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Alexandre Lhosmot, Louis Collin, Geneviève Magnon, Marc Steinmann, Catherine Bertrand, et al.. Restoration and meteorological variability highlight nested water supplies in middle altitude/latitude peatlands: Towards a hydrological conceptual model of the Frasne peatland, Jura Mountains, France. *Ecohydrology*, 2021, 14 (6), pp.e2315. 10.1002/eco.2315 . hal-03469696

HAL Id: hal-03469696

<https://hal.science/hal-03469696v1>

Submitted on 10 Dec 2021

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Restoration and meteorological variability highlight nested water supplies in mid altitude/latitude peatlands. Toward a hydrological conceptual model of the Frasne peatland, Jura Mountains, France.

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Abstract: Peatlands and associated ecosystem services are sensitive to climate changes and anthropogenic pressures such as drainage. This study illustrates these effects on the Forbonnet bog (7 ha), belonging to the Frasne peatland complex (~300 ha, French Jura Mountain) and shows how they can inform about the ecohydrological functioning of peatlands. The southern part of the Forbonnet bog was restored in 2015-2016 by backfilling of artificial drains dating from the end of the 19th century. Piezometric data from 2014 to 2018 allow to evaluate the restoration effect on the Water Table Depth (WTD) and highlight the reactivation of lateral inflows from the surrounding raised peatland complex. Vertical EC profiles permit to identify 3 main peat compartments depending on different water supplies arguing for a nested hydrological functioning. This involves: (i) One-off karst groundwater inputs at the substratum/peat interface supplying the deepest peat layer, (ii) lateral seepage inputs from the neighboring raised wooded peatlands sustaining the intermediate peat level, and (iii) direct

This article has been accepted for publication and undergone full peer review but has not been through the copyediting, typesetting, pagination and proofreading process which may lead to differences between this version and the Version of Record. Please cite this article as doi: 10.1002/eco.2315

rainfall infiltrating the most superficial peat layer. This nested multi-reservoir model operates at various spatio-temporal scales and is consistent with the complex seasonal hydrological and physico-chemical response at the bog outlet, which will be increasingly affected by climate change in the coming decades.

Keywords: Peatland restoration, Jura Mountains, Monitoring, Nested hydrological network, Climate change

1. Introduction

In the context of climate change management, peatlands are recognized as socio-ecosystems, because they are considered to be regulators for fresh water availability and greenhouse gas concentrations (Bonn et al., 2014; Reed et al., 2014). Although representing only 3 % of the global land area, peatlands contain about 10 % of the global freshwater resources (Joosten and Clarke, 2002; Xu et al., 2018) and are hotspots of specific biodiversity providing habitats for highly specialized species. In addition, these systems accumulated a great quantity of organic carbon over the long-term usually throughout the Holocene, and store up to 30 % of the global soil organic carbon (Gorham, 1991; Nichols and Peteet, 2019; Parish et al., 2008). Peatlands interact with watersheds and act as buffer for water, carbon and contaminant owing to their highly specific hydrological characteristics (high porosity, low permeability, great evapotranspiration; Blodau and Moore, 2003; Daniels et al., 2008; Francez et al., 2011; Holden and Burt, 2003; Parish et al., 2008; Rosset et al., 2019). These ecological functions are nowadays considered as socio-ecosystemic services. i.e. processes implying a socioeconomic value and favoring the well-being of our society (*e.g.* Quintas-Soriano et al., 2018).

The sustainability of these socio-ecosystemic services is strongly dependent on peatland structure and water saturation, which is a major environmental issue at different scales. Water Table Depth (WTD) is expected to regulate gas exchange between peatland and atmosphere. Shallow WTD promotes anoxic conditions, limiting aerobic respiration and CO₂ release and favoring at the same time CH₄ emissions by anaerobic processes. In contrast, deeper WTD leads to well oxygenated superficial peat layers, promoting organic matter degradation and CO₂ release (Blodau and Moore, 2003; Jassey et al., 2018; Moore and Knowles, 1989; Waddington and Day, 2007).

In parallel, the diplotelmic (acrotelm-catotelm) model suggests that peat physical and biogeochemical properties of peat vary vertically (Baird et al., 2016; Morris et al., 2011). The uppermost layer called “acrotelm” is highly permeable. It supports plant activities and is subject to punctual aerobic conditions, which explains its crucial role for biodiversity and water-carbon interactions. The deeper layers named “catotelm” are mostly impermeable and favor anaerobic processes (Bertrand et al., 2012; Boelter, 1965; Ingram, 1978).

Peatland hydrology and related ecosystem services have for several centuries been altered by anthropogenic disturbances such as peat extraction (for fuel and horticulture), agriculture and forestry which all require drainage and trigger WTD drop and peat subsidence (Joosten, 2015; Kløve et al., 2014; Parry et al., 2014). Protective measures were taken from local to global scale (Ramsar convention, 1971; EU Habitats and Species Directive, 1992, see FAO, 2020). Nevertheless, it remains that today a large proportion of global peatland area is disturbed. For instance, about 25 % of its original surface is currently used for forestry and agriculture (Page and Baird, 2016).

Anthropogenic disturbances are exacerbated by climate change: increase of air temperature (Shulka et al., 2019) and modification of rainfall seasonality (Trenberth, 2011) both impact WTD (Bertrand et al., 2021) and thus the ecohydrological functioning of peatlands.

The combined effects of climate change and human disturbances lead to increased peat oxidation (including throughout fires) and carbon release to the atmosphere (Page and Baird, 2016; Turetsky et al., 2015). Restoration programs were developed in order to preserve the above-mentioned socio-ecosystemic services (Järveoja et al., 2016; Kimmel and Mander, 2010; Menberu et al., 2016; Parry et al., 2014; Ramchunder et al., 2012; Zerbe et al., 2013) and to mitigate the impact of climate change (Humpenöder et al., 2020). Restoration ultimately aims to shift the carbon balance of disturbed peatland from source to sink (Swenson et al., 2019). In this purpose, rewetting (ditch blocking, drain filling) is a broadly used method to promote higher and more stable WTD. Although some short-term effects on hydrology are commonly observed, a much longer time span of several years to decades is needed to re-establish peat forming vegetation and carbon sequestration (Ahmad et al., 2020; Bortoluzzi et al., 2006; D'Acunha et al., 2018; Lucchese et al., 2010; Price et al., 2003; Zerbe et al., 2013). For instance, McCarter and Price (2013) showed in a Canadian peatland 10 years after restoration that Sphagnum mosses were regenerated, but that mosses water availability, which favor carbon sequestration, was still limited.

In order to mitigate the effects of local and global pressures, it is now required to understand the hydrological feedback of peatland ecosystems to meteorological variability and to propose sustainable management strategies (Ferlatte et al., 2015). Nevertheless, the hydrology of peatlands is not only controlled by *in situ* peatland-atmosphere interactions, but may also interact with the underlying substrates (Drexler et al., 1999; Reeve et al., 2000) or lateral inflows (Bourgault et al., 2019; Ferlatte et al., 2015). These interactions may constrain peatland role in baseflow support for downstream ecosystems (Martin and Didon-Lescot, 2007; Paran et al., 2017). The understanding of watershed-peatland relationships and internal water transfers are therefore crucial for modeling meteorological variability and long-term climatic

impacts on peatland hydrology in a restoration perspective (Ahmad et al., 2020; Hare et al., 2017; Kløve et al., 2014).

In this objective, this study illustrates the interest to combine geochemical and hydrological long-term monitoring in peatlands to evaluate the effects of restoration and meteorological variability on peatland hydrology in a mid-latitude/altitude mountainous bog. The integrated analysis of these data along the horizontal and vertical axes as well as from a temporal point of view allow to decipher mechanisms at the origin of the hydrological functioning, and will constitute the basis for proposing an integrated hydrogeological conceptual scheme, that will be required to better evaluate potential climate change impacts in the coming decades.

2. Context of the study

2a) Site presentation

The study site is located in the Jura Mountains in NE France (46.826 N, 6.1754E, 852 m a.s.l; Fig. 1A) within the Frasné-Bouverans peatland complex. This complex of 293 ha is a regional natural reserve (RNR Frasné-Bouverans) included in the long-term ecological research zone Jurassian Arc (ZAAJ, <https://zaaj.univ-fcomte.fr/?lang=en>) and to the international wetland Ramsar area “Peatlands and lakes of the Jura Mountains”. The study was more especially implemented at the Forbonnet bog (7 ha), south of the Frasné-Bouverans complex. This bog is monitored since 2008 from a hydrological, ecological and biogeochemical point of view (Laggoun-Défarage et al., 2008) and one of the four sites of the French national observatory of peatlands since 2012 (“SNO Tourbières”, <https://www.sno-tourbieres.cnrs.fr/>). Between 2009 and 2019, annual precipitation (rainfall and snowmelt) ranged between 1293 and 2110 mm with a mean of 1618 mm and the annual air temperature between 5.8 and 7.9 °C with a mean of 7 °C.

From a geological point of view (Fig. 1A) the Frasné-Bouverans peatland is located in a 10 km wide synclinal structure (Duraffourg and Palacio, 1981). Within the syncline, partially karstified Jurassic and Cretaceous limestones and marls are overlain by poorly permeable moraines of Pleistocene age (Campy, 1982; Duraffourg and Palacio, 1981). This moraine layer constitutes the basement of the peatland and forms an irregular landscape with small depressions and hills of decametric dimensions. The median peat thickness in the Forbonnet bog is 2.1 m based on 298 manual soundings and reaches a maximum depth of 3.9 m (Collin, 2016).

From an ecological point of view, the bog is mainly dominated by *Sphagnum* genera (*S.recurvum*, *S.magellanicum* and *S.capillifolium* species), associated with herbaceous species, (*Scheuchzeria palustris*, *Eriophorum vaginatum* and *Carex rostrata*), ericaceous plants (*Andromeda polifolia*, *Vaccinium oxycoccus*, *Calluna vulgaris*) and Mountain pine (*Pinus uncinata*). This latter progressively also colonizes the center of the bog. The NE bog extension, towards the main outlet of the bog, is characterized by a birch forest highlighting a rise of the moraine elevation and then a marsh covered by *sedge* and *Molinia caerulea* (Fig. 1B; Collin, 2016). The surroundings of the bog are mainly woodlands with spruce set over mature raised peatlands.

From a hydrological point of view, the Forbonnet bog corresponds to a depression within the peatland complex, and features a natural and occasionally activated overflow reaching the main outlet located NE. At the bog's northern edge a west-east oriented stream drains the more mature part of the peatland complex and reaches the main outlet (Gauthier et al., 2019; Fig. 1B). This general SW-NE oriented water flow pattern was affected by numerous drainage ditches (Fig. 1B). This led to peat subsidence redirecting a part of the water fluxes to the south, toward the "Creux au Lard" karstic sinkhole (Calvar et al., 2018; Magnon and

Grosvernier, 2015) that has been revealed to be connected to the regional karst network (Duraffourg and Palacio, 1981).

2b) The restoration program of 2015-2016

To mitigate the effects of drainage ditches, a restoration program, funded by the European Life Peatland Program, was implemented from 2015 to 2016 (Calvar et al., 2018). This strategy aimed to restore the original watershed surface and to raise the WTD in the bog (Fig. 2).

The work was conducted in 2 phases in September-October 2015 and May-July 2016 respectively (Fig. 1C). (1) To stop water outflow and to counter the subsidence due to peat mineralization, a cofferdam of sheet piles was constructed between the Forbonnet bog and the "*Creux au lard*" sinkhole. An adjacent drainage ditch was backfilled with peat and wood panels. (2) A second ditch was partially backfilled in 2015 and completed in 2016 during the second phase. Also, in 2016 (from 23th May to 21th June), 15 wood palisades were installed perpendicularly to a 300 m long and 20 m wide NW-SE trending drainage axis in order to redirect flows from adjacent areas towards the bog. These palisades were composed of non-treated and 10 cm thick notched spruce beams. They were covered with at least 40 cm thick mounds of peat that are currently vegetated.

3. Materials and methods

3a) Meteorological monitoring

Precipitation (P, ARG100 Campbell Scientific) and Temperature (T, HMP45C Campbell Scientific), were monitored at the SNO Tourbières platform located in the central bog area since November 2008 (Fig. 1B). Due to a technical failure of the rain gauge in 2016, no precipitation data are available from 22/05 to 16/06/2016. To fill this gap, the data were reconstructed with Météo France data from three nearby weather stations at similar altitudes (Pontarlier-831 m at 15.4 km, Levier-713 m at 14.2 km and Mouthe-940 m at 13.6 km) based

on multiple linear correlations between the Forbonnet and the Météo France data before (2009 to 2015) and after (2017 to 2019) the rain gauge failure ($R^2 = 0.955$, p -value = 0.000192 for May; $R^2 = 0.921$, p -value = 0.00104 for June).

Potential Evapo-Transpiration (PET, mm) was calculated at a monthly scale using the Thornthwaite formula (Thornthwaite, 1948; Eq.1):

$$PET = 16 * F * \left(\frac{10 * T_m}{I}\right)^a \quad (\text{Eq.1})$$

$$\text{With } I = \sum \frac{T_m^{1.514}}{5} ;$$

$$a = (6.75 * 10^{-7} * I^3) - (7.71 * 10^{-5} * I^2) + (1.792 * 10^{-2} * I) + 0.49239$$

, whereas T_m is the mean monthly air temperature, F the correction factor based on latitude and I the heat index based on monthly mean temperatures.

The difference between P and PET is defined as an indicator of atmospheric water balance and subsequent water stress (Gao and Giorgi, 2008).

3b) Hydrological monitoring

The hydrological survey was based on 21 piezometers (Fig. 1B). Ten of them (TV1, 2, 3, 4, 5, 6, 11, 12, 13 and 14) were from July 2014 to October 2018 equipped with automatic HOBO U20L04 pressure probes (± 0.2 %). Two piezometers implemented on the scientific platform (SoDry and SoWet) were instrumented with OTT Orpheus Mini pressure probes (± 0.1 %; Toussaint et al., 2020b) and typify a hummock (SoDry)-hollow (SoWet) peat gradient over a distance of 10 m. To limit uncertainties due to peat shrinking and swelling (Camporese et al., 2006; Oleszczuk and Brandyk, 2008), all used piezometers were anchored in the geological substratum except TV3, TV6 and SoWet/SoDry. The error range of the piezometers of ± 1.6 cm corresponds to twice the probe accuracy furnished by manufacturers in order to include manual measurement/calibration errors.

Some piezometers were installed in panpipe configuration in order to screen specific depth levels: (1) TV1, TV2 and TV3 respectively slotted at 220-235 cm, 136-166 cm and 30-50 cm depth; (2) TV4, TV5 and TV6 respectively slotted at 230-240 cm, 87-137 cm and 30-50 cm depth. The two others panpipes systems, (3) D (80-100 cm) and C (175-195 cm), (4) A (250-270), B (100-120) and TV8 (subsurface sampling) were not instrumented but used for manual sampling and hydrochemical characterization. The water level at the bog outlet was monitored since June 2014 with an CTD pressure probe (OTT Hydromet) at a time step of 30 minutes (Fig. 1B). Outlet discharge (Q) was determined from water level using an experimentally-defined rating curve based on 11 manual discharge measurements realized from Nov. 2015 to Feb. 2021.

3c) Physico-chemical monitoring and analytical modelling

In parallel to the hydrological monitoring, Electrical Conductivity (EC) and T were monitored at the bog outlet with the above mentioned CTD probe.

In addition, manual measurements of pH, temperature and EC were carried out directly in the piezometers with a manual multi-meter probe at the depth wished (AP-2000 AQUAREAD). Vertical EC, T and pH profiles were realized manually at the following dates: 22/03/2016, 24/05/2016, 06/07/2016, 16/08/2016. Depth gradients were calculated for each profile and each parameter according to (Eq.3):

$$Gradient = \frac{(lowest\ value - highest\ value)}{(Depth\ of\ lowest\ value - Depth\ of\ highest\ value)} \quad (Eq.3)$$

For the 22/03/2016 and 06/07/2016 campaigns, physico-chemical parameters were completed with major cation analyses. Sampling was performed with a manual pump and a plastic pipe and stored in HDPE bottles. The samples were filtered at 0.45 μ m within 12h after sampling and stored at 4°C in HDPE tubes. The sample aliquots for cation analysis were acidified to pH 2 with ultrapure nitric acid. The cation and anion concentrations were

determined by ICP-AES (Thermo iCAP 6500) and HPLC (Dionex DX 100) at the PEA2^T analytical platform of the Chrono-Environnement laboratory.

Considering the strong correlation between [Ca] and EC ($R^2 = 0.98$) in our samples and the acidity of peat water (mostly $\text{pH} < 7$), we considered calcium as a conservative tracer (de Moel et al., 2013; Hart et al., 2011) allowing to evaluate water and solutes transfer processes throughout the peat column by comparing vertical [Ca] variability with a constant source diffusive equation (Eq. 4) proposed by (Crank, 1975):

$$C(x, t) = C_0 \operatorname{erfc} \left(\frac{x}{2\sqrt{D^*t}} \right) \quad (\text{Eq.4})$$

With $C(x,t)$, the calculated cation concentrations, C_0 (mg.L^{-1}) the initial and assumed constant concentration at the substratum/peat interface, x (m) the distance from this interface, t (s) the time, erfc the complementary error function and D^* the effective diffusion coefficient. D^* was determined from the molecular diffusion in solution for cations adjusted by a tortuosity factor, both respectively set at $2.10^{-9} \text{ m}^2.\text{s}^{-1}$ and 0.22 according to Freeze and Cherry (1979) and Fraser et al (2001).

4. Results

4a) Restoration effect on hydrology

A significant rise of WTD (detailed below, between 20 and 155 cm) was observed after the first restoration phase at the southern part of the bog (Fig. 3). Immediately after this first phase (14/10/2015) WTD did not change significantly because of dry weather conditions (37.6 mm of rainfall from 14/10 to 19/11/2015). WTD then started to increase from 19/11/2015 after important precipitation events (168.6 mm until the 30/11/2015) and continued to rise until June 2016, corresponding to the end of the second restoration phase. During the two following years, WTD remained stable except for two drought events in December 2016 and summer-autumn 2018. Figure 3 shows that during the drought of December 2016 (7 mm between the 24/11/2016

and the 11/01/2017) WTD dropped, but remained above pre-restoration depths. In contrast, the combined effect of heatwave and rainfall drought in summer and autumn 2018 caused a sharp decrease of WTD.

A comparison of the hydrological year 2014-2015 (01/09/2014 - 31/8/2015) and the 2 hydrological years after restoration (01/09/2016 – 31/08/2018) reveals a significant rise of WTD (Welch test, p -value = 0). Some spatial contrasts show a distance gradient from the closest parts of the backfilled ditch (TV14 +155.2 cm, TV12 +42.3 cm) to the further areas (TV13 +23.7 cm, TV11 +19.6 cm; Fig. 1C). At similar distances from the backfilled ditch, TV14 (10 m from the cofferdam) harbored a higher increase than TV12 (35 m from the cofferdam), highlighting the dam blocking effect. This distance effect decreases downstream as observed with SoWet (+0.3 cm, p -value = 0.46) and SoDry (+1.3 cm, p -value = 0.0016). Consistently, scatter plots with WTD vs. Q (Fig. 4) show that there is no break between the data before and after restoration for central piezometers, whereas a clear shift occurs for the upstream (TVs) piezometers.

The data from the panpipe piezometers (TV1/2/3; TV4/5/6) in the southern part of the bog point to hydraulic head (H) changes within the peat column (Fig. 5). The increase (p _value = 0) was stronger for the intermediate (TV2: +54.1 cm; TV5: +20.4 cm) than for the deeper (TV1: +22.5 cm; TV4: +6.6 cm) and shallower levels (TV3: +25.5 cm; TV6: + 16.1 cm), indicating that restoration mainly affected the intermediate peat layer.

Finally, at the outlet, no significant change was observed for the mean discharge before and after restoration.

4b) Temporal variability of the flow rate and electrical conductivity at the bog outlet

Comparing outlet Q and EC meteorological data (P , T , P -PET) allows to visualize the integrated response of the bog to meteorological variations (Fig. 3).

Monthly P-PET is high in winter (252 and 213 mm for Jan. and Feb. 2016, 390 and 329 mm for Jan. and Dec. 2018) as a consequence of low T ($< 3^{\circ}\text{C}$ in average) and high P (> 220 mm/month). In contrast, monthly P-PET is low or even occasionally negative in summer from June to October due to high T and low P (< 146 mm/month, except for Oct. 2019 with 228 mm). During high flow periods in winter (Nov.-May) Q is tightly related to WTD (Fig. 4). In contrast, during low-flow (June-October) Q does not follow WTD variations that temporally stand high WTD levels (Fig. 4). This general behavior is overwhelmed by Q peaks related to strong precipitation events (e.g. Nov. 2015 and Oct-Dec. 2019, Fig. 3).

In parallel, water mineralization (EC) usually presents seasonal cycles (Fig. 3). During high flow periods, EC is low ranging from 66 to 73 $\mu\text{S}\cdot\text{cm}^{-1}$ except for Nov. 2018-May 2019 with 111 $\mu\text{S}\cdot\text{cm}^{-1}$. Conversely, during baseflow (June-October) EC is usually above 100 $\mu\text{S}\cdot\text{cm}^{-1}$ and reaches peaks of 400-500 $\mu\text{S}\cdot\text{cm}^{-1}$.

Assuming EC values of 20 $\mu\text{S}\cdot\text{cm}^{-1}$ for precipitation (e.g., Bertrand et al., 2008) and 550 $\mu\text{S}\cdot\text{cm}^{-1}$ for groundwater, which corresponds to values observed for local karst springs (Fontaine Bouchard and Fontaine Carrée springs located a few kilometers from the bog, Fig. 1A), it is possible to evaluate the contribution of both endmembers to the outlet flow based on simple mixing calculations. At an average seasonal scale precipitation dominates during winter and spring (Nov.-May) with 89 ± 3 % and decreases to 72 ± 8 % for June to October. In 2017, June-October groundwater contribution was however more limited and accounted on average for only 16 %.

4c) Spatio-temporal evolution of the stratification of pore water mineralization

The raw physico-chemical data are available in the supplementary material. Distinct trends can be observed for EC, which allows to separate the sampling points into 3 groups (Fig. 6):

- (1) The lowest mineralization were found in the piezometers TV7, TV10, TV13, BM18 and BM19. Their mean EC was $39 \mu\text{S}\cdot\text{cm}^{-1}$, reaching a maximum of $95 \mu\text{S}\cdot\text{cm}^{-1}$ for BM18. Vertical EC gradients were below $30 \mu\text{S}\cdot\text{cm}^{-1}\cdot\text{m}^{-1}$ with Maximum Depth Measurement (MDM; vertical distance between sampling point and moraine layer) of 0.7, < 0.05, < 0.05, 1.37 and 0.58 m respectively.
- (2) The greatest EC ($> 500 \mu\text{S}\cdot\text{cm}^{-1}$, values close to the EC of local karst springs) were observed in TV1, TV4, C, TV8, TV12. Consistently, these piezometers present the greatest vertical EC gradients of about $150 \mu\text{S}\cdot\text{cm}^{-1}\cdot\text{m}^{-1}$ with low MDM (respectively 0.075, <0.05, 0.35, 0.39 and <0.05 m).
- (3) The other piezometers (TV9, 11 and 14) present different intermediate EC values ranging from 16 to $328 \mu\text{S}\cdot\text{cm}^{-1}$ with a mean of $111 \mu\text{S}\cdot\text{cm}^{-1}$ and EC gradients below $130 \mu\text{S}\cdot\text{cm}^{-1}\cdot\text{m}^{-1}$. MDM are low for TV9, TV11 and TV14 (<0.1 m) and greater for E (1.11 m).

From a temporal point of view, TV9, 11 and 14 present the greatest EC gradient change (standard deviation of 53.4, 37.2 and 49) while the standard deviation median for all piezometers EC gradients is 12.3. In particular, in the deepest part of TV9 EC reached $318 \mu\text{S}\cdot\text{cm}^{-1}$ the 22/03/2016 whereas it did not exceed $69 \mu\text{S}\cdot\text{cm}^{-1}$ in May, July and August. A contrasted pattern was observed for TV11 with a slight and continuous EC increase at its bottom from March to August (130, 114, 168 and $328 \mu\text{S}\cdot\text{cm}^{-1}$).

Modelling a diffusive solute transfer (Eq. 4) may allow assessing recharge-discharge processes on the basis of the Crank's diffusion model (1975). Assuming that the cation exchange capacity remains limited for long-term processes due to saturated exchange sites, concentrations below modeled values imply an advective component in the solute transport. We applied this model for a 3 m thick peat layer with various steady-state time-lapses (from 1000 to 7000 years) in agreement with ^{14}C ages from previous studies which found $910 \pm$

30 years before present (BP) at 2.1 m depth in the southern part of the bog (Goubet, 2016) and 3755 ± 35 and 5665 ± 35 years BP for peat at 2.15 m and 3.2 m depth in the woody/raised surrounding areas to the west of the bog (Gauthier et al., 2019). Despite these uncertainties, the diffusion model highlights that most of the Ca-profiles do not match with purely diffusive transfer of Ca in an immobile catotelmic pore water (Fig. 7) and that advective solute transport is required to explain the observed profiles (Fraser et al., 2001).

5. Discussion

5a) Restoration highlights lateral connectivity with surrounding raised peatlands

The impact of restoration on WTD was much stronger in the SW part of the bog, which is consistent with the apparition of a flooded area at the junction of the cofferdam and a backfilled drain (Figs. 1 and 4). Similarly to other sites (D'Acunha et al., 2018; Holden et al., 2011; Ketcheson and Price, 2011), the restoration effect shows a distance gradient with a decreasing WTD rise to the NE. The observed WTD rise could be consistent with (1) the increase of the watershed surface that increased water volume stored in previously drained peat layers and (2) with a greater water storage capacity of peat resulting from so-called “bog breathing” (e.g., Howie and Hebda, 2018) discussed below and/or accumulation of a new mosses layer.

The increase of the watershed reported by Magnon and Grosvernier (2015; Fig. 2) corresponds to a new supply area of about 8 ha. This lateral connectivity favored by watershed increase could be related to the commonly observed anisotropy of the hydraulic conductivity (K) in peatlands (Fraser et al., 2001; Levison et al., 2014). Beckwith et al. (2003) found in a laboratory study that $K_{\text{horizontal}}$ was 78 % greater than K_{vertical} , with a mean anisotropy K_h/K_v of 3.5.

Due to the observed rapid WTD rise, it is likely that the first (short-term) consequence of restoration was drowning of previously emerged peatland. This additional water could then have had 3 different fates: (1) It could have been evaporated, because a higher WTD would

favor evapotranspiration despite the swelling of peat, which may partly compensate this trend.

(2) The additional water volume may lead to higher outlet discharge or (3) infiltrate toward the underlying karst aquifer.

“Bog breathing” corresponds to peat shrinking and swelling in relation to WTD changes (Howie and Hebda, 2018), resulting in fluctuating water storage capacities and bulk peat densities. Increasing water storage capacity should lead to decreasing outlet discharge for a watershed of constant size. However, at the Forbonnet bog, greater discharge rates were observed after restoration for similar precipitation events (Fig. 3), which is opposite to what has been observed for "bog breathing" elsewhere (Ballard et al., 2012; Wilson et al., 2010). Considering that restoration added 8 ha to the watershed, it is possible that the greater impluvium completely overprinted a possible bog breathing effect. In fact, previous studies have shown that bog breathing could lift up the peat surface from about 1.5 cm (Morton and Heinemeyer 2019) to 20 cm and take several years especially in drained bogs (Howie and Hebda, 2018). Accumulation of new mosses could reach 20 cm, but 10 years after restoration (McCarter and Price, 2013), or 11 cm 20 years after restoration (Ahmad et al., 2020). All these arguments suggest that increase of peat capacity storage (bog breathing and/or accumulation of new mosses), although being a potentially significant mechanism, could remain limited at the current stage in comparison to rapid storage increase due to watershed enlargement.

In contrast, our observations are in accordance with studies illustrating that a water level closer to the surface favors water transfer to the outlet (Bernard-Jannin et al., 2018) by connecting microtopographic heterogeneities such as pool-ridge systems (Oosterwoud et al., 2017). This raises the necessity to address possible WTD threshold effects, especially during superficial flooding as discussed in section 5b.

5b) Superficial acrotelmic horizontal transfers

Consistently, and considering there is no evidence for a significant storage capacity increase after restoration, the complex daily relationship between Q and WTD variations (Fig. 4) implies the existence of complex seasonal mechanism which could be located in the upper part of the peat column, i.e. hydraulically corresponding to the acrotelm. At the whole ecosystem scale, this compartment is recognized to favor (1) superficial horizontal flows in particular during stormflow in wet periods (Holden and Burt, 2003), (2) a high water storage capacity after desaturation, due to greater porosity in this layer (Boelter, 1965; Oosterwoud et al., 2017) and (3) therefore control the complex discharge behavior depending on preceding hydrological conditions (Wilson et al., 2010). For the Forbonnet bog, Bertrand et al. (2021) identified a threshold depth of -19 cm near the central part of the bog consistent with the acrotelm/catotelm boundary and provides an estimation of the upper reservoir dimension.

5c) Diffusive-advective processes between 3 compartments of the peat column

From a vertical point of view, the pressure gradient reversals observed after restoration in the panpipe piezometers near the restored dam ($H_{TV2} > H_{TV1}$ and $H_{TV2} > H_{TV3}$), suggest that restoration had preferentially reactivated lateral inflows at intermediate depth and that the intermediate layer (TV2) potentially irrigates the over- and underlying peat layers. However, this piezometric head reversal is not observed for the second panpipe system further downstream (TV4/5/6). This suggests a possible equilibration of the vertical piezometric head gradients by upward and downward fluxes along the general flow path thanks to hydraulic connectivity between peat layers.

Therefore, our results illustrate that backfilling of drains during peatland restoration not only enhances the lateral hydraulic continuity between surrounding raised bogs and less mature peatland, but locally also introduces vertical water fluxes across at least 3 peat layers. This contrasts with the common diplotelmic model, relevant at the ecosystem scale (section 4a), that

describes peatlands as a two-compartment ecosystems with poorly interconnected acrotelm and catotelm layers due to contrasted physical characteristics (Boelter, 1965; Ingram, 1978).

Consistently, the vertical physico-chemical variability also enables to describe 3 water compartments in the peat profile: (1) a low mineralized water compartment of the uppermost ombrogenic reservoir and corresponding to the hydrologically reactive acrotelm described previously; (2) a minerotrophic compartment near the substratum and (3) a transitional compartment with moderately mineralized water (group 3). The analytical modeling of vertical diffusive cation transfer (Fig. 7, section 4c) argues for the existence of advective transport (horizontal and/or vertical) between these compartments. Rapid EC gradient changes observed in 2016 for the TV9, 11, 12, 14 and DC piezometers (standard deviation $> 33 \mu\text{S}\cdot\text{cm}^{-1}\cdot\text{m}^{-1}$; section 4c; Fig. 6) and a strong concentration change observed for Ca between March and July in piezometers A and B (Fig. 7), are similarly consistent with advective upward fluxes from the basement. Siegel et al. (1995) suggested that even with low permeabilities for the deeper peat layers, a macro scale porosity could locally favor advective transfers. Upward and downward transfers between groundwater and peat columns, with rapid (ten days scale; Fraser et al., 2001) or slower (several years; Siegel and Glaser, 1987) dynamics have already been identified in other peatland ecosystems and lead to questions about possible interactions with the underlying karst aquifer as discussed in the following.

5d) Interactions with the regional karst aquifer

Dilution and concentration effects are common in aquifers due to the variability of rainfall and evapotranspiration (Caetano Bicalho et al., 2012; Jenkins, 1989; Wang et al., 2020). In the Jura Mountains the lack of precipitation during summer and autumn 2018 was historical (LE BARBU et al., 2019) leading to a continuous increase of EC at the bog outlet (Fig. 3). Evapotranspiration alone cannot explain the consecutive EC peaks at the outlet nor the very high EC values of $400\text{-}500 \mu\text{S}\cdot\text{cm}^{-1}$ in bog pore waters. For instance, to shift from 100 to 400

$\mu\text{S}\cdot\text{cm}^{-1}$ only by evapotranspiration, it would be necessary to evaporate 3/4 of the bogs water volume. However, high EC values are consistent with water inputs from the morainic and calcareous basement which are lithologies featuring mineralized pore waters (Drexler et al., 1999; Griffiths and Sebestyen, 2016; Siegel et al., 1995; Siegel and Glaser, 1987; Steinmann and Shotyk, 1997). Considering advective processes delineated in section 5c, water-rock-interaction of vertically or laterally fed water could explain these observations.

At the outlet, although summer (June-Oct.) P-PET seems to slightly (although not significantly, $R^2=0.14$, $p\text{-value}=0.472$; Fig. 8B) induce a dilution of the summer outflow mineralization, it is rather the previous winter (Nov.-May) P-PET which is correlated to summer EC ($R^2=0.91$, $p\text{-value}=0.003$; Fig. 8A). Although summer rainfall remain the major driver of summer discharge ($R^2=0.76$, $p\text{-value}=0.024$; Fig. 8C and D), seasonal groundwater contribution is higher in summer ($28 \pm 8 \%$) than in winter (17.1 %), similarly to values observed by Santoni et al. (2021) in a groundwater-dependent peatland. These elements argue for a mineralized pool of water renewed and stored in the peatland in winter and then transferred to the outlet in summer.

Although the hydrogeological context is favorable and statistical clues argue for these interactions, seasonal change of head pressure gradient as observed elsewhere (Glaser et al., 1997) are not obvious from the WTD evolutions observed in the panpipe piezometers (TV1/2/3 and 4/5/6; Fig. 4). They rather argue for concurrent head pressure changes in the 3 peat layers throughout the hydrological year. This apparent contradiction is however consistent with the mosaic-like hydroecological organization of peatlands as suggested by Baird et al. (2016) and also illustrated by the spatio-temporal variability of the EC profiles. Groundwater inflow could be punctual through pockmarks (Reusch et al., 2015) that could explain the contrast between geochemical and head measurement observations. Such pockmarks between surface water bodies and underlying karst aquifers are known for the Bouverans lake located only 1.8 km

east of the Forbonnet bog (Bichet et al., 2014). Moreover, at the peat basement, while glacial deposits low permeability favor water accumulation as a prerequisite for peat accumulation, this feature does not preclude vertical water flows between regional bedrock and peatland ecosystems as it is identified by Kværner and Snilsberg (2011).

Different areas might therefore be dominated by specific flows depending on the season: (1) downward flows from the surface by rainwater infiltration, (2) upward flows from the substratum by groundwater inputs and (3) lateral drainage flows from surrounding area near the bog's edges. But at the scale of the ecosystem, the relative contributions of these flows vary according to the general hydrological state of the area. This mosaic-like organization of interactions with the regional aquifer, lateral flows and vertical inputs agrees with the hot/cold spot concept developed by Morris et al. (2011), corresponding respectively to spots of rapid or slow water transfers. Our study illustrates this concept by extrapolating the model at the hydrological (surrounding raised peatland) and hydrogeological (regional karstic aquifer) scales and highlights the need to account for these scales in a management perspective (Kværner and Snilsberg, 2011).

6. Conclusion

By combining the evaluation of restoration effects, long-term outlet discharge/conductivity monitoring and vertical pore water profiles it was possible to establish a nested conceptual model for the hydrological functioning of the Forbonnet bog which is entirely integrated in the local/regional hydro-geomorphological context (Fig. 9). Along the vertical axis, three peats layers were identified based on EC and WTD profiles. Each of them is potentially linked to a major source of water: the upperlayer supplied by rainfall; the deeper by groundwater and the intermediate by an advectively controlled vertical mixture between these two end-members and by lateral seepage from surrounding raised bogs, which constitute powerful hydraulic buffer zones as already observed elsewhere (Howie and Meerveld, 2011).

This also illustrates that peatland ecohydrological models may benefit from restoration program analysis, and that managers/researcher need to consider different scales in studying hydrological responses due to restoration or meteorological changes.

Therefore, while this study shows that at the ecosystem scale the diplotemic model appears at the first insight hydrologically relevant at the ecosystem scale, at finer scale, spatio-temporal heterogeneities of water origins and water/solute transfers rather require a 3-compartment model implying vertical meteoric inputs, local lateral flows, and punctual regional upward supply. The proportion of these 3 nested scales of water supply, change throughout the mosaic-like organization of the Forbonnet bog and impact the discharge quality and quantity from seasonal to inter-annual timescale.

Over the long-term, such a nested model should therefore be useful to evaluate and predict further climate change impacts that are still difficult to model mechanistically in such mosaic-like ecosystems (Bertrand et al., 2021). The predicted increase of winter precipitation in northern mid-latitudes (Brulebois et al., 2015; Shulka et al., 2019) could further imply a greater proportion of delayed regional groundwater in the coming decades, especially during vegetation development in summer. Conversely, temperature increase at mid-latitude/altitude mountains like Jura Mountains, could impact the dynamics of this contribution by limiting temporal water storage in snowpack (Meeks et al., 2017).

At the ecosystem scale, stronger overflow could take place in winter due to higher precipitations and lateral flows from the surrounding peatlands. Actually, 2 years after restoration, there is no evidence for an increasing peat capacity storage. In this perspective, more time is needed to assess both a regeneration of bog breathing and the formation of new mosses layers that could increase water storage and limit overflow in winter or buffer the impact of drought events which are predicted to increase (Howie and Hebda, 2018; Shulka et al., 2019).

Acknowledgements

The authors warmly thank the managers of the Regional Natural reserve of Frasne-Bouverans for allowing access to the site. This study is part of the CRITICAL PEAT project funded by the Bourgogne Franche-Comté Region, is supported by the OSU THETA of the Bourgogne Franche-Comté University, the SNO Tourbières (<https://www.sno-tourbieres.cnrs.fr/>) and the French network of critical zone observatories (OZCAR) Network (<https://www.ozcar-ri.org/fr/ozcar-observatoires-de-la-zone-critique-applications-et-recherche/>). The SNO Tourbières observing system was set up thanks to an incentive funding of the French Ministry of Research that allowed pooling together various pre-existing small-scale observation setups. The continuity of the field monitoring was made possible by a continuous CNRS-INSU funding since 2008. A. Lhosmot benefits from a Ph.D. fellowship of the French Ministry of Research. The authors would like to thank Ch. Loup, N. Crini and C. Amiot from the PEA²T analytical platform of the Chrono-Environment laboratory for chemical analyses. M. Nourdin, V. Bichet and P. Grosvernier are also thanked for their personal contributions. Finally, the authors acknowledge the two anonymous reviewers that strongly helped to improve the quality of this work.

Data Availability Statement

The meteorological data (P, T) that support the findings of this study are openly available in Toussaint et al. (2020a) at <https://doi.org/10.5281/ZENODO.3763342>. Central bog piezometer data (SoDry and SoWet, Fig. 4) are openly available in Toussaint et al. (2020b) at <https://doi.org/10.5281/ZENODO.3763766>. These data were derived from the following resources available in the public area: <https://www.sno-tourbieres.cnrs.fr/>. Peat pore water geochemical profiles (Fig. 6, pH, electrical conductivity, depth of measurement) are available in the supplementary material of this article. Other piezometric data (Fig. 5), outlet monitoring

(Fig. 3 and 8) and calcium concentration data (Fig. 7) are available on request to the corresponding author.

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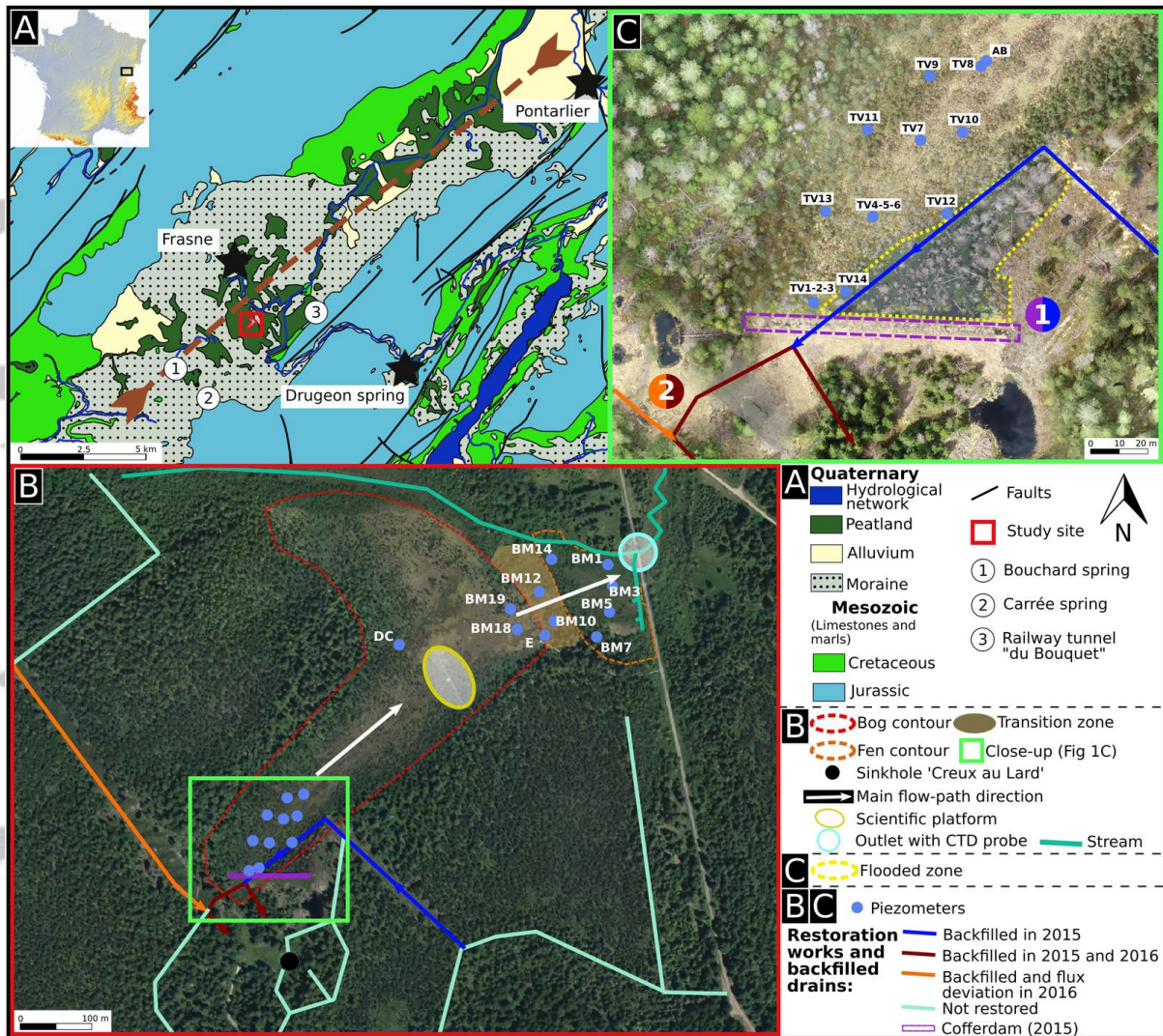


Figure 1. Maps of the study site at different scales: (A) geological map (base map modified from BRGM) and site location; (B) site instrumentation, drainage ditches and restoration work (base map IGN, 2018) and (C) focus on the southern part of the bog highlighting the flooded area (base map Sopreco, 2019).

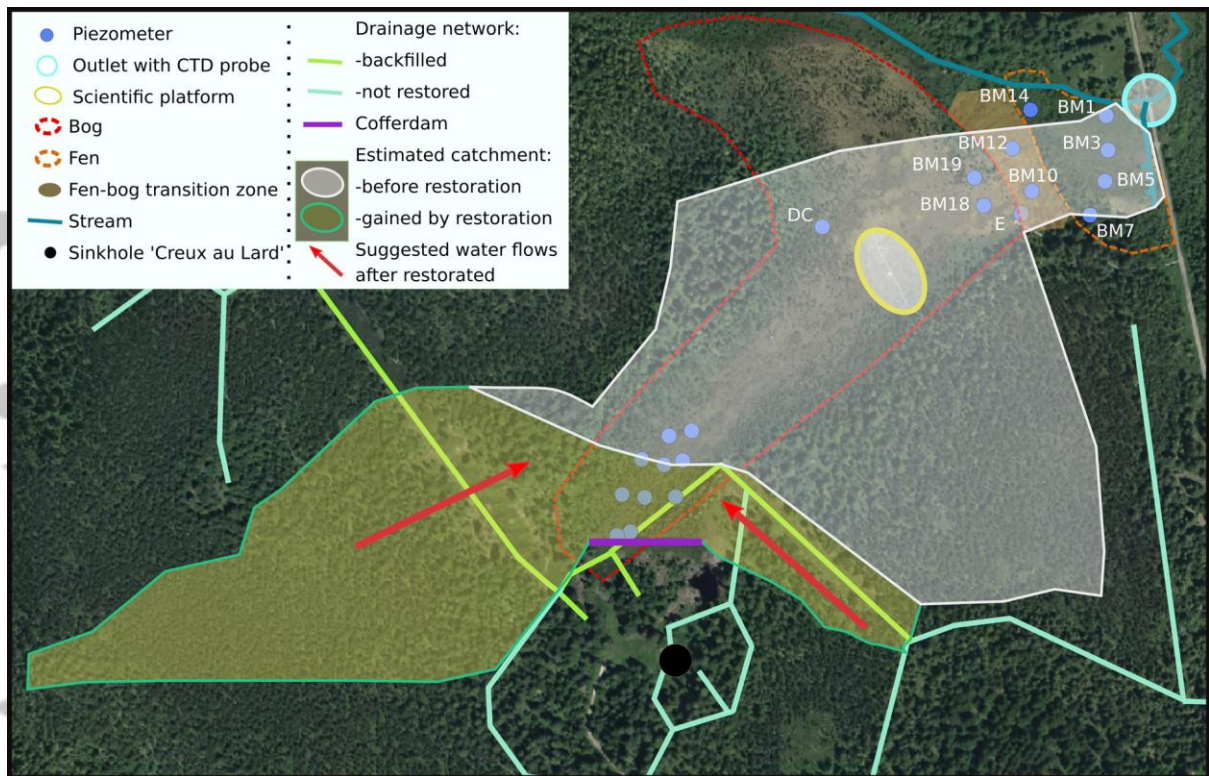


Figure 2. Gained watershed surface and reactivated water flows after restoration (Magnon and Grosvernier, (2015); Collin 2016).

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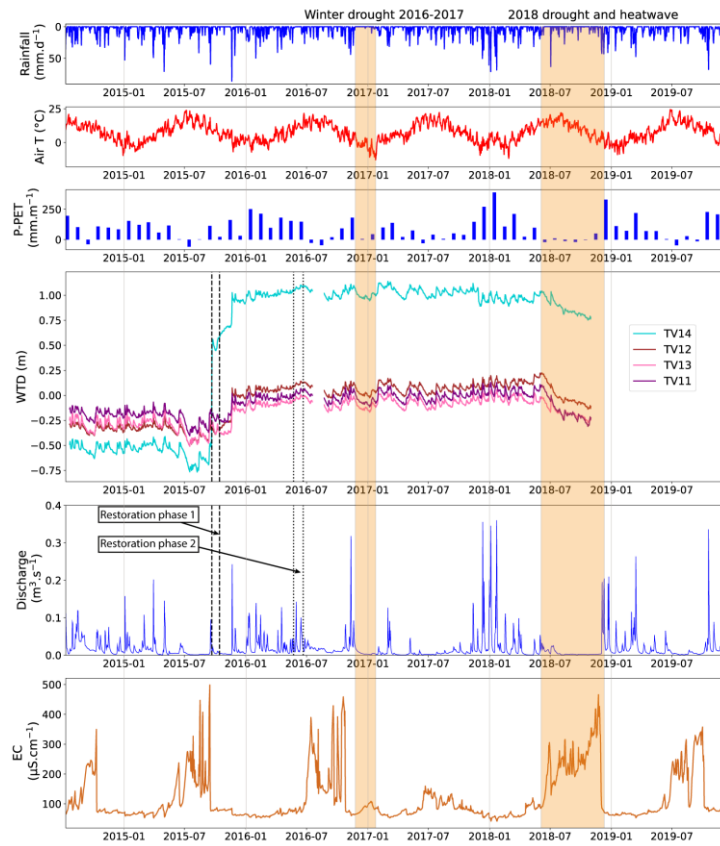


Figure 3. Meteorological and hydrological data over the Forbonnet bog during the study period (2014-2018). Meteorological data are available in Toussaint et al. (2020). Piezometer data are provided by local site managers. Dashed line (from 21th September to 14th October 2015) and dot line (from 23th May to 21th June 2016) represent the two restoration phases.

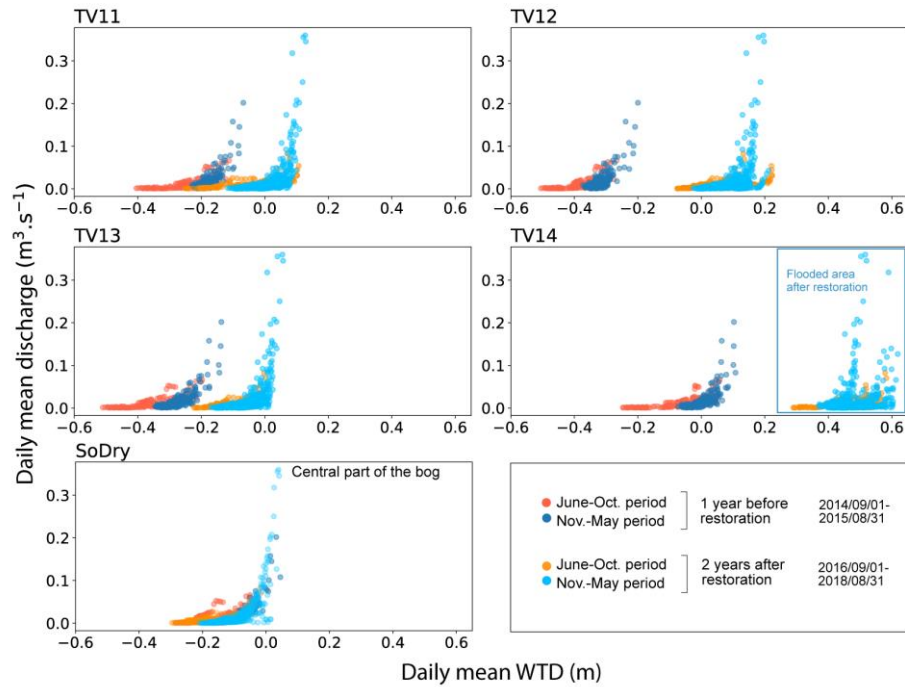


Figure 4. Scatter plots showing daily mean discharge ($\text{m}^3 \cdot \text{s}^{-1}$) function of daily mean WTD (m). For each piezometer, colors highlight temporal period (before/after restoration and June-Oct./Nov.-May periods).

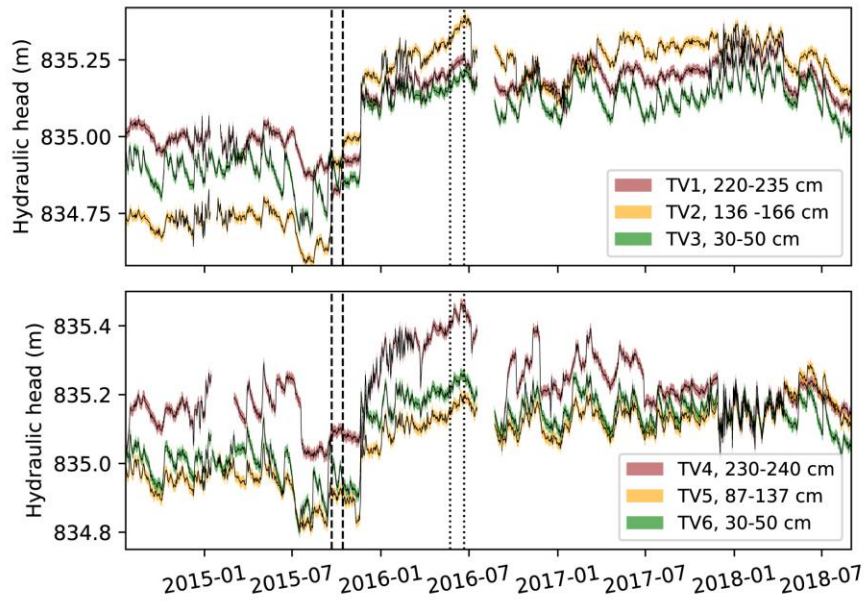


Figure 5. Hydraulic heads in nested panpipes piezometers during the study period (2014-2018). Edge color for curves indicates data uncertainty (twice the accuracy of the sensor). Legend gives slotted depths relative to the peat surface, for each piezometer. Dashed line (from 21th September to 14th October 2015) and dot line (from 23th May to 21th June 2016) represent the two periods of restoration.

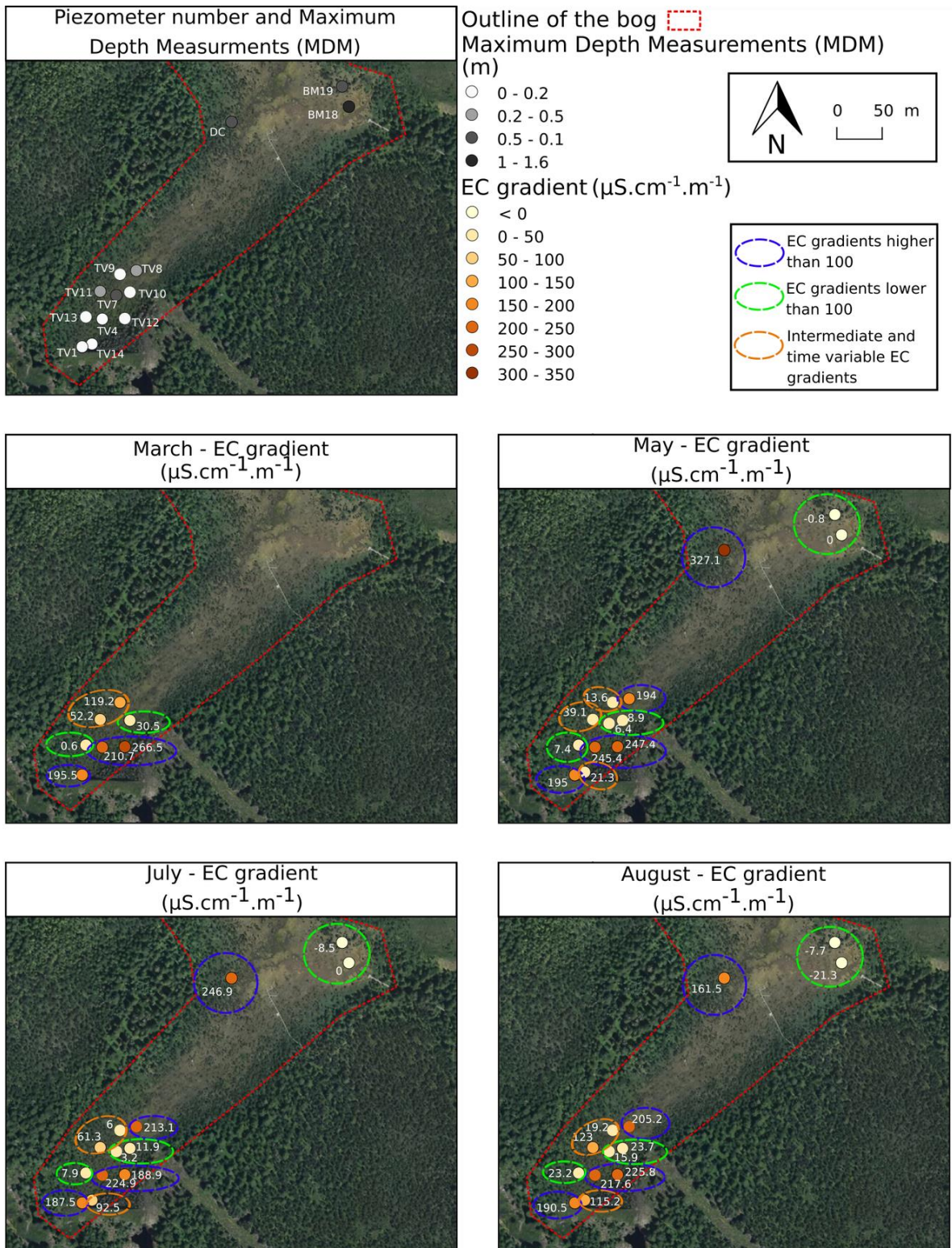


Figure 6. Piezometers EC gradient map for March, May, July and August 2016. Maximum Depth Measurements (MDM) correspond to the vertical distance between piezometer and underlying moraine cover.

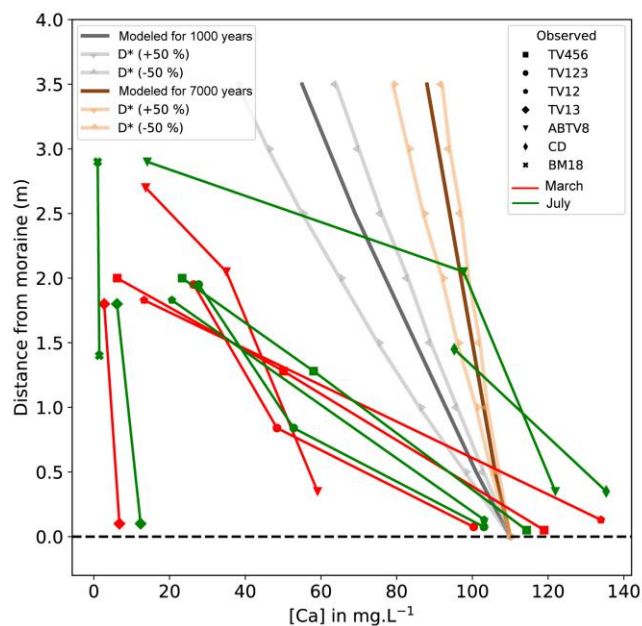


Figure 7. Comparison of modeled and observed Ca concentrations in catotelmic peat with distance from moraine. The modeled concentrations are based on pure solute transfer of Ca from the moraine. The observed values decrease much faster with distance than the calculated values, suggesting the presence of an advective transport.

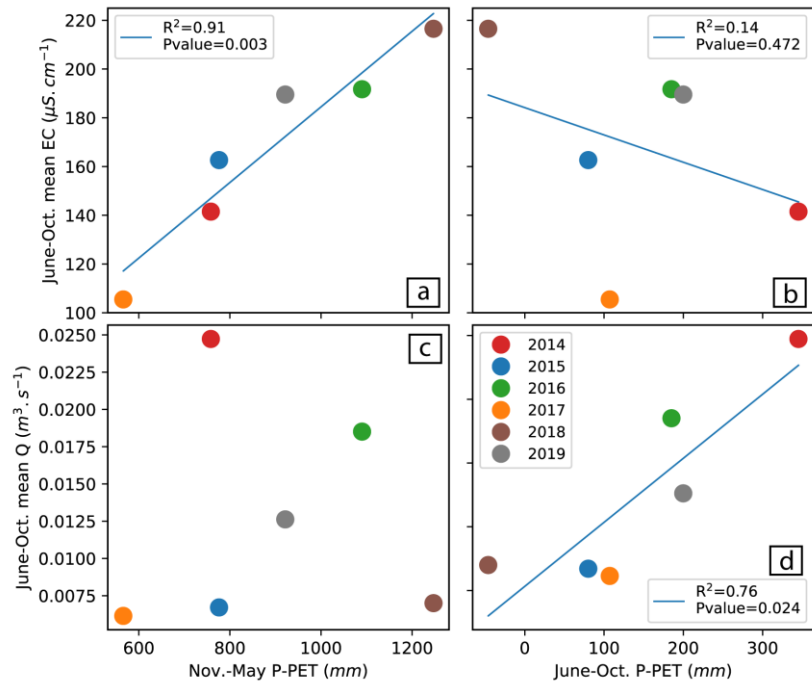


Figure 8. (A) summer (June-Oct.) mean Electrical Conductivity (EC) function of winter (Nov.-May) recharge (P-PET); (B) summer mean EC function of summer recharge; (C) summer discharge function of winter recharge and (D) summer discharge function of summer recharge.

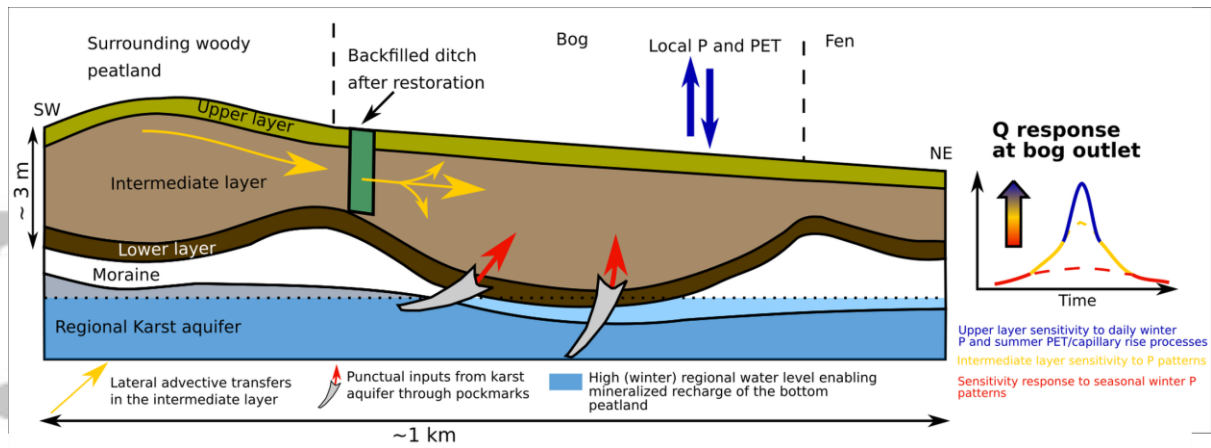


Figure 9. Conceptual scheme of the nested hydrological system of the Forbonnet bog.

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