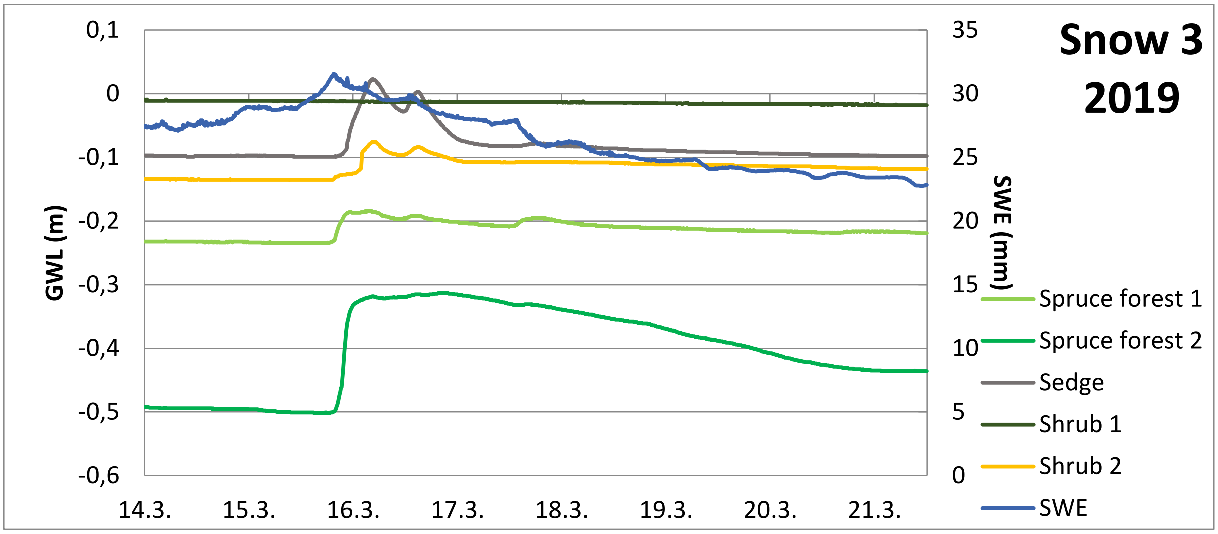


Figure 5. Groundwater level dynamics during the episodes of snow melting 1-3.

The total amount of stored water during the monitored events was approximately 10,000 m^3 (Table 3), of which the most decisive storage capacity belongs to forest cover, while shrub and sedge sites could only retain a maximum of 12 % of the total volumetric infiltrated water.

The differences between the spruce forest 1 and spruce forest 2 sites can be explained by the less dense vegetation cover of the spruce forest 2 site, which indicates not only more rapid snowmelt, but probably also a low *GWL*, as the principal cause, at the beginning of the monitored event. The rise in the *GWL* was not followed by a significant decline, as was observed during the intensive rainfall events.

Table 3. Changes of groundwater level and calculation of volume of infiltrated water during snow melting.

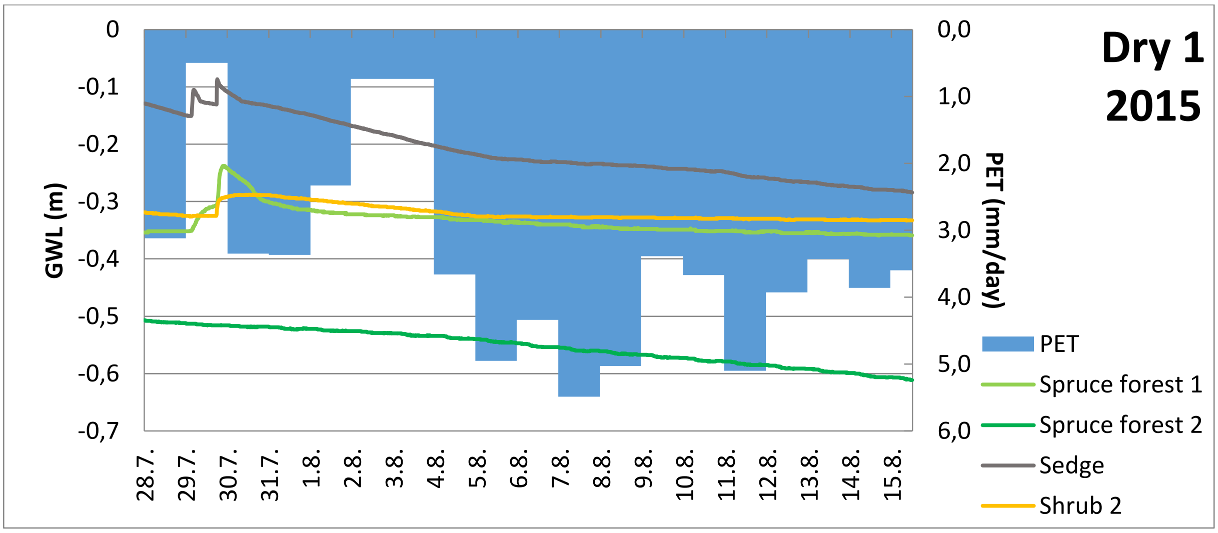
Snow 1 (140 mm)	**Spruce forest 1**	**Spruce forest 2**	**Shrub 1**	**Shrub 2**	**Sedge**
Increase in GWL (mm)	43	191		22	18
Amount of infiltrated water (l/m^2)	10.8	47.8		5.5	4.5
Proportion of infiltrated water (%)	7.7	34.1		3.9	3.2
Volume of infiltrated water (m^3)	8519.2		932.4		132.7
Total volume of water (m^3)	9584.3				

Snow 2 (93.7 mm)	**Spruce forest 1**	**Spruce forest 2**	**Shrub 1**	**Shrub 2**	**Sedge**
Increase in GWL (mm)	37	204		35	21
Amount of infiltrated water (l/m^2)	8.5	51.0		8.7	5.3
Proportion of infiltrated water (%)	9.0	54.4		9.3	5.6
Volume of infiltrated water (m^3)	8664.9		1483.4		154.9
Total volume of water (m^3)	10303.2				

Snow 3 (89 mm)	Spruce forest 1	Spruce forest 2	Shrub 1	Shrub 2	Sedge
Increase in GWL (mm)	37	189	0	37	50
Amount of infiltrated water (l/m^2)	9.3	47.3	0.0	9.3	12.5
Proportion of infiltrated water (%)	10.3	53.1	0.0	10.4	14.0
Volume of infiltrated water (m^3)	8228.0		784.1		368.7
Total volume of water (m^3)	9380.8				

4.3 Dry episodes

The most intensive of the three dry events was in summer (event 1 of 2015), when the daily potential evapotranspiration (*PET*) reached its highest values (Figure 6). Due to *PET*, the sedge part of the peat bog lost a significant amount of groundwater (Table 4). This was most probably caused by the lower vegetation cover and thus more intensive soil evaporation. A specific behaviour was registered at spruce forest 2 site, which showed low *GWL* at the beginning of the monitoring, with only a minor and gradual decrease during the examined dry periods. After the dry 2 event, *GWL* reached its measurable minimum. A part of the peat bog shrub site showed variable behaviour during the monitored period.



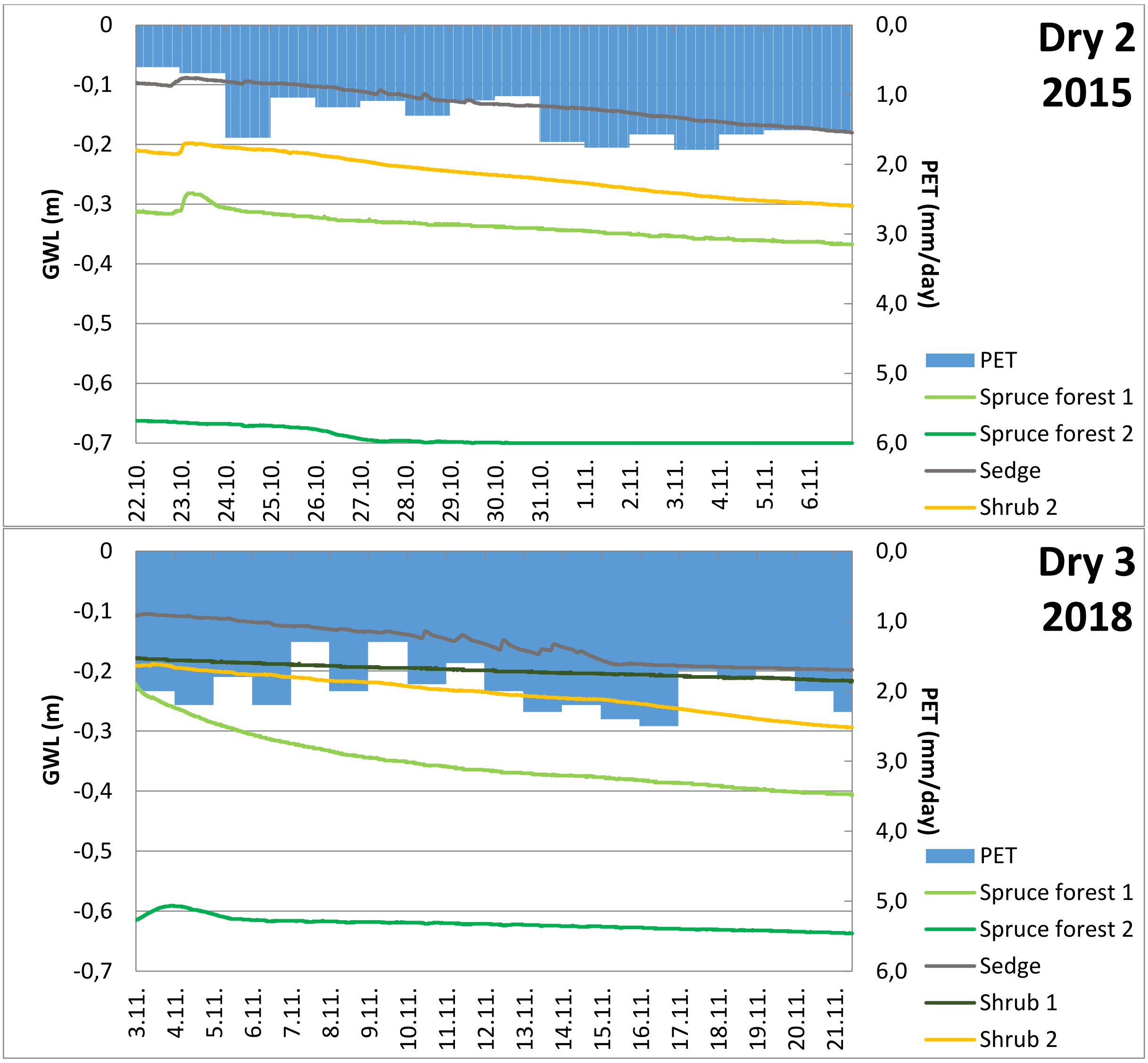


Figure 6. Groundwater level dynamics and daily potential evapotranspiration (PET) during dry episodes 1-3.

The summer episode induced a small decrease of *GWL*, then the level stabilised at a depth of 30 cm below the surface. During the autumn season, without precipitation, a decrease of *GWL* was observed, but after reaching the above-mentioned height, it did not descend any more. Similar behaviour was also registered at spruce forest 1 site, with *GWL* stabilised at a depth of 40 cm below the surface.

Table 4. Changes of groundwater level and calculation of volume of infiltrated water during dry episodes.

Dry 1 (69.2 mm)	Spruce forest 1	Spruce forest 2	Shrub 1	Shrub 2	Sedge
Decrease in GWL (mm)	44	89		36	135
Amount of evaporated water (l/m^2)	8.8	17.8		7.2	27.0
Proportion of evaporated water (%)	12.7	25.7		10.4	39.0
Volume of evaporated water (m^3)	3873.7		1220.6		796.4
Total volume of water (m^3)	5890.7				

Dry 2 (20.4 mm)	Spruce forest 1	Spruce forest 2	Shrub 1	Shrub 2	Sedge
Decrease in GWL (mm)	55	37		93	83
Amount of evaporated water (l/m^2)	11.0	7.4		18.6	16.6
Proportion of evaporated water (%)	53.9	36.3		91.1	23.9
Volume of evaporated water (m^3)	2679.6		3153.3		489.7
Total volume of water (m^3)	6322.6				

Dry 3 (37.3 mm)	Spruce forest 1	Spruce forest 2	Shrub 1	Shrub 2	Sedge
Decrease in GWL (mm)	186	22	38	103	90
Amount of evaporated water (l/m^2)	37.2	4.4	7.6	20.6	18.0
Proportion of evaporated water (%)	99.8	11.8	20.4	55.2	48.3
Volume of evaporated water (m^3)	6058.1		2390.4		530.9
Total volume of water (m^3)	8979.4				

5. DISCUSSION

Concerning the various vegetation types, differences in the *GWL* were observed. After comparing different parts of a peat bog, similar results to those already mentioned by Kučerová *et al*. (2009) and Labadz *et al*. (2010) were obtained. The highest *GWL* is in the centre of the sedge part of the a peat bog, at 10 cm depth below the surface. The long-term average depth of the *GWL* for shrubs is at approximately 20 cm. In waterlogged forests, the observed values vary, depending on the measurement points (spruce forest 1 – 21 cm, spruce forest 2 – 49 cm below the surface).

There are many factors that affect the *GWL* of the examined microsites, i.e. topography, peat porosity, hydraulic conductivity, etc. (Allott *et al*. 2009). The spruce forest site is recovering from a bark beetle disturbance and is at present at different stages of natural regeneration. In addition, in the past, it was affected by the anthropogenic influence of the water regime caused by man-made drainage channels, which are nowadays being blocked and refilled with natural materials. Thus it is not evident which of the listed factors is the key one. On the other hand, the obtained *GWL* values correspond to the results of previous research carried out by Bufková *et al*. (2010) at similar sites in the Šumava Mountains. All the monitored sites proved a similar amplitude and oscillated in the range of approx. 35 cm around their average *GWL*, except for the shrub 1 site, which could have been caused by a shorter observation period. The biggest rise in the *GWL* was registered at the waterlogged forest site – 17 cm during the monitored period – which indicates that the *GWL* elevations can be rapid, as was already confirmed by previous research conducted by Vlček *et al*. (2012).

The analysis of the retention potential of a peat bog during extreme meteorological events proved the relevance of the initial *GWL* at all the monitored sites. The infiltration capacity is thus directly related to the former saturation of a peat bog (Doležal *et al*. 2017). The proportion of infiltrated precipitation and short-term water retention oscillated in a wide range of 3–83 %. After reaching the maximum saturation for the respective vegetation types, which are situated at the surface for sedge and shrub and at max. values of 5–30 cm below the surface for forest cover, the retention potential becomes exhausted and results in the formation of runoff.

The maximum short-term retention potential value during all the intensive rainfall events reached 42 l/m^2. In general, the retention potential during rainfall events proved to be extremely variable, depending on the current soil saturation. Mountain peat bog can, under certain conditions, appear as an area of high retention potential, but in the case of extensive prior saturation, this characteristic is no longer applicable (Kocum *et al*. 2018). Moreover, another effect of the initial *GWL* manifested itself at extreme episodes of rapid snowmelt. In the winter season, the peat bog surface is rarely totally frozen and thus it can, during periods of thaw, infiltrate water from melting snow, as was already mentioned in Kučerová *et al*. (2009). Taking all the monitored snowmelt events into consideration, the total amount of infiltrated water did not exceed 50 %. However, significant differences were observed, depending on the initial soil saturation and the ratio to which the surface was frozen as well. For one particular case, no infiltration of snowmelt water was registered. In particular, the sedge and shrub parts of the peat bog showed low volumes of infiltrated water.

The maximum total volume of short-term infiltrated snow water for the entire period of observation was still around 10,000 m^3. The sedge and shrub cover showed significant declines of the *GWL* during extreme dry events, which were caused by excessive evaporation, resulting from lower vegetation cover. A similar context was already described by Binet *et al*. (2013). Nevertheless, the *GWL* in the above-mentioned vegetation cover did not fall below 40 cm (below the surface) during the entire monitored period, which is a fundamental requirement for keeping the central part of a peat bog in a healthy condition. Long-term drainage of the vulnerable central parts of a peat bog can cause irreversible succession changes (Kučerová *et al*. 2009, Kværner & Snilsberg 2011). As for the extreme rainfall events, high variability of the *GWL* was observed in the forest cover. The most significant decline of the GWL – 19 cm – was registered during a monitored period lasting for 18 days, while the total daily *PET* reached 37 mm during the same period. Nevertheless, even constituent parts of a small headwater area can show high variability of evapotranspiration, which is determined by many factors (Šípek *et al*. 2020).

It is also important to take possible uncertainties of measurement into consideration. The retention potential calculation of a raised bog was based on the volumetric soil moisture data, collected in the same area of interest during the previous pedological research (led by Vlček *et al*. 2012 and Vlček *et al*. 2016). The default values do not necessarily correspond to the authentic humidity at the beginning of the event. The entire set of meteorological data was not recorded directly in the area of interest, but at the nearest meteorological station at Modrava village (at a distance of 5 km from the experimental catchment, at the same altitude), so the starts of the respective monitored episodes could be moderately shifted in time and thus not correspond to the exact formations of precipitation. However, for the total volumetric values of precipitation and the calculations of potential evapotranspiration, the measuring deviations are irrelevant.

REFERENCES

Allott, T.E.H., Evans, M.G., Lindsay, J.B., Agnew, C.T., Freer, J.E., Jones, A., Parnell, M. (2009) *Water tables in Peak District blanket peatlands*. Moors for the future, Edale, Derbyshire, 49 pp.

Binet, S., Gogo, S., Laggoun-Défarge, F. (2013) A water-table dependent reservoir model to investigate the effect of drought and vascular plant invasion on peatland hydrology. *Journal of Hydrology*, 499, 132-139.

Brom, J., Procházka, J., Rejšková, A. (2009) Evaluation of functional properties of various types of vegetation cover using remotely sensed data analysis. *Soil and Water Research*, 4 (special issue 2), 49-58.

Bufková, I. (2009) Ochrana rašelinišť na Šumavě: aneb byly Rokytecké slatě první?. In: Černý, D., Dvořák, L. (eds) Weitfällerské slatě: Sborník výzkumu na Šumavě. 2. Vimperk: Správa NP a CHKO Šumava, 103 pp.

Bufková, I., Stíbal, F., Mikulášková E. (2010) Restoration of drained mires in the Šumava National Park, Czech Republic. In: Eiseltová, M. (ed.) Restoration of lakes, streams, floodplains, and bogs in Europe: principles and case studies. Springer-Verlag, 331-354.

Bufková, I., Stíbal, F. (2012) Restoration of drained mires in the Šumava National Park. In: Jongepierová, I., Pešout, P., Jongepier, J.W., Prach, K. (eds) Ecological restoration in the Czech Republic. Nature Conservation Agency of the Czech Republic, Prague, 78-80.

Doležal, T., Vlček, L., Kocum, J., Janský, B. (2017) Evaluation of the influence of mountain peat bogs restoration measures on the water table level: case study Rokytka peat bog, the Šumava Mts., Czech Republic. Acta Universitatis Carolinae. Geographica. Univerzita Karlova, 52, 2, 1–10.

Evans, M.G., Burt, T.P., Holden, J., Adamson, J.K. (1999) Runoff generation and water table variations in blanket peat: evidence from UK data spanning the dry summer of 1995. Journal of Hydrology, 221(3–4), 141–160.

Holden, J., Burt, T.P., Cox, N.J. (2001) Macroporosity and infiltration in blanket peat: the implications of tension disc infiltrometer measurements. Hydrological Processes, 15(2), 289-303.

Holden, J., Wallage, Z.E., Lane, S., McDonald, A.T. (2011) Water table dynamics in undisturbed, drained and restored blanket peat. *Journal of Hydrology*, 402 (1–2), 103-114.

Janský, B., Kocum, J. (2008) Peat bogs influence on runoff process: case study of the Vydra and Křemelná River basins in the Šumava Mountains, southwestern Czechia. *Geografie – Sborník ČGS*, 113(4), 383–399.

Kocum, J. (2012) Tvorba odtoku a jeho dynamika v pramenné oblasti Šumavy. Disertační práce. Univerzita Karlova, Přírodovědecká fakulta v Praze, Katedra fyzické geografie a geoekologie, 206 pp.

Kocum, J., Janský, B., Vlček, L., Doležal, T. (2018) Hydrological function of a midlatitude headwater peatland. In: Topcuoğlu, B., Turan, M. (eds.) *Peat*. IntechOpen, 141-168.

Kučerová, A., Kučera, T., Hájek, T. (2009) Mikroklima a kolísání hladiny podzemní vody v centrální části Rokytecké slati. In Černý, D., Dvořák, L. (eds.) *Weitfällerské slatě*. Sborník referátů ze semináře 21.1.2009. Správa NP a CHKO Šumava, Vimperk, 50–57.

Kværner, J., Snilsberg, P. (2011) Groundwater hydrology of boreal peatlands above a bedrock tunnel – Drainage impacts and surface water groundwater interactions. *Journal of Hydrology*, 403(3), 278-291.

Labadz, J., Allott, T., Evans, M., Butcher, D., Billett, M., Stainer, S., Yallop, S., Jones, P., Innerdale, M., Harmon, N., Maher, K., Bradbury, R., Mount, D., O´Brien, H., Hart, R. (2010) *Peatland hydrology*. Draft scientific review, commissioned by the IUCN UK Peatland Programmes Commission of Inquiry on Peatlands, 52 pp.

Penman, H.L. (1948) Natural evaporation from open water, bare soil and grass. *Proceedings of the Royal Society of London. Series A, Mathematical and Physical Sciences,* 193(1032), 120–145.

Šípek, V., Hnilica, J., Vlček, L., Hinlicová, S., Tesař, M. (2020) Influence of vegetation type and soil properties on soil water dynamics in the Šumava Mountains (Southern Bohemia). *Journal of Hydrology*, 582, 124285.

Vlček, L., Kocum, J., Janský, B., Šefrna, L., Kučerová, A. (2012) Retenční potenciál a hydrologická bilance horského vrchoviště: případová studie Rokytecké slatě, povodí horní Otavy, JZ. Česko. *Geografie* 117(4), 395–414.

Vlček, L., Kocum, J., Janský, B., Šefrna, L., Blažková, Š. (2016) Influence of peat soils on runoff process: case study of Vydra River headwaters, Czechia. *Geografie*, 121(2), 235-253.

Worrall, F., Burt, T., Adamson, J. (2006) Long-term changes in hydrological pathways in an upland peat catchment—recovery from severe drought? *Journal of Hydrology*, 321(1), 5–20.

7.4 Hydrologický režim a fyzikálně-chemické vlastnosti rašelinných vod

Doležal, T., Vlček, L., Kocum, J., Janský, B. (2020): Hydrological regime and physico-chemical water properties of various types of peat bog sites: case study of Mezilesní peat bog, Šumava Mts. Geografie, 125, 1, 21 - 46.

Hydrological regime and physico-chemical water properties of various types of peat bog sites: case study of Mezilesní peat bog, Šumava Mts.

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ABSTRACT In a period with frequently occurring hydrological extremes, research on areas with a high retention potential is brought into focus. The Šumava Mountains peat bogs are important parts of the landscape in the headwater area of the Otava river basin. The study objective is to describe the variability of discharges and the dynamics of groundwater level changes in various types of peat bogs, and to identify connections between observed physico-chemical water properties. This is assessed by basic statistical methods. The rainfall-runoff process and physico-chemical water properties can be affected by many factors. In this case, strong relations between the observed parameters were identified along with considerable differences in the involvement of various types of peat bog sites in the runoff process. It is evident that the peat bog pattern and its vegetation cover have an essential effect on the hydrological regime and water properties stored in a peat bog.

KEY WORDS peat bog – Šumava Mts. – groundwater level – hydrological regime – headwater area

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1. Introduction

The headwater areas of streams represent resource areas for runoff formation. These areas are very heterogeneous in terms of their physical-geographical and rainfall-runoff aspects. The existence of important areas of peat bog complexes, an ecological phenomenon of local landscape, constitutes an important special characteristic of the Czech part of the Šumava Mts. It is therefore essential, in a specific territory of the headwater area of the Otava river, to deal with the effect of this phenomenon on hydrological regime and runoff formation (Kocum 2012; Čurda, Janský, Kocum 2011). The peat bogs have an important effect on hydrological processes in the landscape (Janský, Kocum 2008). The water storage capacity of the soil is one of the most important water retention components in the landscape. The physical properties of the soil and its general condition are the main factors controlling infiltration into the soil during precipitation events or the length the water can be stored in peaty soil during dry periods (Vlček et al. 2012, Vlček 2017). In the case of the Šumava Mts., the anthropogenic influence on the water regime of peat bogs that were drained in the past or even extracted is also an important factor. Particular interventions are currently being produced by means of restoration measures and blocking up of drainage ditches (Doležal et al. 2017).

Acrotelm plays the most important role in hydrological processes of a peat bog. It is a thin living surface layer of peat-forming vegetation, generally between 10 cm and 40 cm deep, relatively inert. Without acrotelm a bog cannot accumulate peat or control water loss from the deeper layer (Lindsay, Birnie, Clough 2014). The research of peat bogs in the Šumava Mts. is mostly focused on variability of the groundwater level (GWL) in the peat bog, depending on vegetation, or mitigation of anthropogenically disturbed water regime. For example, a research on the peat bog at Schachtenfilz (Bavarian side of the Šumava Mts.) proved that the GWL reached on average higher values in shrub vegetation on an open bog and in waterlogged forest. The GWL on a peat bog covered by grasslands with cottongrass showed greater variation (Bufková, Stíbal 2012). A substantial relationship between the type of vegetation and the GWL is also mentioned by Kučerová, Kučera, Hájek (2009). Vegetation is very responsive to changes of the water regime. A long-term decline or rise can cause succession changes. After longer periods without precipitation, sphagnum mosses become dry; the albedo changes cause decrease in evaporation and increase of surface temperature.

The GWL in many types of peat bogs has a considerable seasonal dynamics. A relatively stable level is found only in peat bogs saturated with source of artesian water. On the contrary, ombrotrophic bogs that are entirely dependant on precipitation, usually show high variability of GWL and considerable GWL decline during summer. The variation of the hydrological and hydro-chemical regime of peat bogs can also be demonstrated by an increased dynamics of soil temperatures in the acrotelm layer. In particular, it is subsequently reflected in the temperature of the water running off from the peat bog (Puranen, Mäkilä, Säävuori 1999).

In terms of the hydro-chemical properties of peat bog water, considerably negative correlations between the GWL and pH were identified; i.e. it was found that water deficiency in the catchment causes an increase in pH, although this research refers rather to river floodplains and transitional mires (Bufková, Prach 2006). A similar dependence between water quantity and pH, for the Šumava Mountains' streams supplied by peat bogs, was observed by Prokš (2010). The water in the Šumava Mountains' peatlands as well as precipitation have usually a low quantity of dissolved substances; the pH values are in particular determined by the dissolved organic carbon concentration, which has a strong seasonal progress related to evaporation and organic matter production. A higher content of organic matter along with low total mineralization naturally result in low pH (Kocum 2012). The interventions to the water regime of peat bogs can also be accompanied by considerable changes in hydrochemism, in particular due to changes in PO_4, Al, Fe concentrations and electric conductivity (Bufková, Stíbal 2012). These studies indicate a connection between the hydro-chemical properties and the quantity of water in the catchment; however, a comprehensive research on the mountain bogs of the Šumava Mts. to compare these water parameters in the catchment has not been carried out yet. Important aspects of the research on physico-chemical water properties concerning peat bogs are the measurements of changes in the electric conductivity and water pH in the stream. In general, the electric conductivity changes with the quantity of contained ions. In terms of cations, Mg, K, Na, but also Fe, Mn or Al occur the most frequently. In the case of very low pH, higher values of electric conductivity can be caused by free hydrogen anions (Worrall, Burt, Adamson 2006).

Studies on the Šumava Mts. comply with the general view of the world literature and prove that the development of peat bog complexes is controlled mainly by hydrological processes. The formation of peat bogs depends mainly on the water storage capacity, on water quantity, its origin and chemical properties. The main factor in the development of peat bog complexes is the GWL as it determines the composition of species of the site (Labadz et al. 2010). Thus, the depth of the GWL determines the balance between organic matter accumulation and decomposition, and therefore even minor changes in the water regime can have an essential effect on development of peat bog complexes (Holden, Chapman, Labadz 2004). Wilson et al. (2011) indicate a strong relationship between the type of vegetation and GWL stability in a peat bog. A higher vegetation cover also results in improvement of the self-regulating process of peat, as the evaporation is reduced due to the higher level of air moisture maintained.

Lots of studies also discuss the issue of the effect of drainage and peat extraction on the water regime and hydro-chemical water properties (Bufková, Stíbal

2012; Holden et al. 2011; Holden, Chapman, Labadz 2004). Anthropogenic disturbance of the water regime due to extraction or drainage can cause irreversible changes in the physical properties of peat (Evans et al. 1999). The GWL is mostly affected by the amount of precipitation, evapotranspiration, by topography and also, locally, the peat porosity and hydraulic conductivity (Wilson et al. 2011; Allott et al. 2009). Holden et al. (2011) point out that the long-term low GWL in drained peat bogs could lead to negative processes in peat. If the GWL is too deep below the surface, oxygen enters the lower layers of the peat bog, which results in peat decomposition. This process subsequently causes changes in the porosity, hydraulic conductivity and runoff characteristics (Allott et al. 2009; Joosten, Clarke 2002). The study by Wind-Mulder, Rochefort, Vitt (1996) concluded that peat bogs affected by extraction showed very low pH values and also a high variability of pH in comparison with intact peat bogs.

Ombrotrophic bogs saturated with rainwater have relatively stable hydro-chemical properties. However, this does not necessarily hold true for the whole bog, e.g. a lagg area (the zone of transition between an ombrotrophic bog and the mineral soils of the surrounding landscape) has variable hydro-chemical properties as it is affected by water from surrounding mineral soils. The influence of mineral water from the surroundings on a peaty soil depends also on its pH, electric conductivity and on the amount of dissolved calcium and carbonates (Howie, Meerveld 2011). Water from peat bogs is acid, and its pH usually ranges between 3.3–5.5, while transitional mires show higher pH ranging approximately from 4.5 to 6. However, the abovementioned values depend on several factors, e.g. climate, the height of the GWL or biological activity (Bergsma, Quinlan 2009; Holden, Chapman, Labadz 2004). Another important water parameter, regarding the study of peat bogs, is electric conductivity. Electric conductivity measurement in the field shows very variable values, from several units to hundreds $\mu S.m^{-1}$. The electric conductivity correlates linearly with the amount of substances dissolved in water. In addition, the conductivity is strongly affected by the soil temperature and peat properties such as the cation exchange capacity of the soil, organic components, peat porosity, pH and soil moisture (Ponziani et al. 2011). Another important parameter of water, which is often monitored in peat bogs, is the dissolved oxygen that is necessary for aerobic respiration at all trophic levels. Oxygen gets into water by means of diffusion from the atmosphere, water mixing and as a product of aquatic organisms. In terms of oxygen solubility, the water temperature is an important factor. In cold water, oxygen is dissolved more easily (O'Driscoll et al. 2016).

The main objectives of the study include:

a) Characteristics of dynamics of GWL and discharges of various types of peat bog sites.
b) Characteristics of physico-chemical properties of surface water and groundwater of various types of peat bog sites.

c) Assessment of statistical dependences of physico-chemical properties of surface water and groundwater in peat bog on the water quantity in the catchment.

2. Site description

The Mezilesní peat bog belongs to the geomorphological district of Kvildské Pláně. This peat bog is situated in the headwater area of the Hamerský brook, which is the right-side tributary to the Vydra river (Fig. 1). It is situated at an altitude of approximately 1,100 metres. The area of the whole peat bog complex is 83 ha large and the peat volume is 1,649 million m³ (Anděra, Zavřel 2003).

In terms of mean monthly discharges, the minimum discharges are reached in February (in the period prior the snow cover melts) or in September (in the late summer, dry period). The best water-rich period is in spring during snowmelt. In this period, peat bogs in the Šumava Mts. are usually fully saturated. In summer

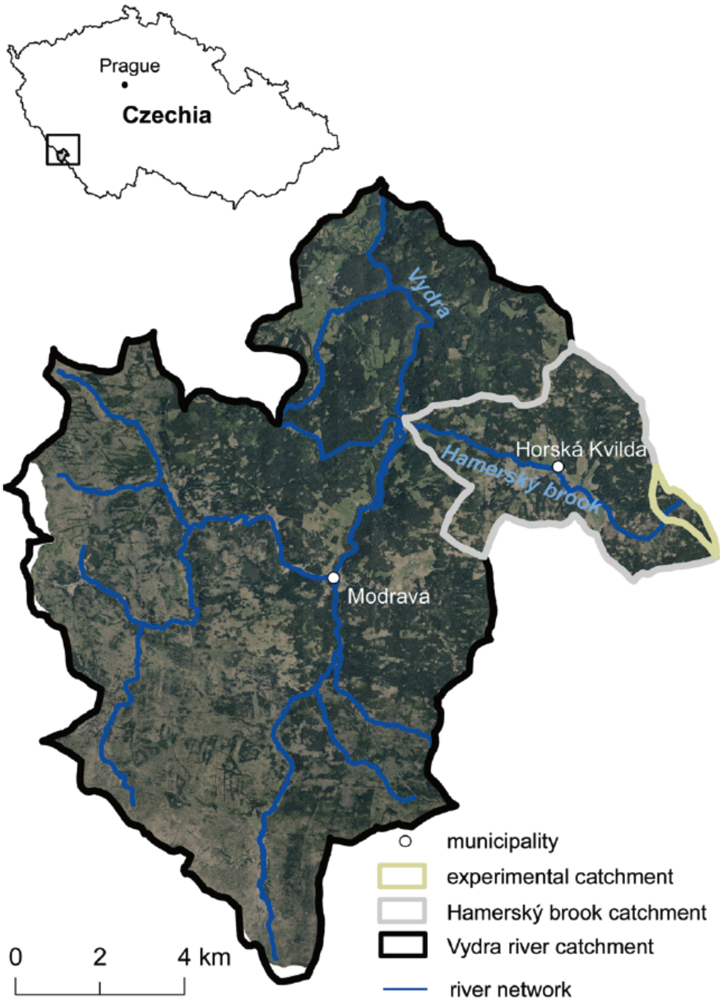


Fig. 1 – Catchment of the Vydra river with designation of the Hamerský brook and experimental catchment in its headwater area

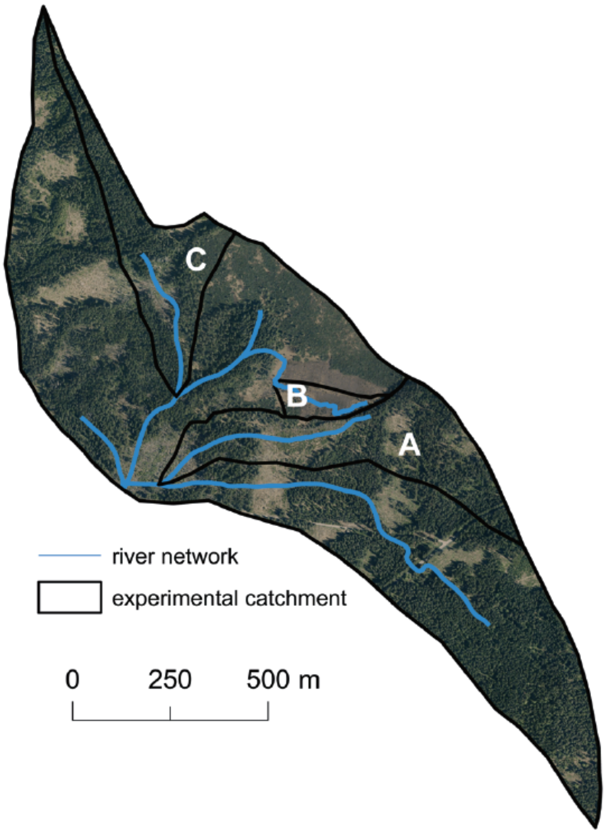


Fig. 2 – Headwater area of the Hamerský brook with designation of experimental catchments

months, discharges show a high variability, which is in particular caused by short periods of intensive precipitation. The value of mean annual precipitation in this region is approximately 1,100 mm (Kocum 2012). In the monitored area, there are peat bogs on sites with a low mean annual air temperature, but during the day the temperature can show considerable amplitudes (Hojdová, Hais, Pokorný 2005).

The headwater area of the Otava river basin belongs to a soil region of Cambisols or Rankers with steep slope areas and a region of entic and haplic Podzols. In the headwater area, there are predominantly hydromorphic soils situated in flat parts of the catchment (Šefrna 2004). The headwater area of the Hamerský brook, where the three experimental catchments are located (Fig. 2), is characterized by the occurrence of hydromorphic soils and organic soils which switch in the forest environment to entic Podzols and Gleysoils with acid pH reaction. The reason for selection of these catchments was due to the fact that on a relatively small area within one peat bog complex, there are several minor catchments with different vegetation cover. The different vegetation composition could have an essential effect on the hydrological regime of the experimental sites due to the difference in evapotranspiration. There is also a large area without vegetation, with bare peat only.

The three experimental catchments cover a total area of 25.4 ha. The smallest basin is an extracted part of the peat bog (catchment B), which is formed mostly by bare peat (Table 1). Shrub is here represented by stands of bog-bilberry (*Vaccinium*

Tab. 1 – Basic vegetation characteristics of experimental catchments

Catchment	A	B	C
Catchment area (ha)	13.6	1.9	9.9
Stream length (m)	599.5	291.8	401.3
Watershed length (km)	2.3	0.7	2.4
Maximum altitude (m a.s.l.)	1,149	1,113	1,135
Minimum altitude (m a.s.l.)	1,092	1,100	1,095
Elevation difference (m)	57	13	40
Forest proportion (%)	67	1	78
Shrub and grass proportion (%)	33	43	22
Proportion of area without vegetation (%)	0	56	0

uliginosum), common heather (*Calluna vulgaris*) and in some places, where peat was disturbed by extraction, there are areas of tussock cottongrass (*Eriophorum vaginatum*) and various species of sedges (*Carex;* Křenová, Hruška 2012). Catchment A (Fig. 2) includes a lagg that is represented by shrubs of the same species as catchment B. In addition, it is enriched with stands of cowberry (*Vaccinium vitis-idaea*), common bilberry (*Vaccinium myrtillus*), bog cranberry (*Vaccinium oxycoccus*) *and various species of grasses*. The lagg in the upper parts, at a distance from the peat bog, passes gradually to waterlogged forest, formed mainly by Norway spruce (*Picea abies*). Catchment C with predominant waterlogged forest is also formed mainly by Norway spruce (*Picea abies*). A part of the catchment is represented by mountain bog with predominant dwarf mountain pine (*Pinus mugo*). The treeless places are mostly formed by shrubs or various species of sedges (*Carex*) and cranberries (*Vaccinium*).

3. Methodology

All measurements were performed during the growing season of 5th May–29th September 2018. Twenty-one field measurements were performed at weekly intervals. Campaigns were held in the period of high water saturation of the soil during the spring after snowmelt and also during a very dry period of hot summer respectively. Each experimental stream was fitted with a hydrometric profile and four tubes at a distance of 2 m and 5 m from the stream on both sides, in which the physico-chemical properties of water and the GWL were measured manually. All GWL values were measured as distance between surface and water level in the tube. Measuring points were selected so as to sufficiently represent each type of vegetation (Fig. 3).

Tube 1 is situated at a distance of 2 m from the left bank, tube 2 at a distance of 5 m from the left bank, tube 3 at a distance of 2 m from the right bank and tube 4

Fig. 3 – Vegetation cover in three experimental catchments. Catchment A – a lagg formed by grass, catchment B – an extracted part - bare peat, catchment C – waterlogged spruce forest. The experimental catchment and the streams are designated in the paper with letters A, B, C and the tubes are designated with numbers 1–4.

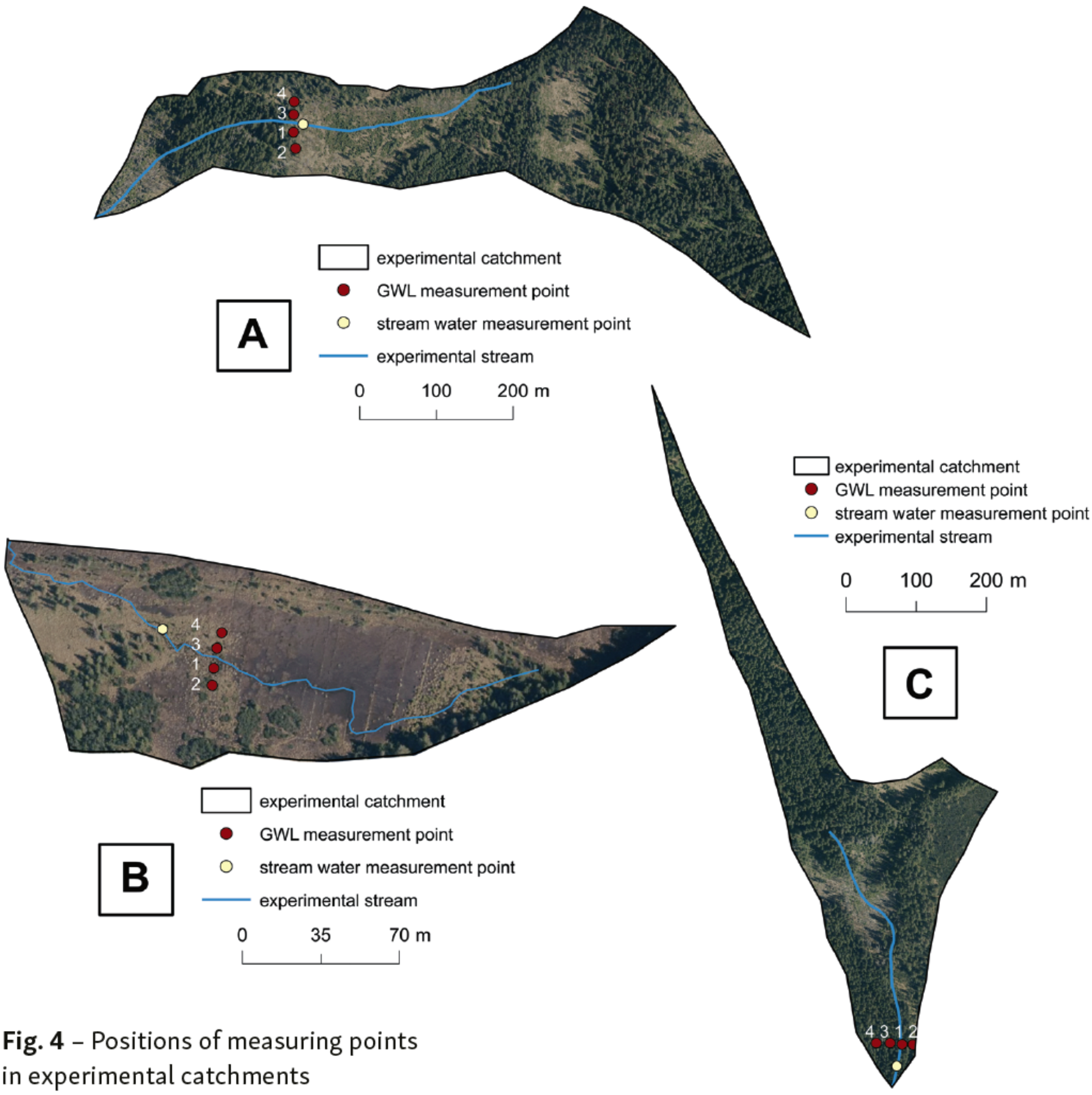


Fig. 4 – Positions of measuring points in experimental catchments

at a distance of 5 m from the right bank. This composition is identical in all three experimental catchments. The tubes and hydrometric profiles were installed as close as possible to the stream outflow from catchment so as to represent each type of vegetation as much as possible. Nevertheless, e.g. for catchment B, this was not possible due to surface factors, as in the lower part of the catchment, there are many minor depressions and the stream branches off to several small erosion furrows, terminating either in these depressions, or in a bog lake. In this case, an alternative site was selected upstream where it was possible to determine unambiguously the stream. In terms of catchment A, a small displacement was caused by a considerable change in vegetation cover in the catchment outflow (Fig. 4).

The measurements were performed episodically, using hand-held equipment. In this research, following parameters were observed: discharge, GWL, pH, electric conductivity, dissolved oxygen and water temperature. The parameters were measured in experimental streams (surface water) and experimental tubes (groundwater). The measurements of physical and chemical parameters of water were assured by calibrated field measurement systems, measuring within the range pH: 0–14, electric conductivity: 0–20,000 µS/cm, temperature: −35 °C to +135 °C and dissolved oxygen: 0–20 mg/l.

Discharges were measured by transmissible Poncelet weirs, which were calculated and adjusted according to parameters of monitored streams in growing season 2018. Discharges were calculated by Basin weir equation (Equation 1), using spillway-discharge coefficient for Poncelet weir (Equation 2):

$$Q = \mathrm{m.b.}\sqrt{2.g}.h^{\frac{3}{2}} \tag{1}$$

$$m = \left[0.405 + \frac{0.003}{h} - 0.03\left(1 - \frac{b}{B}\right)\right][1 + 0.55)(\tfrac{b}{B})^2(\tfrac{S_0}{S})^2] \tag{2}$$

where m represents the spillway-discharge coefficient for Poncelet weir, b the length of the spillway crest, g the acceleration of gravity, h the falling height of water, B the length of weir, S_0 stream flow profile area, S area of spillway crest (Šráček, Kuchovský 2003). All data were subsequently analysed, using the statistical software StatSoft. Possible correlations between specific discharges, GWL and physico-chemical parameters were identified by calculations of Pearson correlation coefficients at the level $p < 0.05$.

The study also evaluates the dynamics of changes in the groundwater level in relation to precipitation and daily potential evapotranspiration (PET). The potential evapotranspiration was calculated by the Penman–Monteith equation (Equation 3).

$$PET_0 = \frac{0.408.\Delta.(Rn - G) + \gamma.\frac{900}{T+273.16}.u.(e_s - e_a)}{\Delta + \gamma.(1 + 0.34.u)} \tag{3}$$

where Δ represents the inclination of the water vapour saturation curve in connection with temperature [kPa.°C^{-1}], Rn the radiation balance [MJ.m^{-2}.day^{-1}], G the flow of heat into the soil [MJ.m^{-2}.day^{-1}], Υ a psychometric constant [kPa.°C^{-1}], u the speed of wind [m.s^{-1}], $(e_s$-$e_a)$ the saturation deficit of air at elevation z [kPa], and T the average air temperature [°C] (Penman 1948).

There is no meteorological station situated in experimental site. All meteorological data used in this paper were measured at the nearest automatic meteorological station at Modrava village, which is 8 km in a straight line.

4. Results

4.1. Water regime

Boxplots indicate (Fig. 5) that the widest range of specific discharges was observed for the stream representing an extracted part of peat bog (stream B). The stream B was the only one to show outliers and, in addition, it was completely dried out twice during the measuring period. High temperatures and direct solar radiation resulted in rapid peat drying. The most stable specific discharges were identified at the stream in waterlogged forest (stream C), which could be caused by high interception and lower evapotranspiration in comparison to other types of vegetation. The lagg (stream A) with its discharge values is rather similar to the extracted

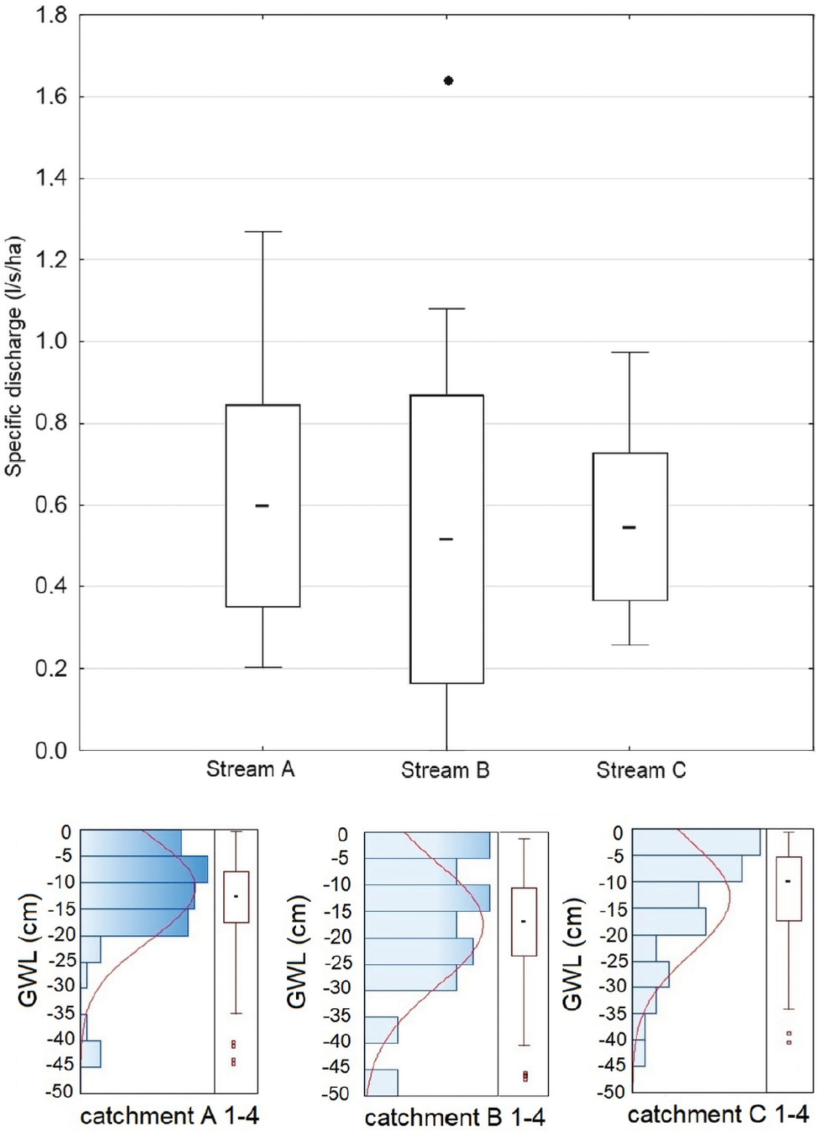


Fig. 5 – Distribution of measured values of discharges and GWL: stream (catchment) A – lagg, stream (catchment) B – extracted part of peat bog, stream (catchment) C – waterlogged forest. Numbers 1–4 designate the aggregate value of all measuring points of the GWL within the experimental catchment.

Tab. 2 – The basic statistical characteristics of discharges and GWL. Stream (catchment) A – lagg, stream (catchment) B – extracted part of peat bog, stream (catchment) C – waterlogged forest. Numbers 1–4 designate the aggregated value of all measuring points of the GWL within the experimental catchment.

Stream	A	B	C	Catchment	A 1–4	B 1–4	C 1–4
Number of measured values	21	21	21	Number of measured values	84	84	84
Average specific discharge (l/s/ha)	0.6	0.51	0.55	Average GWL (cm)	10.97	17.40	12.96
Minimum specific discharge (l/s/ha)	0.2	0	0.26	Minimum GWL (cm)	0	0.50	0
Maximum specific discharge (l/s/ha)	1.27	1.65	0.98	Maximum GWL (cm)	45.00	49.10	42.20
Range specific discharge (l/s/ha)	1.07	1.65	0.72	Range GWL (cm)	45.00	48.60	42.20
Standard deviation	0.25	0.35	0.18	Standard deviation	9.76	11.82	10.45
Coefficient of variance	41	68	33	Coefficient of variance	89	68	81

part; nevertheless, it does not show such considerable extremes. The quantity of measured GWL values in particular intervals was very similar in the forest and in the lagg. The extracted part showed a higher variance in the GWL.

The basic statistical parameters show that the highest variation of discharges was observed in the extracted part of the peat bog, which is confirmed by the standard deviation of flows and the variance (Table 2). In addition, high values of standard deviation and a low coefficient of variance of the GWL indicate the occurrence of considerable variations in the GWL. In this case it was proved that waterlogged forest does not have high variability of flows in comparison with other types of peat bog sites, because the range of measured values and their

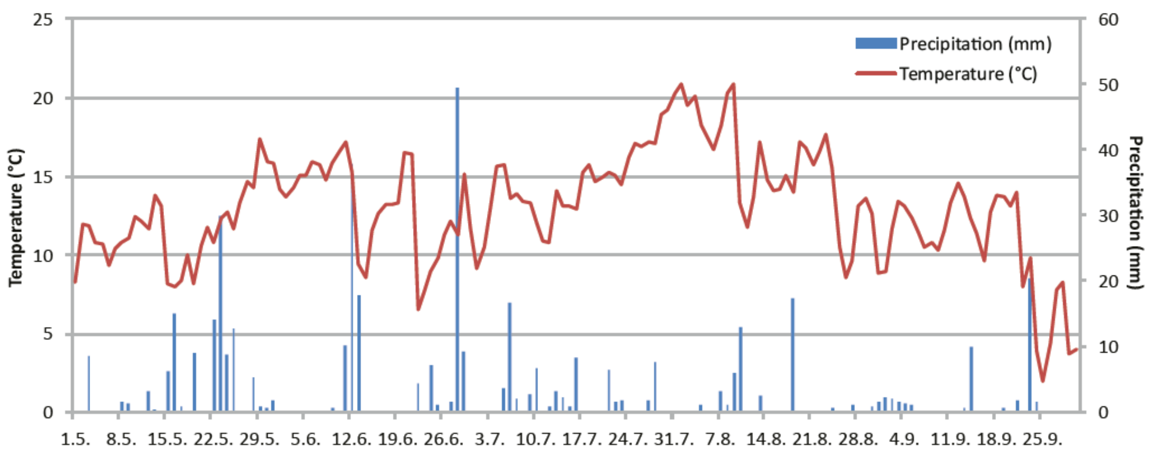


Fig. 6 – Regime of daily precipitation and mean daily air temperature in observed period

standard deviations are very low. In terms of the variability of values of the specific discharges and GWL, the lagg is situated in the middle of both mentioned types.

For a clear illustration of progress of GWL, precipitaton and air temperature in observed period were analysed (Fig. 6). During the summer there were high mean daily air temperatures, especially in late July and early August (over 20 °C). The total of precipitation in observed period was 422.1 mm, which represented approximately average growing season in the region. Maximum of daily precipitation was 49.5 mm (26th June). The highest monthly value of precipitation was in June (140 mm) and the lowest in September (43.8 mm). It is assumed that at the beginning of the growing season peat bog is fully saturated due to snowmelt.

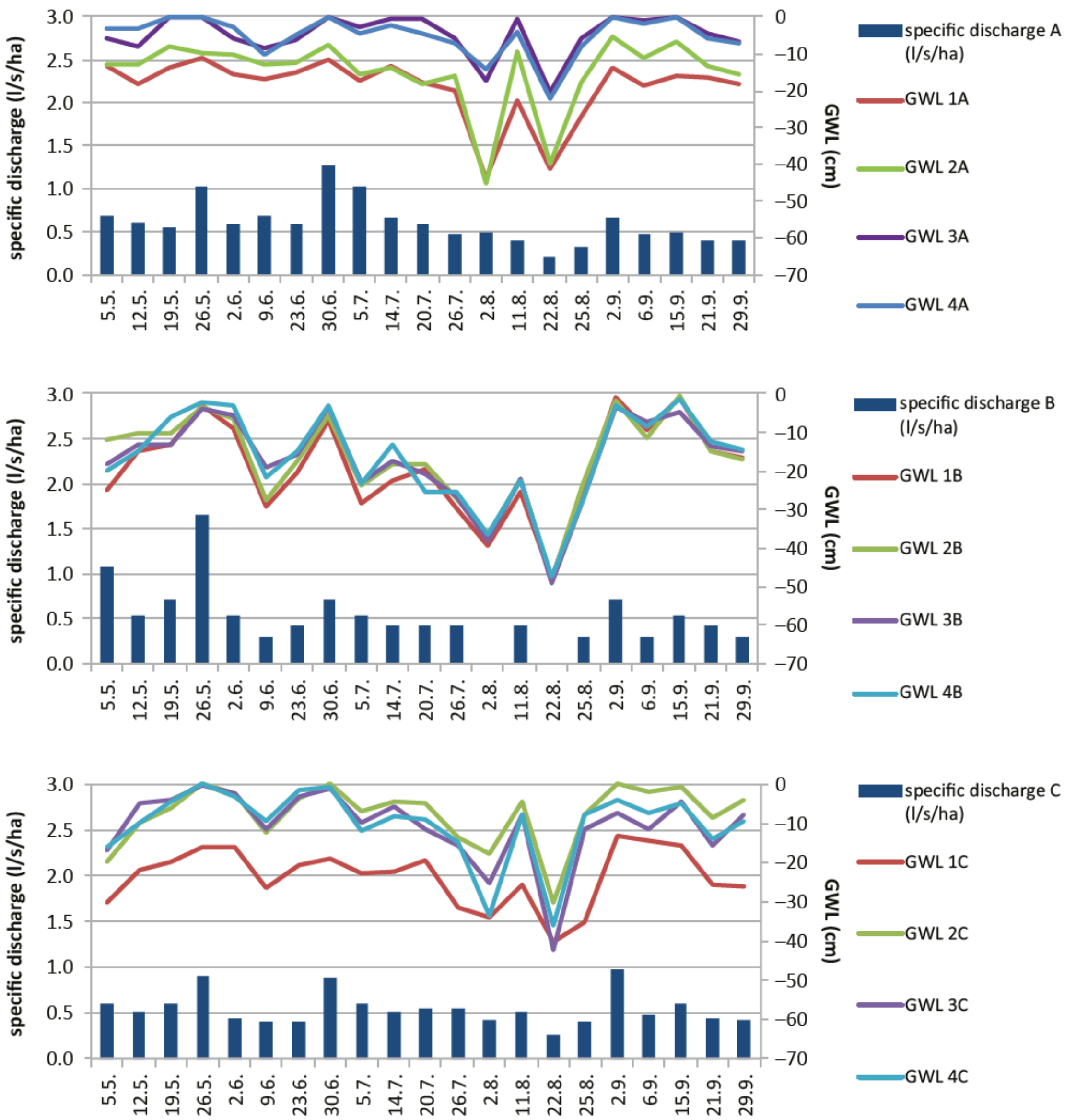


Fig. 7 – Specific discharges and GWL progress in experimental catchments

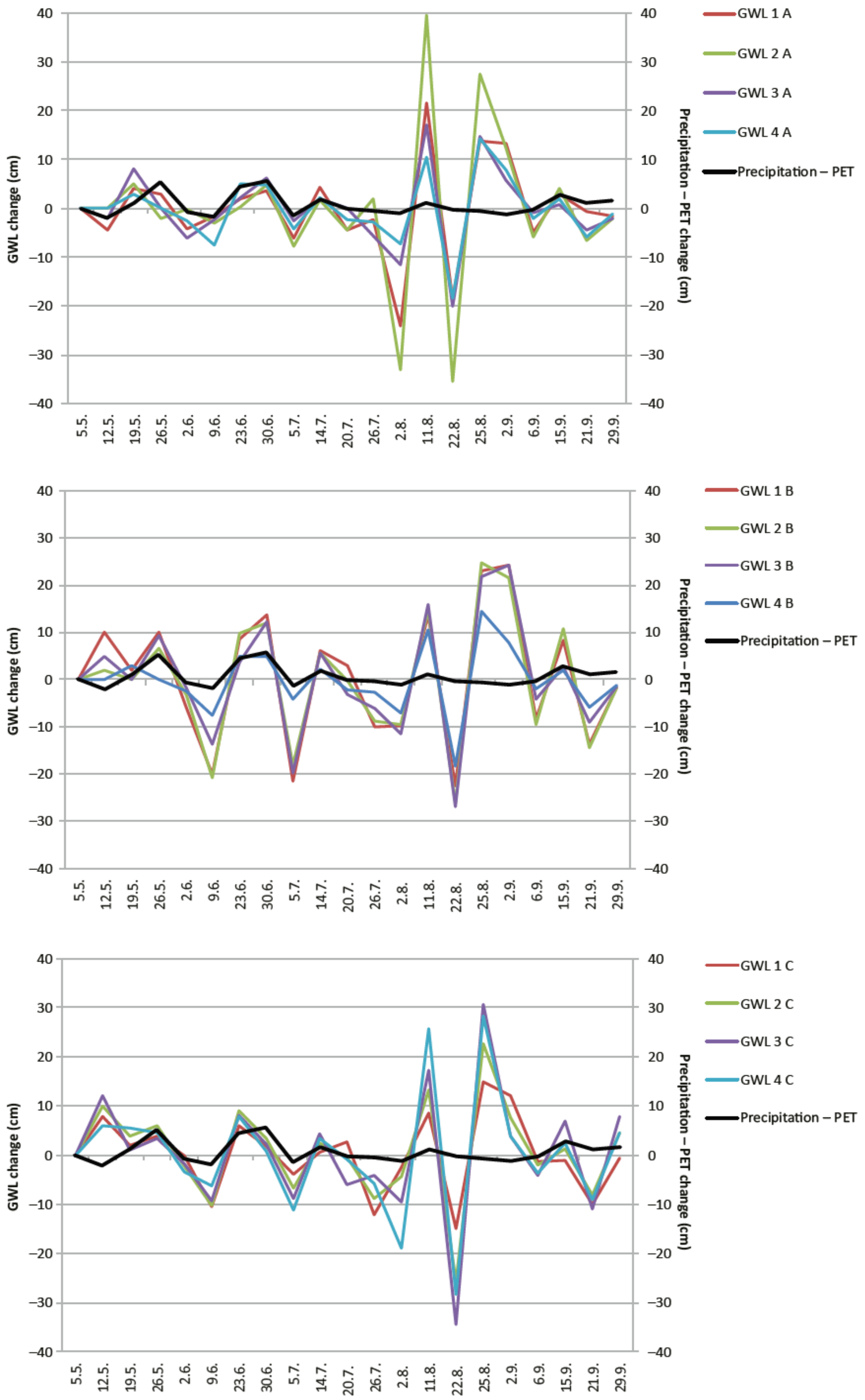


Fig. 8 – Differences in daily precipitation and PET and their comparison with differences in GWL between measurements within experimental catchments

The progress of the specific discharges and GWL within experimental catchments is shown in Figure 7. The charts show a higher variability of discharges in the extracted part of the peat bog, which probably relates to higher variability of the GWL. This fact also indicates that the distance of GWL measurement from the stream is not so important, because all measuring points of experimental catchment B show almost identical values during the whole measuring period. Stream C, in waterlogged forest, shows relatively high values of GWL (as in the case of the lagg). However, the values of the specific discharges are slightly lower, which is probably caused by the considerably more dense vegetation cover and thus by better surface cover of the peat bog complex. The importance of vegetation cover was probably also confirmed by the lower variance of extreme values in comparison with the lagg, both in the maximum and in the minimum flow.

The dynamics of changes in GWL depends mostly on meteorological factors. The differences of the precipitation and daily PET between campaigns in all catchments are very similar to the progress of changes in the GWL (Fig. 8). Catchment A shows the smallest average deviation of the difference in GWL and differences of the precipitation and PET. In this case, the average deviation is 6.4 cm, i.e. the other factors affecting the GWL do not probably have such a strong effect. In terms of catchment B, the average deviation reaches the maximum value of 8.3 cm, which indicates rapid changes in the GWL in the disturbed area of the peat bog. For the waterlogged forest, the average deviation reached the value of 7 cm. In addition, it is necessary to point out that the amount of deviations significantly increased by the four measurements in August, with very high deviations from the average in all catchments.

4.2. Physico-chemical water properties

The extracted part of the peat bog, which surface is not covered by any vegetation, shows the highest mean temperature and variability of the surface water temperature and groundwater temperature. In this case, the vegetation of lagg and waterlogged forest contribute identically to a decrease in temperature that is similar in both types of peat bog (Table 3). The values of dissolved oxygen in all catchments are very similar. The amount of oxygen reaches slightly higher values only in the groundwater of the waterlogged forest. The stream in the extracted part of the peat bog shows a very high electric conductivity. On the other hand, the comparison of values of the groundwater electric conductivity showed that similar values were measured in all experimental catchments. In the observed catchments, the pH reached highest values in the lagg, probably due to soil mineralization at the edge of the peat bog. The waterlogged forest indicated slightly lower values, and a considerably acid environment was identified in the extracted

Tab. 3 – Physical and chemical water properties in observed experimental streams and in groundwater. GWL 1–4 designate aggregate value of all groundwater measurement points within experimental catchment. The stream designates the aggregate value just in the respective experimental stream.

Catchment A	Number of measurements	Mean	Standard deviation
Water temperature (°C) stream	21	10.70	2.25
Dissolved oxygen (mg/l) stream	21	18.80	3.07
Electric conductivity (µS/cm) stream	21	83.76	68.55
Stream water pH	21	5.82	0.76
Water temperature (°C) GWL 1-4	84	11.51	1.88
Dissolved oxygen (mg/l) GWL 1-4	84	16.75	4.12
Electric conductivity (µS/cm) GWL 1-4	84	39.64	8.52
pH 1-4 GWL	84	4.27	0.64
Catchment B	**Number of measurements**	**Mean**	**Standard deviation**
Water temperature (°C) stream	19	17.85	4.54
Dissolved oxygen (mg/l) stream	19	18.91	1.93
Electric conductivity (µS/cm) stream	19	146.47	86.37
Stream water pH	19	3.95	0.41
Water temperature (°C) GWL 1-4	84	14.47	1.91
Dissolved oxygen (mg/l) GWL 1-4	84	16.65	3.36
Electric conductivity (µS/cm) GWL 1-4	84	60.98	17.84
pH 1-4 GWL	84	3.68	0.28
Catchment C	**Number of measurements**	**Mean**	**Standard deviation**
Water temperature (°C) stream	21	10.02	1.79
Dissolved oxygen (mg/l) stream	21	18.92	2.17
Electric conductivity (µS/cm) stream	21	89.43	61.49
Stream water pH	21	4.77	0.79
Water temperature (°C) GWL 1-4	84	11.67	2.09
Dissolved oxygen (mg/l) GWL 1-4	84	18.23	2.83
Electric conductivity (µS/cm) GWL 1-4	84	57.52	19.23
pH 1-4 GWL	84	4.12	0.60

part of the peat bog where there is probably rapid rainout of organic substances, causing a low pH. In addition, there are low values of standard deviations, i.e. the pH is constantly low for the whole measuring period.

The identification of the measure of interdependence between observed parameters was calculated by means of Pearson correlation coefficients that were always assessed separately within particular measurement points. Mutual comparison of all the parameters resulted in a total of 36 correlations.

For a better illustration, the measurement points were added to the table within each catchment where correlation of parameters was identified (Table 4). The measured parameters of GWL and discharge were connected in the category "water quantity". Strong mutual relationships were identified in case of the catchment B, where all measurement points within the catchment showed a correlation between electric conductivity and the amount of dissolved oxygen. Four of five

Tab. 4 – Number of correlations of observed parameters at measurement points within experimental catchments

Catchment A	Water quantity	Water temperature (°C)	Dissolved oxygen (mg/l)	Electric conductivity (µS/cm)	pH
Water quantity		1	1		3
Water temperature (°C)	1			1	
Dissolved oxygen (mg/l)	1			1	2
Electric conductivity (µS/cm)		1	1		1
pH	3		2	1	
Catchment B	**Water quantity**	**Water temperature (°C)**	**Dissolved oxygen (mg/l)**	**Electric conductivity (µS/cm)**	**pH**
Water quantity		4	1		
Water temperature (°C)	4			1	1
Dissolved oxygen (mg/l)	1			5	2
Electric conductivity (µS/cm)		1	5		3
pH		1	2	3	
Catchment C	**Water quantity**	**Water temperature (°C)**	**Dissolved oxygen (mg/l)**	**Electric conductivity (µS/cm)**	**pH**
Water quantity		1		1	2
Water temperature (°C)	1		1	1	
Dissolved oxygen (mg/l)		1		1	1
Electric conductivity (µS/cm)	1	1	1		1
pH	2		1	1	

measurement points showed a mutual relationship between the water quantity and its temperature. Moreover, considerable interdependences were evident between the amount of dissolved oxygen and pH and between electric conductivity and pH. On the contrary, the smallest correlation was identified between the water quantity and pH at two measurement points in the area of waterlogged spruce forest. The same correlation was also identified at three measurement points of lagg. There was also perceived a considerable correlation between the pH and the amount of dissolved oxygen.

Figure 9 shows the calculated correlations of particular parameters within the experimental catchments, identified at multiple measurement points. In terms of catchment A, there were three measurement points with correlation between pH and water quantity (specific discharge and GWL) and two measurement points with correlation between pH and the amount of dissolved oxygen. The progress of the linear functions indicates that the pH in the lagg increased with a declining water quantity in the peat and stream. In the case of groundwater, higher peat aeration probably results in an increase of pH. In two cases it was identified that an increase of pH results in an increase of the amount of dissolved oxygen in water.

At the extracted part of the peat bog, a correlation between the GWL and water temperature (Fig. 10) was identified in four cases. It showed that low GWL meant warmer groundwater. This probably relates to the fact that during sunny days, dark and exposed bare peat absorbs a lot of heat, resulting in a rapid decrease of the GWL and an increase of water temperature. Another reason could be the changes in the heat conductivity of peat after its drying. In addition, a connection was proved also between the dissolved oxygen and electric conductivity. An increasing

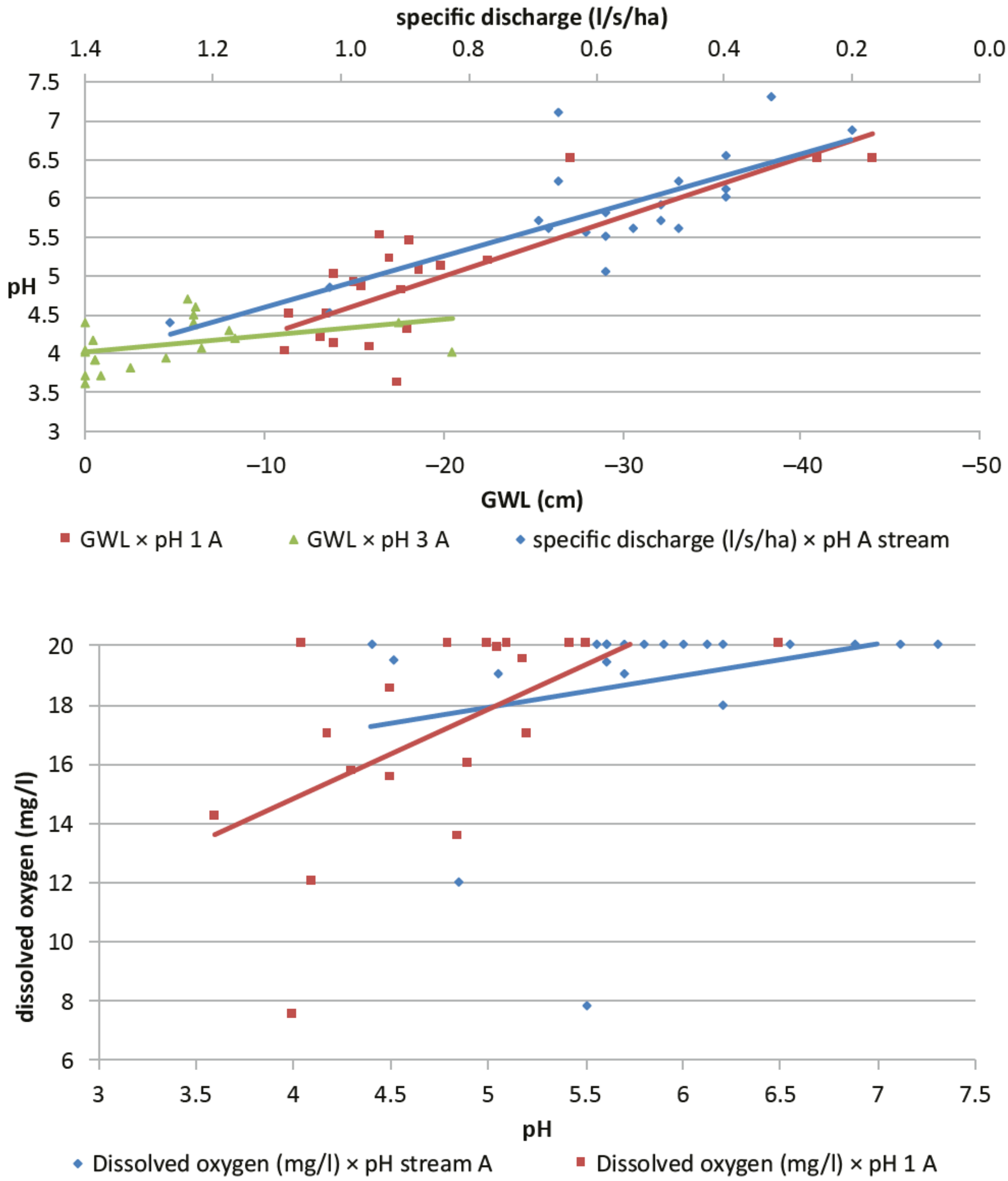
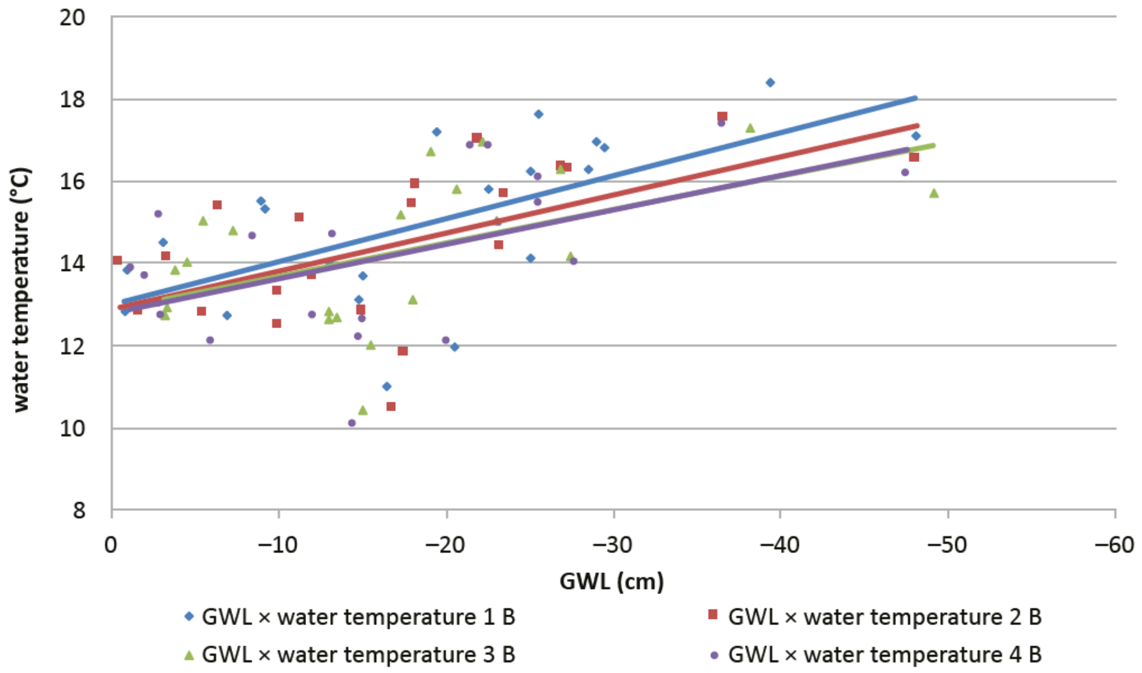


Fig. 9 – Graphic representation of correlations that were identified more than once in experimental catchment A

electric conductivity resulted in this case in higher dissolved oxygen concentration at all measurement points. Just as in the case of the lagg, there is a situation when an increase in pH occurs at lower GWL. Three measurement points were identified where the water pH decreased while the electric conductivity increased.



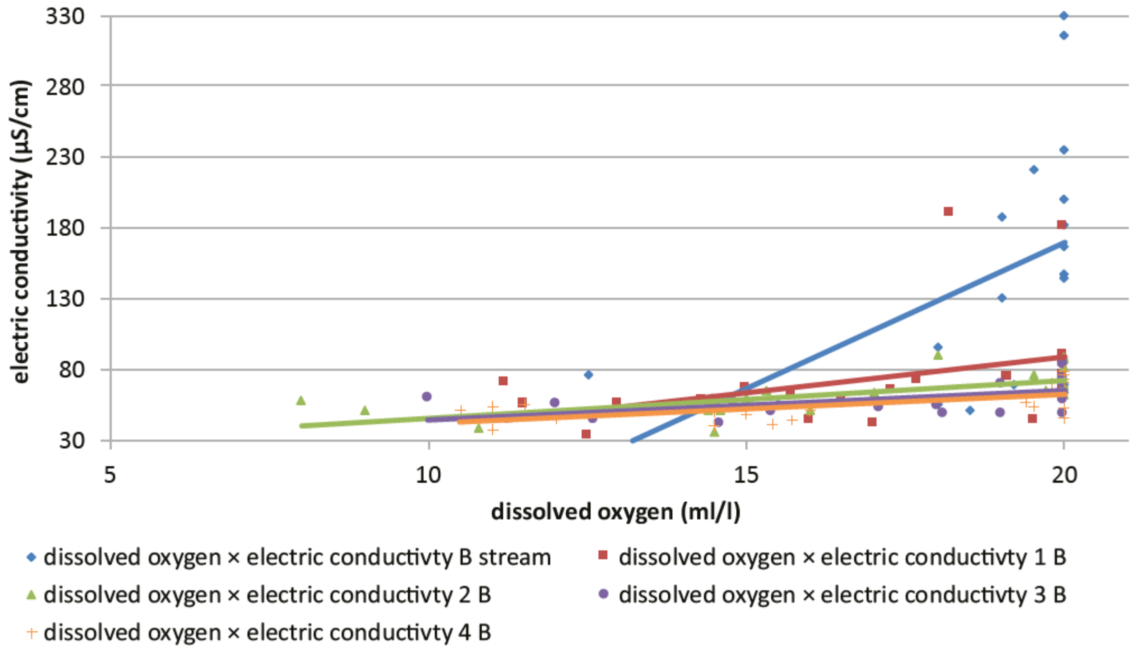
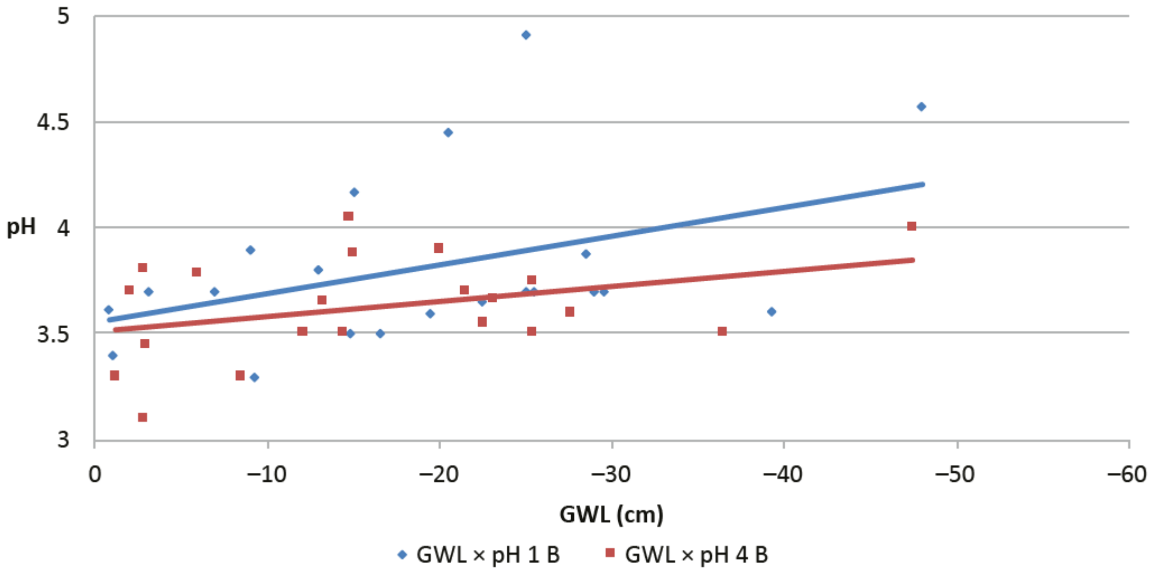


Fig. 10a – Graphic representation of correlations that were identified more than once in experimental catchment B



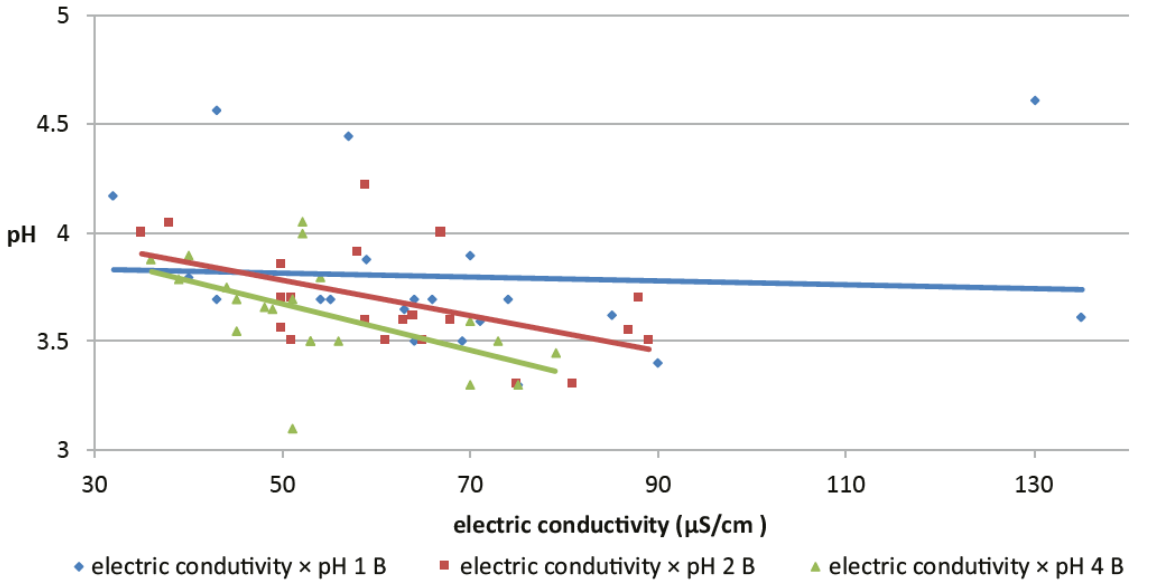


Fig. 10b – Graphic representation of correlations that were identified more than once in experimental catchment B

The catchment in the waterlogged forest showed only one correlation between measured parameters. It was a case of the stream and the nearest GWL point (Fig. 11). As in the previous two experimental catchments, this is the trend when a decreasing quantity of flows or the GWL results in an increase of pH.

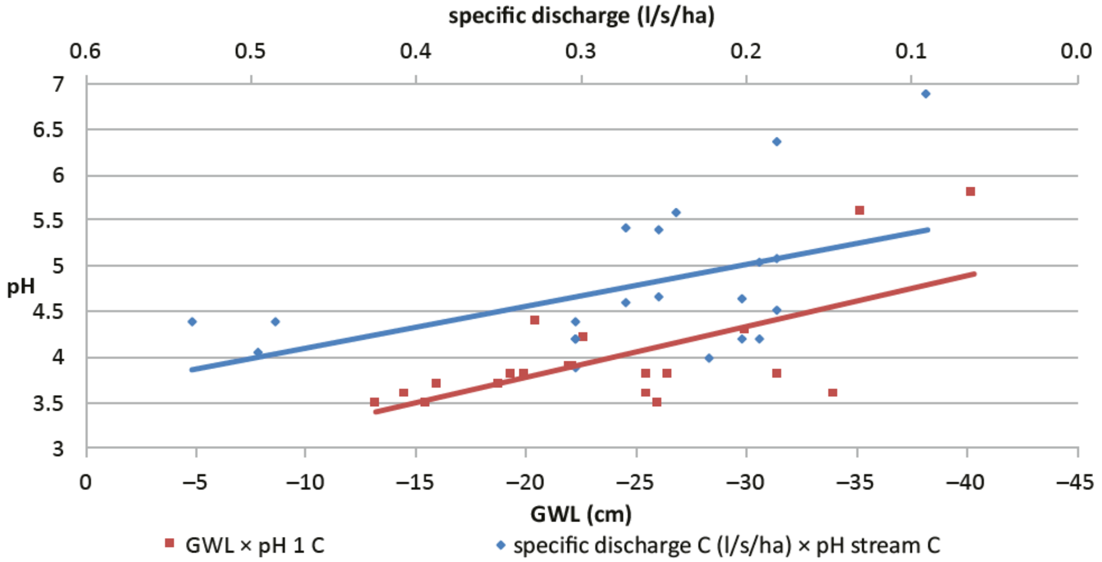


Fig. 11 – Graphic representation of correlations that were identified more than once in experimental catchment C

5. Discussion

The latest studies describing hydrological regime of the headwater area of the Otava river basin indicate that the occurrence of peat bogs in the catchment could be at the origin of higher variability of runoff. This claim was proved by an analysis that showed a very strong dependence of the flow extremity on the proportion of peaty area in the catchment (Čurda, Janský, Kocum 2011). Vlček (2017) also observed that the period of maximum saturation of peat bogs resulted in a rapid rise of runoffs. Furthermore, during the dry periods, the peat bogs store water which means that they do not supply streams. In terms of experimental catchments, it was shown that in observed growing season, the highest variability of the discharges and GWL was identified in the extracted part of the peat bog. The extracted part also reacted negatively during hot summer days when the stream was completely dried up. The most probable cause could be the absence of vegetation. Similar results in anthropogenically disturbed parts of the peat bog were also reported by Holden et al. (2011). In addition, Holden, Chapman, Labadz (2004) point out that even minor changes in the water regime can alter the balance between the decomposition and accumulation of organic matter, which can negatively affect peat bog ecosystems. Statistical characteristics indicated similar progress of the GWL and variability of the discharge in case of lagg and waterlogged forest. Wilson et al. (2011) attribute this behaviour to the buffer effect of vegetation. Trees and shrubs could probably decrease the variability of discharge and also equilibrate a higher and more stable GWL. A similar connection between vegetation cover and the stability and height of the GWL is also described by Labadz et al. (2010) and, in terms of the Šumava Mts., by Bufková, Stíbal (2012) and Kučerová, Kučera, Hájek (2009). In addition, the studies suggest that GWL is especially controlled by the difference between precipitation and evapotranspiration, which was described by Wilson et al. (2011), Allott et al. (2009), and in terms of the Šumava Mts. by Doležal et al. (2017).

The temperature conditions of the surface water and groundwater in observed area are affected by the amount of incoming solar radiation and by the vegetation pattern. High temperatures of water were observed mainly in the extracted part, where there is only bare peat. On the contrary, both the lagg and waterlogged forest showed almost identical temperature conditions. However, Puranen, Mäkilä, Säävuori (1999) point out that temperature variability in acrotelm varies very quickly and considerable differences can be measured within the peat bog. Kučerová, Kučera, Hájek (2009) state that a vegetation effect is identified particularly in the summer during dry seasons when high temperatures result in considerable changes in the albedo and evapotranspiration in the peat bog.

A very low pH was identified in the extracted part of the peat bog. This was also observed in the study by Wind-Mulder, Rochefort, Vitt (1996). A low pH and its higher variability can be explained by a more rapid release of acid organic substances from peat during more intense precipitation events. Lagg showed in this case a relatively high pH. However, Howie, Meerveld (2011) point out that in this regard the lagg can be very variable, as it can be affected by influent mineralized water present in the surroundings of the peat bog. In observed season, the pH values in the groundwater of the waterlogged forest did not differ that much from the pH values of the lagg, but the pH value of the stream was on average lower by almost 1.1. A very strong negative correlation was identified between pH and water quantity in all observed catchments. In a dry period, the water in peat showed higher pH, which can also be attributed to the rainout of organic substances from the peat, which ceases in dry periods. In addition, the similar results in terms of the Šumava Mts. were observed by Prokš (2010).

In this paper, a connection between low pH and higher values of electric conductivity was observed in case of the extracted part. The amount of dissolved substances and the pH can determine the mountain bog pattern and related processes. Similar process was observed in Ponziani et al. (2011). The paper by Wind-Mulder, Rochefort, Vitt (1996) also indicated considerably higher values of electric conductivity in comparison of anthropogenically disturbed parts of peat bog with its intact parts. An interesting finding is that the stream in the waterlogged forest has on average almost identical progress of electric conductivity as the lagg, but it has considerably higher values in terms of groundwater, where they are comparable with the values in the extracted part of the peat bog.

In comparison of the values of dissolved oxygen, it was identified that the three experimental catchments are comparable in this regard. A slightly higher concentration was only identified for the groundwater of the waterlogged forest. O'Driscoll et al. (2016) mentioned a significant temperature effect on the amount of dissolved oxygen. However, these correlations in the case of the experimental catchments were not identified. In terms of dependences, a considerable relationship between dissolved oxygen and electric conductivity was identified, but only in the extracted part of the peat bog. In terms of the lagg, the dependence of dissolved oxygen on pH was identified. However, both of these correlations may be affected by specific physical-geographical factors. For confirmation of these claims, longer data series, which could affect variability of precipitation, is needed. Results were based on experimental measurements in only one growing season and compared to results stated in similar short-term studies of peat bogs. But there are many factors influencing physico-chemical properties of water and hydrological regime. This paper showed differences in progress of observed parameters in various types of peat bog and it was compared to results of related studies. However, all the results from this paper need to be confirmed by long-term measurements.

6. Conclusion

The main objective of this study was to describe differences in behaviour of surface water and groundwater at three peat bog sites with different vegetation covers. Considerable differences were found at the disturbed site. Drainage and peat extraction resulted in low GWL, rapid dynamics of changes in the GWL and discharges. The absence of vegetation in such site indicates a non-standard hydrological regime and its negative effects in comparison with lagg or waterlogged forest. Both of these experimental catchments showed relatively stable discharges and GWL. Another important aim of the study was to describe the basic physico-chemical properties of water in various types of peat bogs. In comparison with intact parts of the peat bog, different water properties were observed in the parts disturbed by extraction. There were considerable declines of the GWL and low specific discharges identified, followed by higher water temperature, unlike in the case of the lagg and waterlogged forest. This aspect contributes to changes in the remaining physical and chemical water properties. In the disturbed part, there were also very low pH values measured. It is probably caused by the absence of vegetation, which allows intensive rainout of organic substances. Moreover, higher values of electric conductivity appeared at the disturbed site. On the contrary, the lagg shows high pH values, which indicate a gradual soil mineralization towards the edge of the peat bog. In terms of other observed parameters, both the lagg and waterlogged forest showed comparable values. The last aspect of the paper was to identify the dependencies of physical and chemical parameters on the quantity of water stored in the catchment. A strong correlation was observed between the water quantity and pH. Precipitation is probably followed by a rapid rainout of organic substances, which caused a decrease in pH at all monitored catchments. In addition, the dependence between a low pH and electric conductivity was identified. This dependence is probably caused by the large amount of free hydrogen anions in the period when the peat bog is suffering from water deficiency. The results of the study are mostly in compliance with the other discussed studies. Nevertheless, this study, based on the observation of a unique vegetation season, could contribute to the general knowledge of these specific sites. In the context of increasing frequency of hydrological extremes, it appears appropriate to continue studying hydrological regime at sites with high retention potential. To understand processes ongoing in various types of peat bogs, continual observation of dynamics of groundwater level and runoff is essential. It is also necessary to focus on those parts of peat bogs that are the most involved in rainfall runoff processes; for this purpose, natural tracers and other modern methods can be used.

References

ALLOTT, T.E.H., EVANS, M.G., LINDSAY, J.B., AGNEW, C.T., FREER, J.E., JONES, A., PARNELL, M. (2009): Water tables in Peak District blanket peatlands. Moors for the future. Edale, Derbyshire.

ANDĚRA, M., ZAVŘEL, P. (2003): Šumava – příroda, historie, život. Baset, Praha.

BERGSMA, B., QUINLAN, C. (2009): Sifton Bog ESA – Conservation Master Plan 2009–2019. Upper Thames River Conservation Authority, Parks, London.

BUFKOVÁ, I., PRACH, K. (2006): Linking vegetation pattern to hydrology and hydrochemistry in a montane river floodplain, the Šumava National Park, Central Europe. Wetlands Ecology and Management, 14, 4, 317–327.

BUFKOVÁ, I., STÍBAL, F. (2012): Restoration of drained mires in the Šumava National Park. In: Jongepierová, I., Pešout, P., Jongepier, J.W., Prach, K. (ed.): Ecological restoration in the Czech Republic, Nature Conservation Agency of the Czech Republic, Prague, 78–80.

ČURDA, J., JANSKÝ, B., KOCUM, J. (2011): Vliv fyzicko-geografických faktorů na extremitu povodní v povodí Vydry. Geografie, 116, 3, 335–353.

DOLEŽAL, T., VLČEK, L., KOCUM, J., JANSKÝ, B. (2017): Evaluation of the influence of mountain peat bogs restoration measures on the water table level: case study Rokytka peat bog, the Šumava Mts., Czech Republic. Acta Universitatis Carolinae. Geographica. Univerzita Karlova, 52, 2, 1–10.

EVANS, M. G., BURT, T. P., HOLDEN, J., ADAMSON, J.K. (1999): Runoff generation and water table variations in blanket peat: evidence from UK data spanning the dry summer of 1995. Journal of Hydrology, 3–4, 221, 141–160.

HOLDEN, J., CHAPMAN, P.J., LABADZ, J.C. (2004): Artificial drainage of peatlands: hydrological and hydrochemical process and wetland restoration. Progress in Physical Geography, 28, 1, 95–123.

HOLDEN, J., WALLAGE, Z.E., LANE, S.N., MCDONALD, A.T. (2011): Water table dynamics in undisturbed, drained and restored blanket peat. Journal of Hydrology, 1–2, 402, 103–114.

HOJDOVÁ, M., HAIS, M., POKORNÝ, J. (2005): Microclimate of a peat bog and of the forest in different states of damage in the Šumava National Park. Silva Gabreta, 11, 1, 13–24.

HOWIE, S.A., MEERVELD, I.T. (2011): The Essential Role of the Lagg in Raised Bog Function and Restoration: A Review. Wetlands, 31, 3, 613–622.

JANSKÝ, B., KOCUM, J. (2008): Peat bogs influence on runoff process: case study of the Vydra and Křemelná River basins in the Šumava Mountains, southwestern Czechia. Geografie, 113, 4, 383–399.

JOOSTEN, H., CLARKE, D. (2002): Wise use of mires and peatlands – background and principles including a framework for decision-making. International Mire Conservation Group and International Peat Society, Totnes, Devon.

KOCUM, J. (2012): Tvorba odtoku a jeho dynamika v pramenné oblasti Šumavy. Dizertační práce. Univerzita Karlova, Přírodovědecká fakulta, katedra fyzické geografie a geoekologie. Praha.

KŘENOVÁ, Z., HRUŠKA, J. (2012). Proper zonation–an essential tool for the future conservation of the Šumava National Park. European Journal of Environmental Sciences, 2, 1, 62–72.

KUČEROVÁ, A., KUČERA, T., HÁJEK, T. (2009): Mikroklima a kolísání hladiny podzemní vody v centrální části Rokytecké slati. In: Černý, D., Dvořák, L. (eds.): Weitfällerské slatě. Sborník referátů ze semináře 21.1.2009. Správa NP a CHKO Šumava, Vimperk, 50–57.

LABADZ, J., ALLOTT, T., EVANS, M., BUTCHER, D., BILLETT, M., STAINER, S., YALLOP, S., JONES, P., INNERDALE, M., HARMON, N., MAHER, K., BRADBURY, R., MOUNT, D., O'BRIEN, H., HART, R. (2010): Peatland hydrology. Draft scientific review, commissioned by the IUCN UK Peatland Programmes Commission of Inquiry on Peatlands, https://www.iucn-uk-peatlandprogramme.org/resources/commission-inquiry/work-commission-2011/peatland-hydrology (20.9.2019).

LINDSAY, R., BIRNIE, R., CLOUGH, J. (2014): Peat Bog Ecosystems: Structure, Form, State and Condition. IUCN UK Committee Peatland Programme Briefing Note No 2, https://repository.uel.ac.uk/item/85872 (20.9.2019).

O'DRISCOLL, C., O'CONNOR, M., ZAKI-UL-ZAMAN, A., DE EYTO, E., BROWN, L.E., XIAO, L. (2016): Forest clearfelling effects on dissolved oxygen and metabolism in peatland streams. Journal of Environmental Management, 166, 15, 250–259.

PENMAN, H.L. (1948): Natural evaporation from open water, bare soil and grass. Proceedings of the Royal Society of London. Series A, Mathematical and Physical Sciences, 193, 1032, 120–145.

PONZIANI, M., SLOB, E. C., NGAN-TILLARD, D. J. M., VANHALA, H. (2011): Influence of water content on the electric conductivity of peat. International Water Technology Journal, 1, 1, 14–21.

PROKŠ, M. (2010): Odtokový režim v povodí Hamerského potoka se zaměřením na reakci pH vody ve vybraných povodňových epizodách. Diplomová práce. Univerzita Karlova, Přírodovědecká fakulta, katedra fyzické geografie a geoekologie, Praha.

PURANEN, R., MÄKILÄ, M., SÄÄVUORI, H. (1999): Electric conductivity and temperature variations within a raised bog in Finland: Implications for bog development. The Holocene, 9, 1, 13–24.

ŠEFRNA, L. (2004): Pedologická charakteristika povodí Otavy ve vztahu k povodním. Sborník příspěvků GAČR 205/Z052/03, 196–212.

ŠRÁČEK, O., KUCHOVSKÝ, T. (2003): Základy hydrogeologie. Masarykova Univerzita v Brně, Přírodovědecká fakulta.

VLČEK, L. (2017): Retence vody v půdách horských oblastí na příkladu Šumavy. Dizertační práce. Univerzita Karlova, Přírodovědecká fakulta, katedra fyzické geografie a geoekologie, Praha.

VLČEK, L., KOCUM, J., KUČEROVÁ, A., JANSKÝ, B., ŠEFRNA, L. (2012): Retenční potenciál a hydrologická bilance horského vrchoviště: případová studie Rokytecké slatě, povodí horní Otavy, jz Česko. Geografie, 117, 4, 371–395.

WILSON, L., WILSON, J., HOLDEN, J., JOHNSTONE, I., ARMSTRONG, A., MORRIS, M. (2011): The impact of drain blocking on an upland blanket bog during storm and drought events, and the importance of sampling-scale. Journal of Hydrology 404, 3-4, 198–208.

WIND-MULDER, H.L., ROCHEFORT, L., VITT, D.H. (1996): Water and peat chemistry comparisons of natural and post-harvested peatlands across Canada and their relevance to peatland restoration. Ecological Engineering, 7, 3, 161–168.

WORRALL, F., BURT, T., ADAMSON, J. (2006): Long-term changes in hydrological pathways in an upland peat catchment—recovery from severe drought? Journal of Hydrology, 321, 1, 5–20.

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7.5 Vliv revitalizačních opatření horských vrchovišť na hladinu podzemní vody

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EVALUATION OF THE INFLUENCE OF MOUNTAIN PEAT BOGS RESTORATION MEASURES ON THE GROUNDWATER LEVEL: CASE STUDY ROKYTKA PEAT BOG, THE ŠUMAVA MTS., CZECH REPUBLIC

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ABSTRACT

The paper evaluates measures taken to restore mountain Peat Bogs and their effect on hydrological regime, with the main focus on groundwater levels. The level of groundwater is a key factor in maintaining the character of mountain Peat Bogs and the main objective of restoration is to increase and stabilize the groundwater level in disturbed Peat Bogs. At the same time, the paper provides a complex overview of the topic, which is being often discussed nowadays, mostly due to a big retention potential of mountain Peat Bogs. The paper is based on detailed measurements of groundwater levels in a selected experimental drainage ditch in the catchment of the Rokytka stream. Basic statistical characteristics, the equation of Penman-Montheit or antecedent precipitation index were used to show the dependence of groundwater level on precipitation or evapotranspiration. The results show a positive influence of the restoration measures on Peat Bogs. In this case it has been confirmed that restoration measures cause increase of groundwater level and decrease its fluctuation in the Peat Bog.

Keywords: Peat Bog, groundwater level, the Šumava Mts., Peat Bog restoration

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1. Introduction

In the context of occurrence of hydrological extremes (floods, droughts), the increase of retention ability of headwater areas has recently become a fundamental question. The headwater area of the Otava River is characterized by a great amount of Peat Bog complexes, whose hydrological regime has not been completely uncovered yet, in spite of numerous analyses (Janský and Kocum 2008). The most recent studies emphasize that the occurrence of Peat Bogs in a catchment increases the extremity of flow (Holden et al. 2011; Holden et al. 2001; Ferda et al. 1971; Čurda et al. 2011).

The increase of a discharge in streams which have been restored is particularly significant at the Šumava Mts. catchments (Čurda et al. 2011). However, Peat Bog restoration measures could also have a negative effect on runoff process during flood events (Holden et al. 2011).

Restoration measures contribute greatly to a decrease of fluctuation of drainage ditches in the cases of mean and low flow. However, in the case of a higher water content caused by intensive precipitation, the barriers, which retain water in a catchment, might have negative effects on area retention capacity. After an excess of retention capacity of these barriers, an intense and rapid increase of flow follows, reaching a higher extremity (Čurda et al. 2011). Organic soils or other waterlogged areas saturated with water can then function as an outflow accelerator. Despite the fact that organic soils have a great retention capacity for water, releasing it gradually to the streams, their retention capacity is not applied in the case of water saturation (Šefrna 2004).

The depth of groundwater in organic soils is a very important factor for Peat Bog ecosystems. In an undisturbed Peat Bog, groundwater is situated very close to surface for most of the year and water fluctuation is largely limited (Holden et al. 2001). The changes in groundwater level concern mainly Acrotelm, which is characterized by higher porosity. Lower situated Catotelm includes more decomposed organic material with smaller pores and lower hydraulic conductivity, so the water movement is extremely limited there (Rizzuti et al. 2004). The combination of the characteristics of Acrotelm and Catotelm thus makes Peat Bogs a significant water reservoir with a unique hydrological regime in the area (Holden et al. 2011). Dynamics of groundwater level is also significant during a low precipitation period. Peat Bog reacts very fast. The rate of groundwater level changes can reach several centimeters per day (Vlček et al. 2012). The main factors influencing groundwater level in Peat Bogs are precipitation, evapotranspiration, topography and, in a local scale, also peat porosity and hydraulic conductivity (Allott et al. 2009).

The main changes in the Šumava Mts. Peat Bogs have been caused by efforts of draining and drying. Peat Bogs have been traditionally drained for the purpose of peat exploitation, agricultural land cultivation, or increase in wood exploitation in waterlogged forest areas. Nevertheless, the extent of surface drains was already considerable at the turn of the 19th and the 20th century. However,

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the major period of drainage digging was in the 70s and 80s of the 20th century. Nowaday, the drainage systems are still visible. Stocktaking researches have displayed that drainage has affected almost 70% of Peat Bogs in the Šumava Mts. (Bufková 2013). The open system of drains causes especially: fast surface flow, steeper culmination, and higher fluctuations of groundwater level (Ballard et al. 2011). Performed restorations can improve these aspects and consequently increase the groundwater level by several centimeters in a year (Worrall et al. 2007). A research from Schachtenfilz in the Bavarian Forest has confirmed that restoration measures increased groundwater level and decreased its fluctuation (Bufková 2013).

Since 1998 a complex restoration program has been implemented in the area of the National Park of Šumava (The Program of Restoration of Šumava Wetlands and Peat Bogs). The program is primarily aimed at a general improvement of disturbed water regime in the Peat Bog area (Bufková et al. 2010). A concept of so called "target water level" has been exercised during the restoration in the Šumava Mts. The method is based on determination of necessary water level, which is particular for each Peat Bog, eventually for their parts, and which is desirable to be achieved by restoration measures. The necessary water level can be described as a maximal tolerated decline of water in a ditch under the dam head, which is bearable for a given type of a Peat Bog (Bufková 2006). However, the increase of water level can be only observed few meters from a restoration because groundwater level is no longer influenced by the drainage system in a further distance from the drainage ditch (Wilson et al. 2011; Holden et al. 2011).

Peat Bogs are physically and ecologically adapted on the depth of groundwater level. The depth has a great significance for ecological niches of vegetative species and hence even for peat development (Kværner and Snilsberg 2011). The response of groundwater level on an exercised restoration is usually very fast; nevertheless, the changes in water chemism and consequent reactions of Peat Bog species are very slow. Peat Bog vegetative species are vulnerable and sudden changes of pH factor or changes in the amount of nutrients after exercising restoration can also have negative effects. Peat Bog restoration consequently includes stabilization and increase of groundwater level and a repeated habitation of the standpoint by Peat Bog species. It is thus important to limit the amount of water drain (Holden 2005).

However, there are still several problems, namely uncertain influence on water drainage ditch and water chemism, uncertain response species on water regime changes, and the price for restoration (Holden 2005).

The main aims of this paper are:

- to establish the influence of a drainage ditch on groundwater level in an experimental catchment;
- to determine dependence of groundwater level behavior on evapotranspiration and precipitation and its influence on groundwater level changes;
- to describe differences in groundwater levels near a functional drainage ditch and near a restored part of a drainage ditch (increased water level in the ditch due to a wooden dam).

2. Site description

The catchment of the Rokytka stream (Fig. 1) is located in the central part of the Šumava Mts. The whole complex of Rokytka Peat Bogs is placed on moderate slopes near the bottom of the Rokytka stream valley. The complex comprises several large and many small mountain Peat Bogs, which are surrounded by forest Peat Bogs, waterlogged pine stands, and minerotrophic sedge Peat Bogs (Bufková 2009). The total area of the Peat Bog is almost 250 ha. The depth of large Peat Bogs is about 5 m. Although in some locations, it can reach up to 7 m (Bufková and Spitzer 2008). Height relations of the catchment are consistent with the location of the central flat areas of the Šumava Mts. The altitude alters between 1089 and 1244 m a.s.l. The Rokytka catchment is rather flat in spite of its high altitude. The average gradient of slope is only 4°. Only exceptionally the gradient of slope reaches up to 10°, with the maximum of 12° (Jelínek 2009).

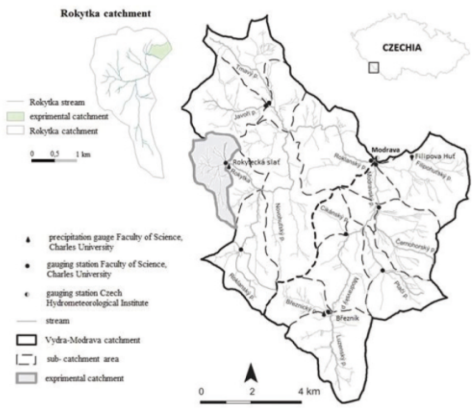


Fig. 1 The catchment of the Vydra River with the positions of gauging stations of the Czech Hydrometeorological Institute and automatic level gauges and precipitation gauges of the Faculty of Science, Charles University. The Rokytka stream catchment and the monitored experimental catchment is highlighted. Source: Kocum, Janský (2009).

The research of the Rokytka Peat Bog was focused on a selected experimental drainage ditch, which is located in the northern part of the catchment, at 1100 m a.s.l., and which drains an area of 0.14 km^2. The drainage ditch was partially dammed by small restoration dams; partially it was left functional, with a depth of 1 m.

In terms of soil cover, the soil region consists of entic Podzol, even of Rankers in steep slope areas (Šefrna 2004).

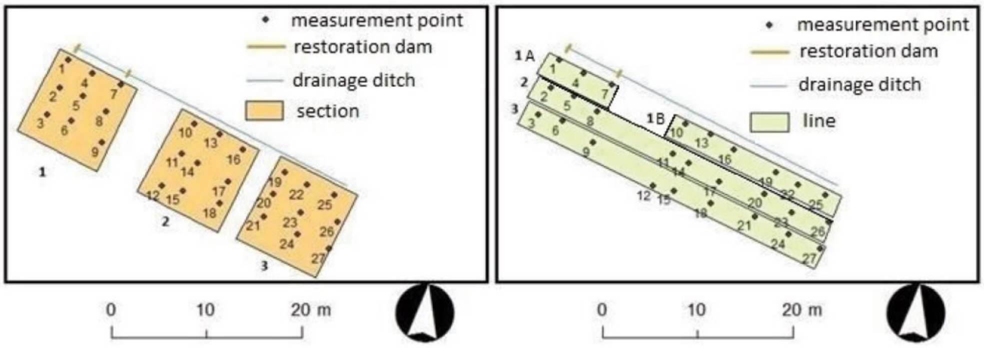


Fig. 2 The scheme of particular measurements of groundwater level and of the segments where the groundwater level was measured.

The soil of the researched catchment is a typical example of soil of the Šumava Mts., where a vertical sequence of soil is typical. Organic soil only occurs in watersheds and at the bottom of the valley. The catchment of Rokytka is mostly covered by Histosols. A peaty Gley can be found in some parts of the stream floodplains. Gleys are represented only marginally. The monitored drainage ditch is located in the part of a Peat Bog with a prevailing occurrence of fibric organic soil. The total area of organic soils in the catchment is 0.87 km^2, which stands for 23% of the whole area (Vlček et al. 2012).

In the area of interest, there are two precipitation gauges with long data series. It is Březník (1150 m a.s.l.) and Modrava (1000 m a.s.l.). The annual precipitation in Březník alters between 1100 and 1300 mm. The long term average of annual precipitation from Modrava is 1100 mm (Kocum 2012).

The annual discharge minimum can be measured at the end of February (before the snow melting) or in September (at the end of summer, a dry period). Discharge maximum is generally in spring, when the snow melts. A significant fluctuation in the outflow can also be seen in the summer period, due to a higher frequency of intensive precipitation. However, this fact does not influence the average monthly discharge variability, which might be caused by short duration of precipitation (Kocum 2012).

The vegetation in the Rokytka catchment is formed by a relict plant community. Low grass with the growth of *Trichophorum caespitosum* can be found there, predominantly. It blends in a mosaic pattern with the hydrophilic vegetation of shallow oblong depressions and the edges of lakes. It is made of a mat of *Sphagnum cuspidatum* and *Sphagnum majus* with the growth of *Carex limosa* and *Scheuchzeria palustris*. One of the constituents of *Trichophorum caespitosum* are obvious, large, and cambered bults with reddish types of peat as *Sphagnum magellanicum*, *Sphagnum russowii* or *Sphagnum rubellum* (Bufková 2009). In the surroundings of the monitored drainage ditch, the herb vegetation is formed by *Vaccinium uliginosum*, *Empetrum nigrum* or *Andromeda polifoli*a are also plentifully represented here. Nevertheless, the most dominant is vegetation primarily *Pinus mugo*.

3. Methodology

Groundwater level measurements were implemented during the period from August 14, 2014 to October 31, 2014. The groundwater level was measured manually in tubes which were inserted into the peat to a depth of 1–1.5 m. The water levels were measured in lines which were copying parts of the drainage ditch and the distance between measurement points was 3 meters. Thus, a regular net with 27 groundwater level measurement points, placed in regular distances, was created. The groundwater level was measured from the surface. For this purpose, particular segments were created from the measuring areas and the groundwater levels were then compared with each other within the scope of the individual sections and lines (see Fig. 2). The line 1 was divided into part A and part B for better accuracy. Part A is located directly to restoration dams and part B is placed in area which is not affected of restoration measures. At the same time, tubes were located by a total station, so the exact location of measurement points is known and therefore maps and interpolations could be created. At each point, 28 values of groundwater level were measured. Further, particular level changes were statistically evaluated in the scope of individual sections and lines to better demonstrate the dependence of groundwater level fluctuation on the distance from a drainage ditch, or from restoration dams. Data of groundwater level from an automatic station in Rokytka Peat Bog were also used. Additionally an interpolation method "natural neighbor" was applied to obtain range information. At first, the whole dataset was analyzed by basic statistical characteristics and data testing. For distribution of measured values of groundwater level in various intervals box plots were used. Statistical characteristics variance, correlation coefficient and directive deviance were calculated. All the statistical

Tab. 1 Statistical characteristics of groundwater level fluctuation in sections and lines.

	line 1A	line 1B	line 2	line 3	section 1	section 2	section 3
Number of measured values	84	168	252	252	252	252	252
Average groundwater level (cm)	13.04	23.04	18.13	10.49	10.95	17.93	19.37
Medium (cm)	12.90	24.70	18.90	10,80	11.20	15.85	18.90
Minimum (cm)	0.10	7.50	2.30	0.10	0.00	6.30	4.80
Maximum(cm)	28.00	50.00	42.00	30.00	28.80	50.00	50.00
Variance	28.98	69.16	53.94	30.67	46.98	63.09	52.72
Directive deviation	5.39	8.31	7.34	5.54	6.85	7.94	7.26

procedures were calculated in a statistical software StatSoft Statistica.

Groundwater level fluctuation was put into context with particular significant factors of rainfall-runoff process. One of them is potential evapotranspiration (equation 1). In this research, Penman-Monteith equation was used for the determination of daily potential evapotranspiration. In this case, the daily potential evapotranspiration is calculated according to the following equation:

$$PET_0 = \frac{0.408 \cdot \Delta \cdot (Rn - G) + \gamma \cdot \frac{900}{T + 273.16} \cdot u \cdot (e_s - e_a)}{\Delta + \gamma \cdot (1 + 0.34 \cdot u)} \quad (1)$$

where Δ represents the inclination of water vapor saturation curve in dependence on temperature [kPa °C^{-1}], Rn radiation balance [MJ m^{-2} day^{-1}], G flow of heat into soil [MJ m^{-2} day^{-1}], Υ psychometric constant [kPa °C^{-1}], u speed of wind [m s^{-1}], $(e_s - e_a)$ saturation deficit of air in elevation z [kPa], T average air temperature [°C] (Penman, 1948).

The antecedent precipitation index API (equation 2) is also applied in this paper. The index is used for determination of catchment saturation and it expresses the influence of precipitation which occurred in previous days to the given date. It thus demonstrates the ability of a catchment to absorb more precipitation. For the purpose of this paper, the API index was calculated for five previous days according to the following equation:

$$API = \Sigma 0.93^i \cdot P_i \, (2)$$

where "i" stands for the number of the day counted back from the date, which API is counted for, P daily amount of precipitation, [mm] in the i-th day before the causal precipitation (Mishra and Singh 2003).

4. Results

4.1 Statistical evaluation of groundwater level fluctuation

When basic statistical characteristics were used, significant differences between the areas lying near restoration dams and those with the absence of restoration were ascertained. At first, statistical differences were tested in sections using analysis of variance ANOVA. Significant differences on the probability level ($p < 0.05$) were proven. The resulting coefficients were significantly higher than the critical value of the distribution (31.5 > 3). It means that high differences between data sets were detected. However, the analysis was not able to show where exactly the differences occurred. Due to this fact, a t-Test was used on the probability level ($p < 0.05$). The only significant difference was detected between section 1 and section 3. In other cases, the differences were not prominent.

In the case of lines, a uniquely high difference can be found in the biggest proximity of the drainage ditch (line 1A and 1B), Fig. 3. The difference between average groundwater levels in these two lines during a monitored period was 10 cm. The data from a distance of 6 m from the drainage ditch are very similar to line 1A from the location with restoration measures. Consequently, it can be presumed that in the distance of 6 m from restoration measures, the behavior of the groundwater level seems natural. The amplitude in the distance of 6 m from the drainage ditch was only 29.9 cm. It is very similar to line 1A which contains data from the area with restoration measures. A similar divergence can be seen during evaluation of particular sections. In the scope of section number 1 (Fig. 3), which is the closest one to the exercised restoration, it was proven that the groundwater level is less fluctuated and that it remains near the surface. In this section, the average groundwater level was 10.95 cm and the amplitude was 8 cm. On the contrary, in the furthest section, the average groundwater level, for the monitored period of time, was 19.37 cm and the amplitude reached up to 45.2 cm. Thus, in the proximity of restoration, the level of groundwater is located 8.42 cm higher on average.

In Table 1, statistical characteristics of groundwater level fluctuation in sections and lines are shown. For example, line 1B, which is the nearest to the drainage ditch without restoration measures, expresses the highest numbers in variance and in directive deviations. On the contrary, the furthest line 3 and line 1A with restoration measures have significantly lower values. When sections were compared, it was observed that section 1, which is the closest to restoration, disposes of the lowest numbers

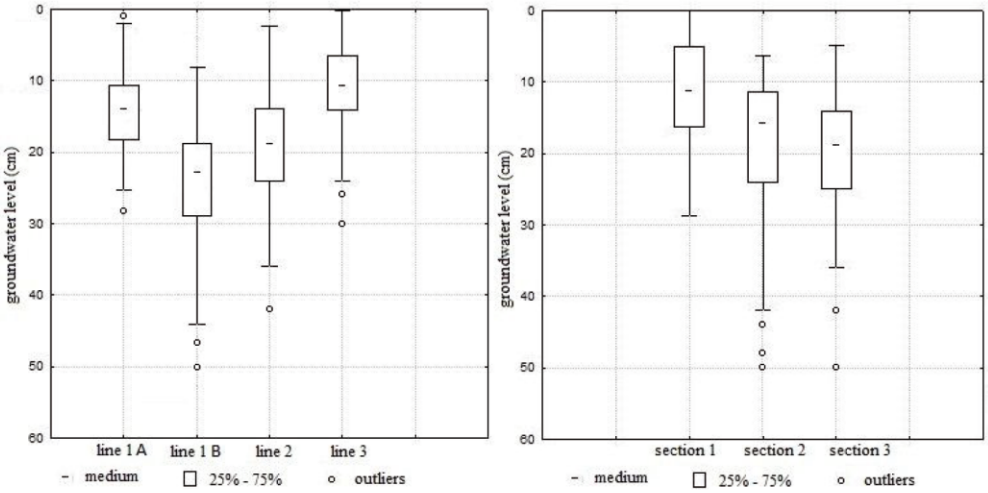


Fig. 3 Groundwater level fluctuation in the scope of particular sections and lines.

in variance and in directive deviations. Further sections show expressively higher values which point to a positive influence of restoration measures on fluctuation and on groundwater level.

The highest groundwater level can be found at measuring points in the proximity of restoration. The above mentioned statement, that the higher the distance from the drainage ditch, the higher the groundwater level, is also confirmed here. Another important finding is that in the distance of 6 m from the drainage ditch, the drainage effect can no longer be seen.

4.2 The dependance of groundwater level on meteorological factors

In particular sections, the groundwater level fluctuation was compared to the difference between precipitation and evapotranspiration between measurements.

It was found that these two factors are the most essential ones for groundwater level. A dependence was visible in Fig. 4, moreover, with quite high coefficients of determination. The dependence was proven most significantly in the section located the closest to the drainage ditch. Consequently, a daily groundwater level fluctuation can be rather accurately estimated and explained with the regard of these two factors.

A few values of decrease of groundwater level were also observed when precipitation was higher than evapotranspiration. This fact could be caused by an interception. During a lower precipitation period, rainwater is probably intercepted by trees, which growth rather densely in the experimental area.

For the quantification of the dependence, coefficients of determination and other correlation coefficients were calculated. Correlation coefficients in Table 2 reach very high values. The dependence of groundwater level on precipitation and evapotranspiration is statistically significant in sections 1 and 3.

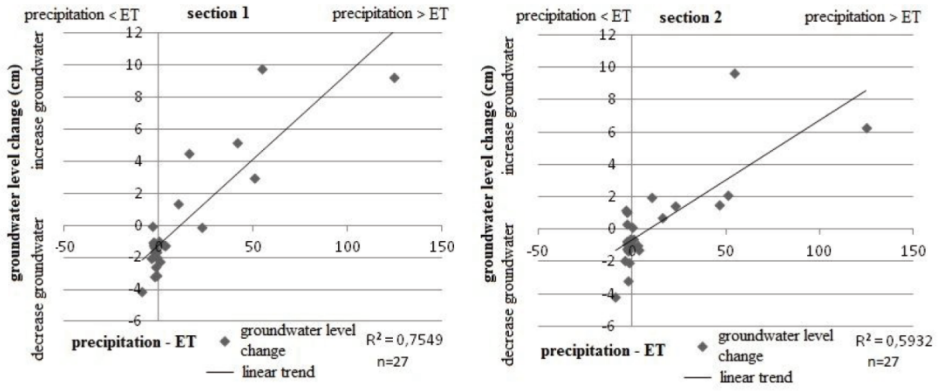
Tab. 2 Correlation coefficients and determination coefficient in particular sections, mark * is statistical significant at probability level, $p < 0.05$.

	section 1	section 2	section 3
Coefficient of determination (level change; ET-precipitation)	0.869*	0.770	0.842*
Correlation coefficient (level change; ET-precipitation)	0.755*	0.593	0.708*
Correlation coefficient (average level; API)	−0.532*	−0.306	−0.511*

The same approach was exercised at the automatic station at Rokytka Peat Bog. The difference is in the fact that the station is in a sufficient distance from the drainage ditch and therefore it is not influenced by melioration. It was revealed that the difference of precipitation and evapotranspiration describes the actual course of groundwater level more precisely (see Fig. 5). The moving average of the difference of precipitation and evapotranspiration follows precisely the process of groundwater level changes. The changes of groundwater level in an area with no human influence in Peat Bog can be expressed more precisely and its course corresponds better to meteorological factors.

4.3 Groundwater level fluctuation during chosen episodes

The variability of groundwater level is also an important factor. Two episodes have been selected for an evaluation. The first one, an episode of intensive precipitation



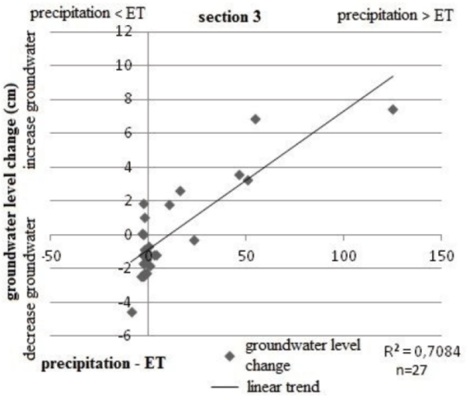


Fig. 4 Groundwater level change in the dependence on the difference of precipitation and evapotranspiration in particular sections.

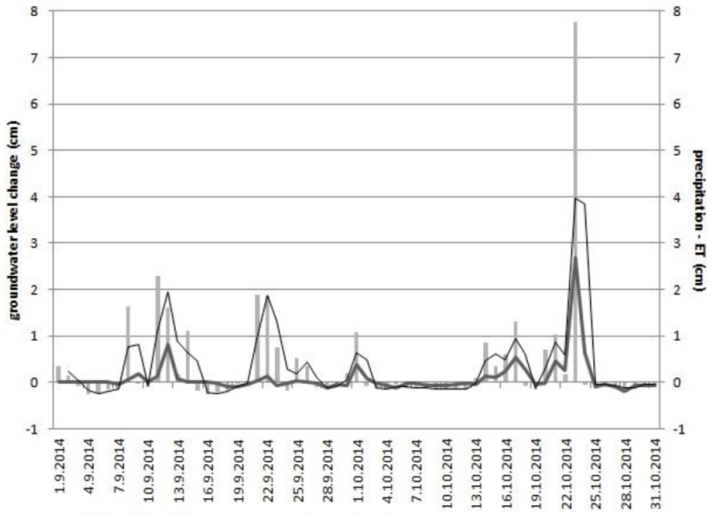


Fig. 5 The fluctuation of daily average groundwater level in dependence on the difference of precipitation and evapotranspiration in Rokytka Peat Bog, from the September 1, 2014 to October 31, 2014. Source: Data from the automatic station of Faculty of Science, Charles University.

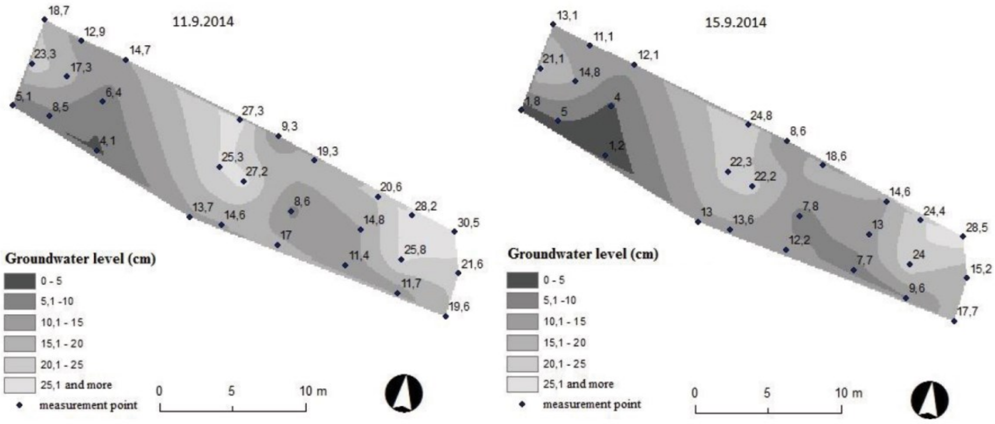


Fig. 6 Changes of groundwater level during a selected episode of intensive precipitation between the September 11, 2014 and September 15, 2014. The given numbers in the graph represent measured groundwater level in centimeters on a given day.

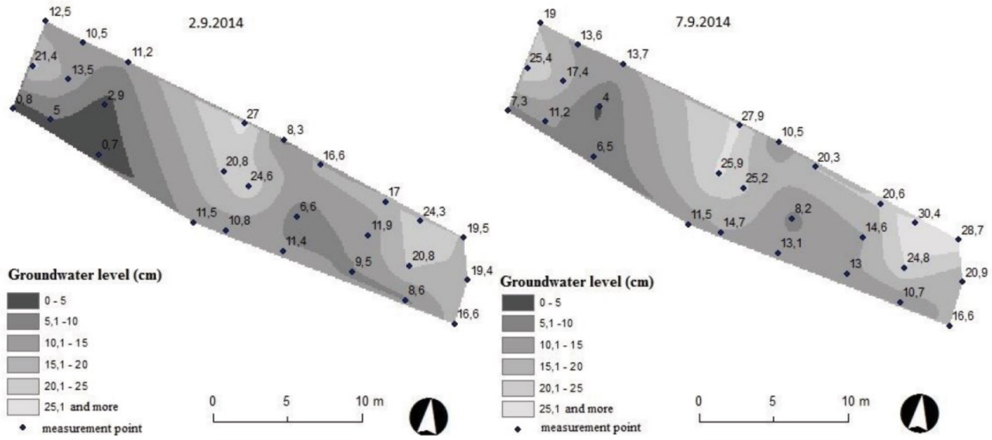


Fig. 7 Changes of groundwater level during a selected episode of drought between the September 2, 2014 and September 7, 2014. Given numbers in the graph represent measured groundwater levels in centimeters on a given day.

(55.4 mm), was analyzed between the September 11, 2014 and September 15, 2014 at the Rokytka catchment. It is obvious that groundwater level along the drainage ditch shows a high amplitude, see Fig. 6. With longer distance from the drainage ditch, the groundwater level increases and its change during an episode decreases. The level is the highest in the section close to restoration dams. Their influence is perceived as positive, as they raise groundwater level. They also have a stabilizing effect. However, the results also imply that in a certain distance from restoration dams, their effects can no longer be seen and groundwater level fluctuates naturally as in the Peat Bogs, which are not influenced by a drainage. It is also evident that the decreases or increases of water level are very variable and there are noticeable differences between individual points, in spite of the fact that it is a small homogenous area. The difference between the spot with the highest and with the lowest decrease is 6.4 cm. On the contrary, in the areas near restoration dams, the groundwater level was increasing very gradually and a similar increase was reached at all the measurement points.

Another observed episode was during a dry period, when there was only 1.4 mm of precipitation from the September 2, 2014 to September 7, 2014 (see Fig. 7). The smallest changes of groundwater level in a period with low precipitation were reached in the middle line of the observed area, precisely in the distance of 3 m from the drainage ditch. It is interesting that in this episode, rather big amplitudes can be found, even in the area of restoration. It can be caused by the fact that before the period of drought, the groundwater level was very high, precisely right under the surface; hence, following decreases could have progressed faster there. The biggest difference between water levels is significant again and it is even up to 9.2 cm during the monitored five day range. It has been confirmed repeatedly that in the areas located further from restoration, the groundwater level is distinctly lower and, moreover, there is a remarkable and fast fluctuation of groundwater, which is not beneficial for the evolution of mountain Peat Bogs.

5. Discussion

The results of the research conducted in Rokytka Peat Bog are in correspondence with other national and international researches concerned with the same subject (Bufková 2013; Vlček et al. 2012; Worrall et al. 2007; Wilson et al. 2011). The most significant finding is that restoration measures have the ability to increase groundwater level and to decrease its amplitude. In this case, it was statistically observed especially in line 1A, which contains data from restoration measures. At the point of restoration, the mean groundwater level was up to 10 cm higher than at a place with no restoration, similar to Worrall et al. (2007). Holden's research (2011) brings the difference of 4 cm. Ketcheson and Price (2011) demonstrated that the size of groundwater level change depends on many factors. Similar results in the Šumava Mts., particularly in the Schachtenfilz Peat Bog, were achieved, for example, by Bufková (2013), who declared that after three years from a restoration, the average groundwater level increased and its fluctuation was considerably reduced, especially in the most disturbed parts of the Peat Bog and in the forest cover of peat and pine grove. In our case it was also statistically significant.

The groundwater level increases more steeply perpendicularly from the drainage ditch in the area of restoration in contrast with the area without restoration. Thus, it can be presumed that restoration measures have an important role in the attenuation of negative effects of a drainage ditch. It was confirmed by a research of Haapalehto et al. (2011) who found that damming and filling of the ditches resulted in a raised groundwater level in the restored bog and fen systems by several centimeters. However, in a certain distance, these differences decrease and groundwater level behaves naturally as in a Peat Bog with no human disturbance. This statement is also confirmed by the work of Wilson et al. (2011). Although, it was revealed that particular changes of groundwater level can be rather variable, in spite of the fact that the researched area was small and very homogenous. Vlček et al. (2012) notes that daily relations of groundwater level are quite fast and they can reach several centimeters during a day. In this case, even higher values were observed in some areas, due to the fact that in the proximity of the drainage ditch, significantly higher amplitudes appeared, compared to the further located areas from it. Thus, it was proven that the influence of restoration measures on groundwater level is positive in this case.

This research has also demonstrated that the evaluation of factors of precipitation and evapotranspiration is sufficient to clarify precisely the changes of groundwater level in Peat Bogs. It is manifested predominantly by high values of correlation and determination coefficients. On the other hand, a little decrease was observed in the groundwater level, even though there was a higher precipitation than evapotranspiration. It could be caused by interception during a low intensity rain. Kellner (2003) and Allott et al. (2009) note that the size of evapotranspiration from wetlands is variable. General factors influencing the evapotranspiration are surface conditions, such as roughness, temperature, and dryness, together with the air temperature, humidity, and solar radiation. Since the climatic factors vary significantly, it is hard to generate absolute values without a great uncertainty.

The correlation of groundwater level with the index of previous precipitation manifested itself in the same manner. An important aspect of the paper was also to refer to space variability of groundwater level changes. Though the observed area is small and homogenous, a high variability in groundwater level changes appeared even during selected episodes. The difference between maximum increase and decrease reached up to 9.2 cm during five days. Ketcheson and Price (2011) reminded that topographical variability and the location of the dams can strongly influence the magnitude of the groundwater level rise at any given location. Price et al. (2003) notes that groundwater level also depends on the depth of the ditch, the distance between ditches, and the hydraulic conductivity of the peat. For example, Wilson et al. (2011) observed conflicting results in a few experimental locations, when after drain blocking, groundwater level was deeper than before blocking. It shows that in a Peat Bog, there are some places with non-standard local specific water regime.

Changes of groundwater level were especially prominent in the areas far from restoration. It was also discovered that the lowest deviations (0.48 cm) were achieved at the automatic station, which is not influenced by melioration. The moving average of the difference between precipitation and evapotranspiration at this point demonstrated that it copies the natural course of groundwater level changes very accurately. This statement is confirmed by another research (Vlček et al. 2012), which was also implemented in Rokytka Peat Bog; similar daily changes were observed, approximately 2–3 cm according to the temperature and precipitation. A rapid and sudden increase of groundwater level during intensive precipitation was also confirmed (Kučerová et al. 2009). Despite the fact that many factors are involved in groundwater level changes, the dependence can be largely explained by the difference between precipitation and evapotranspiration. It was observed also in Wilson et al. (2011).

6. Conclusion

The headwater area of the Otava River is characterized by a high portion of Peat Bogs, thus it is an area with a very specific hydrological regime. In the context of the occurrence of hydrological extremes, it is very important to pay attention to the retention potential of the areas.

The most significant outcome of the paper is the demonstration of positive effects of a restoration on groundwater level. It was proven that restoration decreases fluctuation and increases groundwater level, which is essential for a natural evolution of a mountain Peat Bog. The main factors of groundwater level fluctuation are predominantly evapotranspiration and precipitation; the explanation of this phenomenon is demonstrated by high values of determination and correlation coefficients. The paper thus contributes to the understanding of the retention potential of peat complexes, which is essential for the understanding of the hydrological regime of the Otava River. This part of the Šumava Mts. can consequently offer other valuable findings, due to its significant retention potential.

REFERENCES

ALLOTT, T. E. H., EVANS, M. G., LINDSAY, J. B., AGNEW, C. T., FREER, J. E., JONES, A., PARNELL, M. (2009): Water tables in Peak District blanket peatlands. Moors for the Future, Report 17, 49 p.

BALLARD, C. E., MCINTYRE, N., WHEATER, H. S., HOLDEN, J., WALLAGE, Z. (2011): Hydrological modelling of drained blanket peatland. Journal of Hydrology 407, 81–93. https://doi.org/10.1016/j.jhydrol.2011.07.005

BUFKOVÁ, I. (2006): Revitalizace šumavských rašelinišť. Zprávy České Botanické Společnosti, Praha 41(21), 181–191.

BUFKOVÁ, I., SPITZER, K. (2008): Šumavská rašeliniště. Správa Národního parku a Chráněné krajinné oblasti Šumava, Vimperk, 203 p.

BUFKOVÁ, I. (2009): Ochrana rašelinišť na Šumavě aneb byly Rokytecké slatě první? In: Černý, D., Dvořák, L.: Weitfallerské slatě. Sborník z výzkumu na Šumavě, 2. Správa NP a CHKO Šumava, Vimperk, pp. 12–22.

BUFKOVÁ, I., STÍBAL, F., MIKULÁŠKOVÁ, E. (2010): Restoration of drained mires (Šumava National Park, Czech republic). Proceedings 7th European Conference on Ecological Restoration Avignon, France, pp. 23–27. https://doi.org/10.1007/978-90-481-9265-6_16

BUFKOVÁ, I. (2013): Náprava narušeného vodního režimu rašelinišť v národním parku Šumava. Ochrana přírody 2, 17–19.

ČURDA, J., JANSKÝ, B., KOCUM, J. (2011): Vliv fyzicko-geografických faktorů na extremitu povodní v povodí Vydry. Geografie 116(3), 335–353.

FERDA, J., HLADNÝ, J., BUBENÍČKOVÁ, L., PEŠEK, L. (1971): Odtokový režim a chemismus vod v povodí Horní Otavy se zaměřením na výskyt rašelinišť. Sborník prací ČHMÚ 17, 22–126.

HAAPALEHTO, T. O., VASANDER, H., JAUHIAINEN, S., TAHVANAINEN, T., KOTIAHO, S. J. (2011): The Effects of Peatland Restoration on Water-Table Depth, Elemental Concentrations, and Vegetation: 10 Years of Changes. Society for Ecological Restoration International. Restoration Ecology 19(5), 587–589. https://doi.org/10.1111/j.1526-100x.2010.00704.x

HOLDEN, J., BURT, T. P., COX, N. J. (2001): Macroporosity and infiltration in blanket peat: The implicaions of tension disc infiltrometer measurements. Hydrol. Process. 15(2), 289–303. https://doi.org/10.1002/hyp.93

HOLDEN, J. (2005): Peatland hydrology and carbon release: why small-scale process matters. Philosophical Transaction, The Royal Society 363, 2891–2913. https://doi.org/10.1098/rsta.2005.1671

HOLDEN, J., WALLAGE, Z., LANE S., McDONALD A. (2011): Water table dynamics in undisturbed, drained and restored blanket peat. Journal of Hydrology 402, 103–114. https://doi.org/10.1016/j.jhydrol.2011.03.010

JANSKÝ, B., KOCUM, J. (2008): Peat bogs influence on runoff process: case study of the Vydra and Křemelná River basins in the Šumava Mountains, southwestern Czechia. Geografie – Sborník ČGS 113(4), 383–399.

JELÍNEK, J. (2009): Akumulace a tání sněhové pokrývky v povodí Rokytky v hydrologických letech 2007 a 2008. Diplomová práce. Univerzita Karlova v Praze. Přírodovědecká fakulta, katedra fyzické geografie a geoekologie, 85 p.

KELLNER, E. (2003): Wetlands – different types, their properties and function. Technical report. Department of Earth Sciences /Hydrology, Uppsala University, 62 p.

KETCHESON, S. J., PRICE, J. S. (2011): The Impact of Peatland Restoration on the Site Hydrology of an Abandoned Block-Cut Bog. Wetlands 31(6), 1263–1274. https://doi.org/10.1007/s13157-011-0241-0

KOCUM, J., JANSKÝ, B. (2009): Retence vody v pramenných oblastech Vydry a Křemelné – případová studie povodí Rokytky. In: Černý, D. a Dvořák, L. (eds.): Weitfallerské slatě. Sborník referátů ze semináře 21. 1. 2009. Správa NP a CHKO Šumava, Vimperk, pp. 26–48.

KOCUM, J. (2012): Tvorba odtoku a jeho dynamika v pramenné oblasti Šumavy. Disertační práce. Univerzita Karlova, Přírodovědecká fakulta v Praze, Katedra fyzické geografie a geoekologie, 206 p.

KUČEROVA, A., KUČERA, T., HÁJEK, T. (2009): Mikroklima a kolísání hladiny podzemní vody v centrální části Rokytecké slati. In: Černý, D., Dvořák, L. (eds.): Weitfallerské slatě. Sborník referátů ze semináře 21. 1. 2009. Správa NP a CHKO Šumava, Vimperk, pp. 26–48.

KVÆRNER, J., SNILSBERG, P. (2011): Groundwater hydrology of boreal peatlands above a bedrock tunnel – Drainage impacts and surface water groundwater interactions. Journal of Hydrology 403, 278–291. https://doi.org/10.1016/j.jhydrol.2011.04.006

MISHRA, S. K., SINGH V. P. (2003): Soil Conservation Service Curve Numer (SCS-CN) Methodolgy. Dordrecht, Kluwer Academic Publisher, 511 p.

PENMAN, H. L. (1948): Natural evaporation from open water, bare soil and grass. Proceedings of the Royal Society of London. Series A, Mathematical and Physical Sciences 193(1032), 120–145. https://doi.org/10.1098/rspa.1948.0037

PRICE, J. S., HEATHWAITE, A. L., BAIRD, A. J. (2003) Hydrological processes in abandoned and restored peatlands: an overview

of management approaches. Wetlands, Ecology and Management 11, 65–83. https://doi.org/10.1023/A:1022046409485
RIZZUTI, A. M., COHEN, A. D., STACK, E. M. (2004): Using hydraulic conductivity and micropetrography to assess water flow through peat-containing wetlands. Int. J. Coal Geol. 60(1), 1–16. https://doi.org/10.1016/j.coal.2004.03.003
ŠEFRNA, L. (2004): Pedologická charakteristika povodí Otavy ve vztahu k povodím. Sborník příspěvků GAČR 205/Z052/03, pp. 196–212.
VLČEK, L., KOCUM, J., JANSKÝ, B., ŠEFRNA, L., KUČEROVÁ, A. (2012): Retenční potenciál a hydrologická bilance horského vrchoviště: případová studie Rokytecké slatě, povodí horní Otavy, JZ. Česko. Geografie 117(4), 395–414.
WILSON, L., WILSON, J., HOLDEN, J., JOHNSTONE, I., ARMSTRONG, A., MORRIS, M. (2011): The impact of drain blocking on an upland blanket bog during storm and drought events, and the importance of sampling-scale. Journal of Hydrology 404, 198–208. https://doi.org/10.1016/j.jhydrol.2011.04.030
WORRALL, F., ARMSTRONG, A., HOLDEN, J. (2007): Short-term impact of peat drain-blocking on water colour, dissolved organic carbon concentration, and water table depth. Journal of Hydrology 337, 315–325. https://doi.org/10.1016/j.jhydrol.2007.01.046