PhD thesis
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Surface water - groundwater interactions at different spatial and temporal scales

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Preface

This thesis is the result of a three-year PhD programme carried out at the Department of Geosciences and Natural Resource Management, Faculty of Science, University of Copenhagen under the supervision of Associate Professor Peter Engesgaard. Eva Bøgh, Associate Professor at the Department of Environmental, Social and Spatial Change, Roskilde University acted as an external supervisor.

In accordance with the guidelines given by the Faculty of Science, University of Copenhagen the thesis comprises a summary as well as the following papers:

I/ Sebok E, C. Duque, P. Engesgaard and E. Bøgh. Spatial variability in streambed hydraulic conductivity of contrasting stream morphologies: channel bend and straight channel. Manuscript ready for submission

II/ Sebok E, C. Duque, J. Kazmierczak, P. Engesgaard, B. Nilsson, S. Karan and M. Frandsen. High-resolution Distributed Temperature Sensing to detect seasonal groundwater discharge into Lake Væng, Denmark. Accepted for publication at Water Resources Research

III/ Sebok E, C. Duque, P. Engesgaard and E. Bøgh. Application of Distributed Temperature Sensing for a coupled mapping of sedimentation processes and spatiotemporal variability of groundwater discharge in soft-bedded streams. Under review at Hydrological Processes


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Abstract

As there is a growing demand for the protection and optimal management of both the surface water and groundwater resources, the understanding of their exchange processes is of great importance. In the meanwhile surface water and groundwater interact on various spatial and temporal scales, thus making the accurate characterisation of these complex exchange processes difficult. This PhD study aims at describing this natural spatial and temporal variability at several spatial and temporal scales, mainly exploring the use of Distributed Temperature Sensing (DTS) in lowland catchments. For this reason field studies were carried out at different spatial scales and over different seasons in a stream and a lake in Denmark.

Spatial variability in groundwater discharge can possibly be related to the natural heterogeneity of streambed/lakebed properties which was quantified in a small, 6x5 m long stream stretch. A wide range of vertical hydraulic conductivities were observed even on this small spatial scale also showing variability across the stream as well as between the seasons. A similarly large spatial variability in groundwater discharge was observed in a 150 m² lake section with vertical lakebed temperature profiles and seepage meter-based flux estimates. This spatial heterogeneity in groundwater discharge is also reflected in Sediment-Water Interface (SWI) temperatures that can be mapped by DTS with high spatial and temporal resolution. Correspondingly, DTS deployed on the SWI in a 70x5 m stream section detected great variability in temperatures both across and along the stream, thereby indicating a heterogeneous distribution of groundwater discharge. DTS measurements on the kilometer scale at the SWI of the stream also revealed a heterogeneous discharge pattern with several concentrated groundwater discharge sites. This variability was confirmed by differential gauging and groundwater fluxes estimated by vertical streambed temperature profiles. Catchment scale variability in groundwater discharge was detected by δ²H and electrical conductivity-based hydrograph separation.

Seasonal variability in discharge, as observed in the lake, was mostly related to changes in hydraulic heads, while in the soft-bedded stream, temporal variability can also be attributed to the changes in streambed morphology.

The temperature contrast between streams and groundwater is low, 6-8 °C at maximum, in the investigated stream catchment, thus groundwater discharge rarely results
in large changes in SWI temperatures. For this reason DTS data were combined with vertical temperature profile, and in the lake also seepage meter-based fluxes to quantify groundwater discharge and to assess the performance of DTS. The combination of the different measurements show that DTS was capable of detecting major discharge sites, but due to its spatial averaging interval of one meter, it may mask small-scale heterogeneity in discharge where the traditional point-scale flux estimates give more accurate information about its spatial distribution. Lowland streams can also have soft, mobile streambeds composed of glacial sandy deposits which were shown to influence DTS data by sedimentation and scouring processes. A new methodology was therefore developed for the long-term monitoring of surface water-groundwater exchanges in soft-bedded streams.
Dansk Resume

Da der er et voksende krav om beskyttelse og optimal forvaltning af både grundvands og overfladevands ressourcer, har en grundig forståelse af deres interaktion stor betydning. Imidlertid sker interaktionen imellem grundvand og overfladevand gennem forskellige rumlige og tidslige skalaer, derfor er den præcise karakterisering af denne udveksling vanskelig. Formålet med denne PhD afhandling er at beskrive den rumlige og tidslige variabiliteten af denne interaktion primært ved brug af Distributed Temperature Sensing (DTS) i lavlandede oplande. Derfor er der udført feltstudier på forskellige rumlige skalaer og årstider i et dansk vandløb og sø.

Rumlig variabilitet i grundvandsudstrømning er forbundet med heterogenitet af vandløbsbundens materialer, hvilket blev undersøgt i en 6x5 m vandløbsstrækning. Variation i vertikale hydraulisk ledningsevne blev observeret på tværs af vandløbet og imellem de forskellige sæsoner. Grundvandsudstrømningen viste ligeledes stor rumlig variabilitet over et 150 m² udsnit af en søbund ud fra af målinger af direkte udstrømning udført med seepage meter og vurdering af udstrømningsrater baseret på temperaturprofiler målt under søbunden. Denne rumlige variabilitet i grundvandsudstrømning blev også observeret i temperaturfordelingen i overgangen mellem sediment og vand (Sediment-Water Interface, SWI) og kortlagt med højopløselig DTS målinger. Ligeledes indikerede DTS målinger på SWI i en vandløbsstrækning på 70x5 m stor variabilitet i temperaturfordeling på tværs og på langs af vandløbet og viste dermed en lignende fordeling af grundvandsudstrømningen. DTS målinger på SWI på kilometer skala afslørede også heterogene udstrømningsmønstre med flere koncentrerede udstrømningssteder. Dette resultat blev endvidere bekræftet af målinger af variation i vandføring og udstrømningsrater udført ved hjælp af temperatur profiler under vandløbsbunden. Hydrografseparation baseret på δ²H og elektrisk konduktivitetsevne viste også rumlig variabilitet i udstrømningen på oplandsskala.

Forskell i grundvandsudstrømning henover årstider som observeret i søen var primært relateret til forandringer af det hydrauliske trykniveau, mens det i vandløbet med mobile sedimenter på vandløbsbunden også forventes at være relateret til forandringer af morfologien i vandløbets bund.

Temperaturkontrasten imellem grundvand og vandløb i det undersøgte vandløbs opland er maksimalt 6-8 °C, derfor forårsager grundvandsudstrømning sjældent store
forskelle i SWI temperaturer i vandløb. På grund af det blev DTS målinger sammenlignet med vertikale temperaturprofiler og i søen også seepage meter fluxe for at være i stand til at kvantificere mængden af grundvandsudstrømningen. DTS metoden var i stand til at kortlægge de signifikante udstrømningsteder, men som følge af den midling af temperaturerne der sker over 1 m er det muligt at DTS maskerer småskala variabiliteten i SWI temperaturer og dermed også udstrømnings fluxe hvorimod de traditionelle punktmålinger sikrer en mere detaljeret beskrivelse af koncentrerede udstrømingszoner. Det blev endvidere observeret at de mobile smeltvandsnedløbesedimenter der findes i lavlandede vandløbe har inflydelse på DTS målinger udført på vandløbsbunden. Derfor er en ny metode blevet udviklet til gennemførelse af langsigtet monitering af udvekslingen imellem grundvand og overfladevand i vandløb med mobile vandløbsbunde.
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Chapter 1

Introduction

1.1 Surface water-groundwater interactions

Since the past decades there is a growing demand to adequately characterise the transport of nutrients and pollutants to the surface water bodies in order to optimize water management and reach a good ecological state of waters as prescribed by the Water Framework Directive. As the surface water and groundwater are connected, these transport processes can also happen via groundwater flow paths, reaching the surface water bodies through discharging groundwater. On the other hand groundwater can also be contaminated from loosing surface water bodies. The understanding of surface water-groundwater interactions is therefore of great importance.

The interaction of surface water bodies with groundwater is controlled by the position of these water bodies compared to the groundwater flow systems, geological characteristics of their beds and their climatic settings (Winter, 1999). Next to the directionality of these interactions, their quantification (Kalbus et al., 2006) has also been in the focus of research in recent years, as the quantification of fluxes can lead to more elaborate and precise water budgets of our water resources (Kidmose et al., 2013), thus also to a more considerate use of our water supplies.

These interactions are not only important from the water management perspective, but also have enormous ecological influence through numerous physical, chemical and biological processes at the sediment-water interface as well as the hyporheic zone (Triska et al., 1993; Brunke and Gonser, 1997; Dahm et al., 1998). For instance riverbed clogging can physically filter small particulates and bacteria at induced infiltration sites (Schubert, 2002) and numerous studies showed nitrate removal in the hyporheic zone (Pinay et al., 2008; Zarnetske et al., 2011).

At the same time surface water-groundwater exchange is both spatially and temporally variable, with spatial variability mostly controlled by spatial variability in
streambed/lakebed hydraulic conductivity (Kalbus et al., 2009) and temporal variability governed by the fluctuations of the groundwater table and thus recharge (Winter, 1981). Thus it is a difficult task to thoroughly characterise surface water-groundwater interactions, and special investigation methods are required.

Ideally these exchanges should not only be mapped over large areas in the surface water bodies, but also quantified at several spatial and temporal scales. At the same time methods for the quantitative description (seepage meters, heat as a tracer) only yield point-scale information of groundwater fluxes. With the exception of temperature time series analysis or automated seepage meters coupled with flow meters, these measurements also lack the fine scale temporal aspect. Distributed Temperature Sensing (Selker et al., 2006b; Tyler et al., 2009) on the other hand is capable of mapping surface water-groundwater interactions over large areas and different time scales, but next to detecting spatial and temporal variability in discharge, the quantification of groundwater fluxes to the surface water bodies is rarely possible. On the catchment scale, hydrograph separation based on environmental tracers can give an approximation of the net fluxes, while providing information of a much coarser spatial resolution.

As in Denmark the surface water bodies are mostly groundwater-dominated, the understanding of groundwater pathways, the corresponding discharge zones and the net fluxes is crucial for an optimal management of water resources. For this reason the PhD research was carried out in a stream and in a lake with the aim of detecting surface water-groundwater interactions at various spatial and temporal scales with different methods.

1.2 Motivation and objectives

Distributed Temperature Sensing (DTS) is a recently developed hydrological tool using a fiber optic cable as a sensor deployed at the Sediment-Water Interface (SWI) of surface water bodies to monitor variability in SWI temperatures, thus detecting groundwater discharge using heat as a tracer. The technique can provide temperature measurements down to a sampling interval of 0.25 m, with the extent of the mapped areas only limited by the instrument specifications and the length of the applied fiber optic cable.

So far DTS has been mostly deployed in low-discharge, hard-bedded streams, laying out the fiber optic cable several hundred meters along the stream, mapping the spatial distribution of SWI temperatures. Although this mapping of SWI temperatures can also be carried out in lake environments, DTS has never been applied in lakes with the purpose of detecting groundwater discharge. Consequently its use in lake environments, in the quasi-motionless water body is so far not evaluated.
1. Introduction

In the stream studies groundwater discharge to the stream was quantified with a mixing equation based on step changes in streambed temperatures and differential gauging upstream and downstream of the detected discharge site (Selker et al., 2006b; Briggs et al., 2011). This method however can be successfully applied only if the temperature contrast between the surface water and groundwater is large enough and the stream discharge is relatively low, otherwise a temperature equilibrium between the surface water and the discharging groundwater is immediately achieved with only a slight change in SWI temperatures as was observed in lowland streams (Krause et al., 2012; Karthikeyan et al., 2012).

The quantification of groundwater discharge to surface water bodies based on the step changes in SWI temperatures detected by the DTS was usually carried out by the combination of the DTS technique with differential gauging as an independent method to quantify groundwater discharge. On the other hand there are only a few examples (Lowry et al., 2007; Briggs et al., 2012) where DTS measurements were combined with other methods, especially point flux measurements of groundwater discharge. Thus the limitations and capabilities of DTS compared to other point-scale flux-estimation methods were not yet assessed.

For this reason the PhD research was centered around the use of DTS in lowland catchments in Denmark where the temperature contrast between the groundwater (8°C yearly average) and surface water (between 1-23°C in the lake in the study period and an annual variation between 2-16°C in the investigated stream) is low during most of the year and streams are usually soft-bedded due to glacial sand deposits. The objectives of the present PhD research were:

1. to investigate the small-scale natural spatial and temporal heterogeneity of the streambeds/lakebeds,

2. to develop new monitoring methodologies for surface water-groundwater interactions using Distributed Temperature Sensing (DTS) in so far undiscovered environments,

3. to assess the performance of DTS as a tool for detecting groundwater discharge at various spatial and temporal scales,

4. to assess and combine methods of different spatial resolution to detect and quantify spatial and temporal variation in groundwater discharge from the small to the catchment scale,

5. to detect variability in surface water-groundwater exchange at the catchment scale.
1. Introduction

Paper I deals with objective 1, and aims to assess small-scale heterogeneity both in horizontal and vertical streambed hydraulic conductivity to investigate if the scale of spatial variability in groundwater discharge (as presented in Paper III) is related to a similar heterogeneity in streambed hydraulic conductivity.

Paper II elaborates on objectives 2 and 3 by applying DTS in a lake environment during a winter, spring and a summer month with the fiber optic cable deployed in looped pattern at the lake bottom to enhance the resolution of the measurements. DTS data were also compared with seepage meter and temperature profile-based vertical groundwater flux estimates (objective 4).

Paper III presents a new methodology to apply DTS in soft-bedded streams where sedimentation processes are likely to change the position of the fiber optic cable relative to the streambed (objectives 2, 3). This study also shows that DTS can be used to monitor sedimentation processes in soft-bedded streams.

Paper IV addresses objectives 3, 4 by comparing DTS measurements with Acoustic Doppler Current Profiler (ADCP) and temperature-based vertical flux measurements along a 2.5 km long stream stretch. The study also involves the analysis of sub-catchment scale responses to rain events based on hydrograph separation using $\delta^{2}H$ and electrical conductivity as tracers (objective 5).

Paper V presents a new, simple way to detect groundwater discharge on the catchment scale by linear regression analysis of air temperature and stream temperatures measured at two depths in the water column (objective 5).
Chapter 2

Methods

The interaction between groundwater and the surface water bodies can be studied at several spatial and temporal scales by different methods. This chapter gives a short introduction to the methods applied in the PhD research and discusses their capabilities and limitations.

2.1 Temperature as a tracer

Temperature as a tracer to detect groundwater discharge has been successfully applied for 50 years (Anderson, 2005). The method relies on the natural temperature contrast between surface water and groundwater making use of the relatively stable groundwater temperature as opposed to the larger annual and diurnal temperature oscillations of surface waters (Silliman and Booth, 1993). As an advantage of the method, temperature is a naturally occurring tracer which is very robust and easy to monitor compared to other chemical tracers (Constantz, 2008) and the field instrumentation is relatively non-invasive.

2.1.1 Vertical temperature profiles

Assuming that the natural temperature oscillations of the surface water body are attenuated beneath the SWI or dampened by upward groundwater flow, temperature measurements at several depths below the SWI can be used to indirectly estimate vertical groundwater fluxes. One approach is to fit the analytical solution of the steady state heat transport equation (Bredehoeft and Papadopulos, 1965; Stallman, 1965) to the observed data (Schmidt et al., 2007; Jensen and Engesgaard, 2011) giving point-scale estimates of fluxes both in space and time. Although the data acquisition for this method is relatively simple, Anibas et al. (2009) found that the steady-state solution can only reliably predict fluxes in the winter and at the end of the summer.
Temporal variability in fluxes can also be determined by methods using longer temperature time series assuming a sinusoidal temperature change at the SWI (Silliman et al., 1995). More recent methods are based on the phase and amplitude shift of the diurnal oscillation compared to the surface water temperature signal due to dampening below the SWI (Hatch et al., 2006; Keery et al., 2007; Jensen and Engesgaard, 2011; McCallum et al., 2012). While temperature time series can also quantify temporal variability in discharge, as a limitation of both approaches, they only give point-scale estimates of fluxes. Moreover, they assume vertical flow, thus are not capable of estimating lateral flow components and flux estimates are also influenced by streambed heterogeneities (Ferguson and Bense, 2011). Although most of the studies using sediment temperature profiles were carried out in streams, temperature has also been used in lakes to compare (Kishel and Gerla, 2002) and to estimate (Kidmose et al., 2013) groundwater fluxes.

2.1.2 Distributed Temperature Sensing

As opposed to the vertical temperature profiles yielding point-scale information, Distributed Temperature Sensing (DTS) can monitor SWI temperatures over large areas. The sensor of the method is a fiber optic cable, along which average temperature measurements are collected with a precision down to 0.01 °C (depending on the installation and calibration) (Selker et al., 2006b; Tyler et al., 2009) and sampling interval of 2-0.25 m (depending on instrument specifications). During measurements a laser pulse is sent through the fiber optic cable and the instrument measures the intensity of the reflected light. Based on the return time, the distance where the signal comes from can be determined. Most of the light is reflected at the same wavelength as emitted, but small portions are either reflected at a longer (Stokes backscatter) or shorter (Anti-Stokes backscatter) wavelengths than the original pulse (Raman scattering). The amplitude of the Anti-Stokes backscatter is linearly dependent on temperature, thus at shorter wavelengths the intensity of the return signal is dependent on the intensity of illumination and also the temperature (Selker et al., 2006a), this way the Stokes/Anti-Stokes ratio is temperature-dependent.

DTS has been widely used in streams (Westhoff et al., 2007; Lowry et al., 2007; Briggs et al., 2011; Krause et al., 2012), rivers (Mwakanyamale et al., 2012), salt-marsh channels (Moffett et al., 2008) and polders (Hoes et al., 2010), where laying the fiber optic cable on the SWI, longitudinal sections up to km scale were mapped for groundwater discharge. In some studies groundwater discharge was manifested as a step change of a few degrees in streambed temperatures (Selker et al., 2006b; Westhoff et al., 2007; Briggs et al., 2011), thus knowing the stream discharge upstream and
2. Methods

downstream of the detected discharge location, the quantity of groundwater discharge was calculated with a simple mixing equation (Selker et al., 2006b; Briggs et al., 2011). In other cases (Krause et al., 2012; Karthikeyan et al., 2012) the mixing analysis could not be applied due to the lack of large changes in SWI temperatures. As a disadvantage of the method, in such cases the DTS measurements alone cannot be used to quantify groundwater discharge.

Similarly to the vertical temperature profiles, the fiber optic cable can also be wrapped helically around a PVC pipe, giving temperature measurements with a vertical resolution of a few millimetres, thus the fluxes and their temporal variability between different depths below the SWI can be precisely determined (Vogt et al., 2010; Briggs et al., 2012). Nevertheless these installations can only yield point-scale flux estimates.

2.2 Seepage meter

Seepage meters are widely used to measure groundwater discharge directly both in lakes (McBride and Pfannkuch, 1975; Lee, 1977; Rosenberry et al., 2010) and in streams (Landon et al., 2001; Rosenberry, 2008). Depending on the area covered by the half-barrel (Lee-type) seepage meter, the method gives an average net groundwater flow through the SWI over a larger area than vertical temperature profiles, also including flow components other than vertical. Subject to the magnitude of the fluxes and size of the seepage bags, groundwater fluxes are integrated over a period of time, thus seepage meter measurements usually lack the fine scale temporal resolution on the scale of minutes to days, although coupled with flowmeters they can achieve fine scale temporal resolution (Paulsen et al., 2001; Rosenberry et al., 2013). Combined with piezometers screened under the SWI, seepage meters can also be used to determine hydraulic conductivities (Rosenberry and Pitlick, 2009; Wojnar et al., 2013).

Fluxes measured by seepage meters are also subject to external impacts such as velocity head effects in lotic environments (Rosenberry, 2008) or the effect of waves (Rosenberry et al., 2013). Results also have to be corrected for friction loss in the seepage meters (Rosenberry, 2008). Due to the difficulties arising from streamflow, bedform heterogeneities and sediment transport, the use of seepage meters in streams is less frequent.

2.3 Stable isotopes and environmental tracers

$^2$H and $^1$H are natural stable isotopes of hydrogen, also occurring in waters where their concentration is only altered during fractionation processes such as evaporation and
condensation. $^1$H is the lighter isotope thus more prone to go into vapour phase during evaporation, similarly the heavier $^2$H isotope is more likely to remain in the liquid phase during fractionation. Therefore, during the fractionation processes the $^2$H to $^1$H ratio changes. This ratio, the enrichment or depletion in $^2$H, can be expressed by the delta values, where the $^2$H to $^1$H ratio of the water sample is compared to the $^2$H to $^1$H ratio of the standard (VSMOW: Vienna Standard Mean Ocean Water):

$$
\delta^{2}H = \left( \frac{^{2}H/^{1}H_{sample}}{^{2}H/^{1}H_{standard}} - 1 \right) \times 1000
$$

$^2$H can be considered as a conservative tracer, thus its concentration only changes with mixing of waters of different origin. As precipitation, groundwater and surface water all go through different fractionation processes, their $\delta^{2}H$ values are different. Precipitation shows high variability in $\delta^{2}H$, while groundwater $\delta^{2}H$ values are relatively stable. Stream samples usually consist of a mixture of these two components, while due to evaporation lake water is enriched in $^2$H. Based on these differences $\delta^{2}H$ (and also $\delta^{18}O$) has been generally used in stream studies as a basis of hydrograph separation (Sklash and Farvolden, 1979; Hoeg et al., 2000) or in lake studies to estimate groundwater contribution in the lake water budget (Krabbenhoft et al., 1990).

### 2.4 Piezometers, permeameters

Piezometers are commonly used to monitor hydraulic heads and thereby provide a general interpretation of flow conditions over several spatial scales. Piezometers can also be installed directly in streams and lakes to observe vertical head gradients and consequently to determine the direction and magnitude of vertical flow (Käser et al., 2009; Krause et al., 2012) or combined with slugtests also to quantify groundwater fluxes (Conant, 2004). The natural heterogeneity of horizontal hydraulic conductivity determined by slugtests in these piezometers (Cey et al., 1998; Landon et al., 2001; Leek et al., 2009) can also explain some of the spatial variability in groundwater discharge to surface water bodies.

Meanwhile the location of groundwater upwelling zones is rather determined by the vertical hydraulic conductivity of streambed/lakebed sediments. Based on the Hvorslev method (Hvorslev, 1951) or its simplification (Chen, 2000), in-situ permeameters are widely used to determine vertical hydraulic conductivity (Genereux et al., 2008; Cheng et al., 2011).
2.5 Discharge measurements

Groundwater discharge contributes to the net increase of the streamflow, thus discharge measurements along the stream can also reflect the magnitudes and spatial distribution of the groundwater contribution. Recently the Acoustic Doppler Current Profilers (ADCP) provide for high precision discharge measurements with short measurement times (Mueller and Wagner, 2009), although the spatial resolution of the stream discharge data is limited by the spacing of the measurement locations to obtain net increases in discharge above the uncertainty of measurements.
Chapter 3

PhD research

3.1 Field sites

Surface water-groundwater interactions were studied over several spatial and temporal scales in a stream and in a lake at two catchments in Western Denmark (Figure 3.1a). On the small scale, studies were carried out in Holtum stream (Papers I, III), located in the upper part of the Skjern River catchment (Figure 3.1b) and at Lake Væng (Paper II) situated in the catchment of the Gudenå River (Figure 3.1c). To map medium scale spatial variability in groundwater discharge, a 2.5 km long section of Holtum stream was investigated (Paper IV) and catchment scale differences and responses were also studied in the catchment and sub-catchments of Holtum stream (Papers IV,V).

Figure 3.1: The location of field sites in Denmark (A) and an overview map of the Holtum catchment (B) and the catchment of Lake Væng (C).
3.2 Main findings

3.2.1 Natural heterogeneity of streambeds

Spatial variability in groundwater discharge is largely controlled by natural heterogeneity in streambed hydraulic conductivity (Kalbus et al., 2009). Paper I therefore aims to quantify this natural heterogeneity in streambed hydraulic conductivities both in the horizontal ($K_h$) and vertical ($K_v$) direction in a small-scale, 5x6 m area in Holtum stream. 40 piezometers were installed with a screen 0.5 m below the streambed and at the same locations, permeameter tests on 0.5 m long sediment columns were carried out. The test locations were grouped into 8 transects to account for variability in $K_h$ and $K_v$ across the stream. To detect seasonal variability, the tests were conducted at the same locations in December 2011 and August 2012. Vertical head gradients in the piezometers were also measured to obtain estimates of the direction and magnitude of vertical groundwater flux.

The results suggest very large spatial variability both in $K_h$ and $K_v$ with a range of $K_h$ values of 0.15-54 $\text{md}^{-1}$ and $K_v$ values of $7 \times 10^{-4}$ - 6 $\text{md}^{-1}$. From December 2011 to August 2012 $K_v$ values not only show the expected general increase in absolute values due to the reduced water viscosity of the summer month, but also a much higher temporal variability than $K_h$. This can be explained by the mobile streambed sediments which, due to their redistribution, change the material properties of the tested vertical material column between the surveys, while sediments 0.5 m below the streambed stay relatively stable as indicated by the low temporal variability in $K_h$ values.

This study also showed that $K_v$ values are determined by a fine organic layer located 0.12-0.3 m below the streambed. This layer could be the reason for observing generally lower $K_v$ and anisotropy values than previously reported for streams with sandy streambeds (Chen, 2000; Cheng et al., 2011; Landon et al., 2001). At test locations where this organic layer was not present or was discontinuous, higher $K_v$ values were measured. This shallow organic layer can also indicate a shallow hyporheic zone. Despite the low $K_v$ values, vertical head gradients indicated groundwater upwelling everywhere in the study section in December 2011 and only three out of 40 test locations with downwelling in August 2012, further emphasizing the importance of spatial heterogeneities on discharging groundwater in the catchment.

3.2.2 Spatial variability in groundwater discharge

Papers II and III studied surface water-groundwater interactions on the small scale. DTS measurements in Paper II were carried out in Lake Væng, in a 25x6 m area, while in Paper III along Holtum stream in a 70x5 m stretch (also including the study area
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of Paper I). In both cases as opposed to the single longitudinal deployment (Lowry et al., 2007; Briggs et al., 2011) or a layout with two cable rows (Krause et al., 2012), the fiber optic cable was looped at the SWI to allow temperature measurements of a grid of 1x0.5 m. Both studies show SWI temperature differences even over 0.5 m distance indicating a similar variability in groundwater discharge, reflecting the large natural heterogeneity of streamed materials as shown by the streambed hydraulic conductivity measurements of Paper I.

Paper II presents the results of the first DTS study carried out in a lake environment with the purpose of detecting groundwater discharge into the lake. DTS data indicated scattered high-discharge sites in Lake Væng as it was previously observed by Kishel and Gerla (2002). Due to the special geology of the area, the discharge sites were detected farther from the shore, not following the traditional exponentially decreasing discharge pattern from the shoreline (McBride and Pfannkuch, 1975). Groundwater fluxes measured by seepage meters and estimated based on lakebed temperature profiles show a variability by a factor of 200 (between 2.7x10\(^{-3}\) and 0.62 md\(^{-1}\)) in this small area.

The stream DTS survey of Paper III shows that SWI temperatures are far from homogeneous across the stream, thus in wide, high-discharge streams the single longitudinal DTS layout frequently applied in small, low-discharge streams (Westhoff et al., 2007; Lowry et al., 2007) may not be able to accurately detect the natural spatial variability. Similarly to the modelling study of Storey et al. (2003), potential concentrated groundwater discharge sites in the stream were mostly detected scattered along the streambanks with their location controlled by streambed morphology.

Spatial variability of groundwater discharge on the medium scale was investigated in Paper IV, where the fiber optic cable was deployed longitudinally at the SWI of Holtum stream over a 2.5 km long section. Due to the extent of the mapped area and the longitudinal layout, this study does not show the same large spatial heterogeneity in discharge as the small-scale measurements (Paper II and III), but similarly to other studies (Lowry et al., 2007; Briggs et al., 2011) several areas of concentrated groundwater discharge can be identified.

The observed step changes in SWI temperatures at the concentrated discharge sites were however low, on the scale of 0.5-1\(^{\circ}\)C, without a recovery curve as it was previously observed in other lowland streams (Krause et al., 2012; Karthikeyan et al., 2012). Moreover, as the extent of these zones was also small, quantification of fluxes by differential gauging and a mixing analysis (Selker et al., 2006b; Westhoff et al., 2007) was not possible. Meanwhile vertical groundwater fluxes estimated from streambed temperature profiles at the discharge areas indicated by the DTS also show large variability (0.06 and 0.86 md\(^{-1}\)). Based on the vertical head gradients presented in Paper I and
the results of the $\delta^2$H-based hydrograph separation of Paper IV, it can be assumed that Holtum stream is a groundwater-dominated stream, thus discharge sites indicated in this study by the DTS are high-discharge sites as opposed to low groundwater discharge through the rest of the streambed.

Papers IV and V address catchment scale variability in discharge. Paper IV used $\delta^2$H and electrical conductivity-based hydrograph separation to assess sub-catchment responses to rain events. Out of the four sub-catchments studied, the most upstream one with the largest proportion of agricultural areas and tile drains reacted the fastest and in the highest degree to the rain events, with an event water percentage of 30-80%. The sub-catchment of a tributary of Holtum stream is completely groundwater-dominated, showing only a very low percentage of event water, 30% during the largest and most intensive rain event. The sub catchment also described in Papers I and III reacts to rain events with an event water percentage of 10-60%, but its response is more delayed and dampened compared to the most upstream catchment.

Paper V presents a simple method to detect groundwater discharge based on the linear regression analysis of air and stream water temperatures, measured at two depths. Linear regression analysis was carried out at 8 stations along a 15 km stream section (comprising all the sub-catchments of Paper IV). At stations with high groundwater discharge there is a considerable difference in slope and intercept between the regression line of the air temperature and the stream water temperature measured close to the surface and air temperature and stream water temperature measured at the bottom of the water column. Due to the groundwater influence the intercept of the regression line increases, while the slope decreases as was also observed by O’Driscoll and De-Walle (2006). This qualitative indication of groundwater discharge was also confirmed by discharge and flux measurements. The highest measured flow accretions (0.08-0.24 m$^3$s$^{-1}$km$^{-1}$) and estimated vertical groundwater fluxes (0.6-1.2 md$^{-1}$) were observed at stations where the stream water temperature at the bottom of the water column showed only a slight correlation with air temperature and a mean diel amplitude below 0.6 °C due to discharging groundwater. Low diel amplitudes as indicated by reduced diurnal oscillations in temperature were also observed in another Danish stream (Pedersen and Sand-Jensen, 2007) and were attributed to the influence of groundwater. Next to the high discharge sites, other stations were qualified as low-discharge sites as corresponding to the findings of Paper IV, vertical groundwater fluxes in the order of 0.1 md$^{-1}$ could still be observed. The location of high-discharge sites in the catchment does not follow any distinct pattern, they are scattered along the stream, showing spatial variability on the catchment scale.
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3.2.3 Temporal variability in groundwater discharge

Based on in-stream hydraulic head measurements, Paper I demonstrates that a small-scale stream section changed from entirely gaining conditions to a partially loosing condition from December 2011 to August 2012. Temporal variability in both vertical and horizontal streambed hydraulic conductivity was also observed concluding that $K_v$ shows larger temporal variability than $K_h$. Similarly to Genereux et al. (2008) this temporal variability in $K_v$ is attributed to the highly mobile streambed (shown by Paper III) where the upper sediment layers are constantly changing resulting in variability in $K_v$ as opposed to $K_h$ which was measured 0.5 m below the streambed in a quasi-stable environment. These temporal changes in $K_v$ could indicate similar temporal changes in groundwater discharge. Confirming the findings of this study, DTS data in Paper III showed changes in the location of potential discharge sites between October, November and December 2011. In a similar stream Conant (2004) found a lower variability in discharge between seasons. In the study presented in Paper III however, the larger temporal variability in the distribution of discharge is most likely due to changes in streambed morphology induced a sedimentation research experiment, which by the dislocation of bedforms can also result in the dislocation of upwelling zones.

Paper II observed temporal variability in upwelling within the winter, spring and summer seasons in a small area in Lake Væng. In this study the location of one high-discharge zone, the farthest away from the lakeshore, remained constant during the seasons, although its spatial extent and relative importance decreased from February to August 2012. In the meanwhile new discharge zones appeared closer to the lakeshore. Temporal variability on the annual scale was observed by the aid of ice holes, where the concentrated discharge of warm groundwater prevented ice formation on the lake surface. Ice hole observations from the winter of 2010 and 2012 also confirm the temporal variability in discharge detected by DTS. While in other studies temporal variability in discharge fluxes was correlated with rainfall data (Downing and Peterka, 1978), similarly to the findings of Shaw and Prepas (1990) these temporal changes can possibly be explained by changes in hydraulic head at the Western lakeshore where most of the discharging groundwater originates.

Sub-catchment scale temporal variability in groundwater discharge was addressed in Paper IV. Temporal variability in event fractions was observed at each investigated sub-catchment. This variability is most likely due to differences in intensity and antecedent wetness conditions between the rain events in April, May and September 2012. The largest temporal variability in responses can be observed at the most upstream sub-catchment which also has the fastest and largest response to rain events, on the other hand the groundwater-fed tributary shows the least variability in response, due to a
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constant, relatively high input of groundwater.

3.2.4 Assessment of DTS in new environments and comparison with other methods

DTS in lakes (Selker et al., 2006a; Suarez et al., 2011) has not been applied so far with the purpose of detecting groundwater discharge as done in numerous cases in streams (Lowry et al., 2007; Briggs et al., 2011). The deployment of DTS in lakes is different from stream applications due to the quasi-motionless water body. Paper II shows that SWI temperatures measured with the fiber optic cable are more exposed to external conditions as in case of stream applications. Depending on the water depth, solar radiation can influence DTS recordings by the heating of the fiber optic cable (Neilson et al., 2010), while multi-level lake water measurements in Lake Væng revealed a homogeneous temperature distribution in the water column during the night, indicating that night time DTS data is free from the effects of solar radiation. Similarly cold surface inlets can make the interpretation of DTS data more uncertain during summer, when the denser cold water sinking and propagating at the bottom of the lake can be misinterpreted as groundwater discharge.

Paper III shows the complications of using DTS in a groundwater-dominated, lowland, soft-bedded stream, where sedimentation processes constantly modify the position of the cable relative to the streambed. Consequently, DTS did not exclusively measure SWI temperatures, but in case of sediment deposition rather temperatures below the streambed. Temperatures at deposition sites resemble the groundwater signal (are colder than SWI temperatures during summer and warmer during the winter) and diurnal oscillations are also dampened. Combining this with detailed streambed elevation surveys, Paper III demonstrates the capability of DTS to monitor sedimentation processes in soft-bedded streams.

As also observed by Krause et al. (2012) in a lowland stream and Karthikeyan et al. (2012) in a Danish stream, the groundwater-induced temperature anomalies at the SWI were low, in our study only of maximum 1-2 °C, thus without the large step changes as observed in previous studies (Briggs et al., 2011; Lowry et al., 2007) the delineation of potential groundwater discharge sites from sedimentation sites is difficult. Paper III also proposes a method to overcome this problem by the aid of detailed streambed elevation surveys and the use of diel metrics of temperature data while also directly quantifying the effect of streambed sediments on the temperature recorded by DTS. The study shows that a 0.1 m thick sediment layer on the cable can already alter DTS recordings, while also showing that this range of streambed elevation change occurs naturally in soft-bedded streams.
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The performance of DTS was compared with traditional methods, seepage meter and temperature-based flux estimates on the small-scale in a lake (Paper II) and differential gauging and temperature-based fluxes on the medium-scale in a stream (Paper IV). So far several studies estimated groundwater fluxes by step changes in SWI temperatures and differential gauging (Selker et al., 2006b; Briggs et al., 2011), while only a few attempts have been made to compare DTS results to flux estimates of traditional point scale methods: seepage meter measurements (Lowry et al., 2007) or flux estimates based on head gradients (Krause et al., 2012). On the small-scale Paper II showed that DTS had difficulties detecting high-discharge sites of reduced size as indicated by the ice hole observations, it could only identify with confidence discharge sites of above 0.3 md$^{-1}$ discharge flux. DTS was neither capable of reflecting the real spatial heterogeneity in discharge as shown by seepage meter and temperature-based fluxes. These discrepancies are probably due to the spatial averaging interval of the instrument (Rose et al., 2013), where in these installations temperature recordings will be averaged over 1 m of fiber optic cable as opposed to the concentrated discharge sites of reduced spatial extent, thus temperature recordings will be more homogeneous than in reality.

On the medium scale however, similarly to the results of Briggs et al. (2011) DTS data agrees very well with differential gauging (100-200 m long sections) and temperature-based fluxes on the approximate location of discharge sites. This better agreement on the medium-scale can possibly be explained by the scale of the measurements as at a longitudinal deployment length of several hundred metres, the spatial averaging of one meter does not mask as much information as during small-scale measurements.
Chapter 4

Conclusions and perspectives

As surface water and groundwater are connected resources, pesticides and nutrients polluting one, will also influence the ecological state of the other. For this reason it is of great importance to understand the complex interaction of these systems and also to be able to account for the natural spatial and temporal variability of the exchange processes. A way of investigating these interactions is the detection and quantification of exchange fluxes at several spatial and temporal scales. So far there are several methods to quantify point fluxes (temperature and seepage meter-based estimates) both in lake and stream environments and large areas can also be mapped using SWI temperatures as a tracer for groundwater discharge, but the combination of the two approaches is rarely encountered.

The objective of the present PhD research was to detect variability in groundwater discharge at several spatial and temporal scales. The main method applied in the study was Distributed Temperature Sensing (DTS) which was used at various scales and environments. On the small-scale, DTS data from Holtum stream indicated large differences in groundwater discharge across and along the stream. DTS was combined with vertical temperature profiles and seepage meters in Lake Væng, showing that groundwater discharge can vary by a factor of 200 in an area of 25x6 m. On the kilometer scale DTS was able to detect several concentrated discharge sites along Holtum stream which were also confirmed by differential gauging and vertical temperature profiles. Sub-catchment scale variability was also observed with hydrograph separation based on δ²H and electrical conductivity.

Temporal variability in discharge was described in the small-scale studies. In Lake Væng the discharge locations and also their spatial extent changed within months probably due to changes in hydraulic heads, while in Holtum stream the temporal changes in discharge were also related to sedimentation processes and changes in streambed morphology.

The PhD research also addressed the topic of using DTS in surface waters with
low temperature contrast between the surface water and groundwater. As within these conditions no large changes in SWI temperatures can be observed, the quantification of groundwater discharge is not possible by the traditionally applied mixing analysis. For this reason most DTS measurements were combined with independent point-scale flux measurements. These measurements confirmed the ability of the DTS to detect major discharge sites, while it was not capable of detecting the real spatial heterogeneity in groundwater discharge and concentrated discharge sites of reduced spatial extent. This is probably a result of the instrumentation, where measured Sediment-Water Interface (SWI) temperatures are averaged over one meter distance along the fiber optic cable.

DTS was also deployed in a soft-bedded stream, where it monitored sedimentation processes based on their induced temperature anomalies. Next to this, a new monitoring methodology was applied to differentiate sedimentation-induced temperature anomalies from the temperature signal of groundwater discharge. DTS was also used in a lake environment where due to the quasi-motionless water, external influences also had to be taken into account when analysing SWI temperature data. Both cold inlets and solar radiation were shown to considerably influence SWI temperatures and consequently data interpretation.

All the studies conducted in the PhD research, starting from small-scale measurements to catchment scale processes, showed that groundwater is an essential element of the surface water budget. This presents the need to monitor surface water-groundwater exchange processes in lowland catchments. As in several cases DTS was able to detect concentrated discharge sites and temporal variability in groundwater discharge over large areas, installations can possibly be used in the future for long-term monitoring purposes as well, both in lake and stream environments.

This PhD research presented the use of DTS in a lake and soft-bedded stream environment. There are however, many possible applications, environments and external influences yet unexplored. Another great advancement of the DTS method would be the development of a general monitoring methodology, including the upscaling of exchange fluxes with the combined application of DTS and traditional flux estimation methods.
References


Chapter 5

Paper I

Spatial variability in streambed hydraulic conductivity of contrasting stream morphologies: channel bend and straight channel

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\textit{manuscript ready for submission}
Abstract

Streambed horizontal hydraulic conductivities ($K_h$) were determined from in-stream slugtests, vertical hydraulic conductivities ($K_v$) were calculated with in-stream permeameter tests and hydraulic heads were measured to obtain vertical head gradients (VHG) at 8 transects, each comprising 5 test locations, in a groundwater-dominated stream. Seasonal small-scale measurements were taken in December 2011 and August 2012, both in a straight stream channel with homogeneous elevation and downstream of a channel meander with heterogeneous elevation. All streambed attributes showed large spatial variability. $K_h$ values were the highest at the depositional inner bend of the stream, while high $K_v$ values were observed at the erosional outer bend and near the middle of the channel. Calculated $K_v$ values are related to the thickness of the organic streambed sediment layer and also showed higher temporal variability than $K_h$ due to sedimentation and scouring processes affecting the upper layers of the streambed. Test locations at the channel bend showed a more heterogeneous distribution of streambed properties than test locations in the straight channel, while within the channel bend higher spatial variability in streambed attributes was observed across the stream than along the stream channel.

5.1 Introduction

Understanding surface water-groundwater exchange is a key to understand processes influencing stream ecology such as the discharge and mixing of pollutants (Böhlke and Denver, 1995; Flewelling et al., 2012) or biogeochemical processes (Brunke and Gonser, 1997) in the hyporheic zone. Heterogeneity in streambed hydraulic conductivity ($K$) is for example a main factor controlling spatial variability in surface water-groundwater interactions (Kalbus et al., 2009) not only on the streambed scale, but also in a generalized form of leakage coefficients in case of large, catchment-scale models. Due to the natural complexity of the stream-aquifer systems, it is therefore difficult to characterise and quantify the spatial and temporal variability of these processes.

The determination of streambed $K$ is difficult despite the great variety of methods that can be applied. These methods include slugtests (Cey et al., 1998; Landon et al., 2001; Leek et al., 2009), in-situ permeameter measurements (Chen, 2000;
different methods used to estimate K can also yield different results (Landon et al., 2001; Cheong et al., 2008) partly due to the measurement scale and the directionality of measurements. Slug tests give average information about the horizontal hydraulic conductivity ($K_h$) of a streambed sediment layer at a specific depth were the screen is located. Permeameters can be used to measure hydraulic conductivity both in the horizontal and vertical direction (Chen, 2000, 2004), with the vertical tests giving information on the order of tens of centimeters of a streambed column possibly consisting of various sediment layers. The use of grain size analysis yields estimates without directions (bulk K) of a measurement volume depending on the size of the disturbed sample. Due to the scale of measurements $K_h$ values obtained from slug tests and on-shore pumping tests (Hunt et al., 2001; Nyholm et al., 2002) are generally higher than K values estimated from grain size analysis ($K_g$) and vertical hydraulic conductivities ($K_v$) are usually the lowest (Cheng and Chen, 2007; Song et al., 2009).

Another difficulty of characterising surface water-groundwater exchanges is the natural heterogeneity of both $K_h$ and $K_v$ of streambed materials. Leek et al. (2009) found variations of three orders of magnitude in $K_h$ and statistically significant differences within different streambed depths from 0.3 m to 1.2 m below the streambed level. Cheng et al. (2011) found a correlation scale of less than 1.5 m in $K_v$ at four sites which corresponded to the distance between test points, while Chen (2011) observed a decreasing trend in $K_v$ with depth. Although observing a high degree of spatial variability in $K_v$, Vertical Head Gradient (VHG), and nitrate concentration in the horizontal direction, Kennedy et al. (2008) found that $K_v$ is the least spatially variable streambed attribute.

Changes in water viscosity, sedimentation and scouring processes (Genereux et al., 2008; Levy et al., 2011), or biogeochemical processes such as clogging can result in temporal changes in K (Blaschke et al., 2003). With the exception of Genereux et al. (2008) and Levy et al. (2011), who linked temporal variability in $K_v$ to erosion and deposition of streambed sediments, studies about streambed hydraulic conductivity generally lack the temporal aspect.

As for the spatial aspect, Dong et al. (2012) found large differences in $K_v$ values between the point bar and the meandering channel and several studies observed great variability in K across river channels of a width of 50-400 m (Chen, 2004, 2005; Chen et al., 2009). Genereux et al. (2008) however, found uniform K values across a 10 m
wide stream channel. Chen (2005) and Genereux et al. (2008) related cross-channel changes in \( K_v \) to changes in water depth and indirectly to differences in flow velocity. With the exception of Genereux et al. (2008), the measurements were carried out at test locations with several metres of spacing. Despite a multitude of studies investigating streambed hydraulic conductivity patterns, so far there is no comprehensive small-scale survey relating spatial and temporal variability in \( K_h, K_v \), VHG, and therefore also the ratio of \( K_h \) to \( K_v \) (from now on referred to a streambed anisotropy) to streambed morphology.

In this study we have therefore determined \( K_v, K_h \), streambed anisotropy and vertical head gradients (VHG) in two areas of a relatively small stream section; downstream of a channel meander and in a straight channel stretch during winter and summer. The objectives have been to: (i) measure small-scale spatial and seasonal variability in streambed hydraulic conductivity and its anisotropy, (ii) relate this variability to channel morphology, and (iii) compare the measured and calculated properties of different streambed sediments. Spatial and temporal variability in streambed properties were studied by performing slug tests, in-situ permeameter tests, and hydraulic head measurements at 40 locations in the stream in December 2011 and August 2012. Streambed sediments were described by removing sediment cores at the 40 test locations in August 2012. Spatial patterns and correlation between streambed attributes were assessed using Principal Component Analysis.

### 5.2 Field site

The study was conducted in the perennial, gaining, lowland Holtum stream located in the Skjern river catchment in Jutland, Western Denmark (Figure 5.1a). The stream has a catchment area of 70.4 km\(^2\) with predominantly agricultural (56%) and forested (23%) land use. The upper sediments of the shallow aquifer in the catchment area are dominated by glacial sand and silt from the Weichsel period (Houmark-Nielsen, 1989). The mean annual stream discharge was 1.3 m\(^3\)s\(^{-1}\) in 2011 and 1.4 m\(^3\)s\(^{-1}\) in 2012 measured at the gauging station 2 km downstream from the study site (Figure 5.1b). Stream water temperature ranged between 1 and 16 \(^\circ\)C during the year. The groundwater-fed stream has a width of 3.5-5 m and a depth of 0.5-0.7 m at the 40 m long stream section. The stream flows from East to West with a sharp bend in the channel (Figure 5.1c). Due to this bend, the right streambank is being eroded while sedimentation processes occur at the left bank.

Field measurements took place between 12-14 December 2011 and 1-4 August 2012 when the daily mean discharge was 1.7 and 1.0 m\(^3\)s\(^{-1}\), respectively. To relate spatial
and seasonal variability in streambed attributes ($K_h$, $K_v$, VHG) to differences in stream morphology, two stream sections 35 m apart were selected for investigation: one in the straight stream channel and one downstream a meander bend (Figure 5.1c). The site downstream the meander bend (Figure 5.1d) (from now on referred to as meander section) contained six transects, each with 5 test locations forming a measurement grid of approx. 1 m spacing along and 0.5 m spacing across the stream. This meander section of 6 x 5 m showed abruptly changing streambed elevation with a maximum difference in streambed elevation of 0.52 m and 0.47 m in December and August, respectively, between the depositional left stream bank and an erosional right stream bank (Figure 5.1d). The straight channel section of 1 x 5 m (Figure 5.1e) (from now on referred to as straight section) comprised two transects, had a homogeneous streambed elevation, with maximum elevation differences of 0.12 m between the test locations both in December and August (Figure 5.1e). During both seasons test locations were named after the month and their serial number in the downstream direction, e.g. D27 is test location 27 in December (Figure 5.1d).

5.3 Methods

By having test locations 1-10 in a straight channel and test locations 11-40 downstream of a channel bend (Figure 5.1c), it is possible to compare streambed attributes of these channel types. The five test locations across the channel also make it possible to compare spatial heterogeneity and seasonal variations in streambed attributes in the depositional inner bend and erosional outer bend of the stream.
5.3.1 Horizontal hydraulic conductivity

In December 2011 and August 2012, 40 plastic piezometer pipes of 2.5 cm diameter and 0.1 m long screens were installed by direct push method 0.5 m below the streambed. The clean-pumped piezometers were left for at least 15 hours to stabilize, then one to seven slugtests were carried out in each piezometer. Even the time span of the slug tests indicated high spatial heterogeneity in $K_h$ as in the slowest piezometers only one slug test was carried out and lasted for an hour, while in the fastest ones seven slugtests could be finished in 5 minutes.

![Measurement setup at the test locations showing the position of piezometers (A) where slugtests and VHG measurements were carried out and the transparent pipes for the falling head tests (B).](image)

The slug test data was analysed by fitting the confined Hvorslev solution (Hvorslev, 1951) to the data using the AQTESOLV 3.5 software, assuming that the piezometer partially penetrates the aquifer. At each test location, the arithmetic mean of the $K_h$ values was calculated for December and August, respectively.

5.3.2 Vertical hydraulic conductivity

In-situ vertical hydraulic conductivities were measured by the permeameter method as described by Hvorslev (1951). Transparent PVC pipes of 5 cm diameter and 1.75 mm wall thickness were installed next to the piezometer pipes. Each pipe was installed to 0.5 m depth below the streambed, thus trapping a streambed sediment column of 0.5
m length (Figure 5.2b). After reaching a stable water level in the pipes, they were completely filled up with stream water thus creating a gradient in hydraulic heads. The recovery of the initial water level was recorded by measuring the water level in the pipes first every 10 minutes, later on an hourly basis. \(K_v\) values were calculated on the basis of the solution provided by Hvorslev (1951):

\[
K_v = \frac{\pi D}{11m + L_v} \ln\left(\frac{h_1}{h_2}\right)
\]

(5.1)

where \(D\) is the diameter of the pipe (m), \(L_v\) is the length of the streambed sediment column in the pipe (m), and \(h_1\) and \(h_2\) are hydraulic heads observed in the pipe (m) corresponding to measurement times of \(t_1\) and \(t_2\) (min). The term \(m\) expresses the ratio of horizontal \((K_h)\) and vertical hydraulic conductivity \((K_v)\):

\[
m = \sqrt{\frac{K_h}{K_v}}
\]

(5.2)

Chen (2000) found that if the length of the sediment column \((L_v)\) is several times larger than the diameter of the pipe \((D)\) than the formula provided by Hvorslev (1951) (Eq. 5.1) can be simplified to:

\[
K_v = \frac{L_v}{t_2 - t_1} \ln\left(\frac{h_1}{h_2}\right)
\]

(5.3)

Chen (2000) observed that this solution underestimates the \(K_v\) of streambed sediments with small anisotropy.

At each test location \(K_v\) will be calculated by five solutions: i) assuming isotropic conditions \((m=1)\), ii) with an anisotropy ratio of 9 \((m=3)\), iii) assuming strongly anisotropic conditions \((m=10)\), iv) by substituting \(K_h\) values observed at the test locations in \(m\), and v) with the solution given by Chen (2000) (Eq. 5.3). \(K_v\) for each method is given as the arithmetic mean of \(K_v\) values calculated by all possible \(t_1, t_2, ..., t_n\), time step combinations. Anisotropy in \(K\) was calculated by the ratio of \(K_h\) and \(K_v\) values at the test location.

### 5.3.3 Hydraulic conductivity from grain size analysis

Following the December 2011 campaign four sediment cores, and after the August 2012 field campaign all 40 streambed sediment cores of the permeameter experiments were removed with the permeameters for a visual qualitative description of streambed
sediments. Sediment cores with the most streambed material and the best preserved streambed structure were selected for further analysis. Dry sieve grain size analysis was carried out on 21 streambed sediment samples taken from cores A2, A6, D12, A27, D38, and A39. The largest and smallest sieve aperture was 8 mm and 0.063 mm, respectively with an additional 11 sieves in between. Fractions finer than 63 microns were not analysed as they represented less than 0.43% of the sample weight. Estimated hydraulic conductivity from grain size distribution ($K_g$) will be compared to $K_h$ estimates from August 2012 when most of the streambed samples were taken.

Vukovic and Soro (1992) showed that the application of different empirical formulas to calculate $K_g$ can result in differences of a factor of 100 using the same dataset. For this reason six empirical solutions, all having the domain of application to medium and coarse-grained sand, were used to calculate $K_g$. The Hazen, USBR, Kozeny, Schlichter and Terzaghi methods were used as described by the general equation given by Vukovic and Soro (1992):

$$K_g = \frac{g}{\nu} \times C \times \varphi(n) \times d_e^2$$  \hspace{1cm} (5.4)

where $K_g$ is the hydraulic conductivity, $g$ is the acceleration of gravity, $\nu$ is the kinematic coefficient of viscosity, $C$ is a dimensionless coefficient depending on the properties of the porous medium, $n$ is the porosity, $\varphi(n)$ is the porosity function, and $d_e$ is the effective grain size diameter. The used parameters of the general equation for each empirical formula are given in Table 5.1. A kinematic coefficient of viscosity of $1.31\times10^{-6}$ m$^2$s$^{-1}$ was used conforming to the stream temperature of 10-12 °C measured in August 2012. The porosity was estimated according to (Vukovic and Soro, 1992):

$$n = 0.255(1 + 0.83^\eta)$$  \hspace{1cm} (5.5)

where $\eta$ is the coefficient of uniformity of the material given by:

$$\eta = \frac{d_{60}}{d_{10}}$$  \hspace{1cm} (5.6)

d$_{10}$ and d$_{60}$ being the particle diameters corresponding to cumulative fractions of 10% and 60%, respectively. Based on previous studies, Shepherd (1989) derived the following formula to estimate hydraulic conductivity of channel sediments:

$$K_g = C \times d_e^{1.65}$$  \hspace{1cm} (5.7)
where \( d_e \) is the particle diameter corresponding to 50% cumulative fraction and \( C \) is a dimensionless coefficient.

### 5.3.4 Vertical head gradient

Hydraulic heads were measured in the piezometers 15 and 16 hours after the installation in December 2011 and August 2012, respectively. Vertical head gradients (VHG) at the test locations were calculated according to

\[
VHG = \frac{h_i - h_j}{l}
\]

where \( h_i \) is the depth of the stream water level from the top of the piezometer, \( h_j \) is the water level in the piezometer and \( l \) is the depth of the piezometer screen below the streambed (Figure 5.2a), thus negative values reflect inflow to the stream.

### 5.3.5 Relating streambed attributes to stream morphology

A Principal Component Analysis (PCA) was carried out to detect the patterns emerging from the datasets of \( K_h \), \( K_v \), VHG, the elevation of the piezometer screen, and the calculated streambed anisotropy of the test locations from December 2011 and August 2012 and to relate this pattern to streambed morphology. The meander and the straight section will be compared based on the scatter of their respective test locations on the biplot.

The non-parametric Kruskal-Wallis test is used to assess similarities in distribution between different populations of data without assuming a normal distribution. This test was used to assess the similarities between the different geomorphological environments and the test locations across the stream. To compare the differences across the stream channel, \( K_h \), \( K_v \), VHG, and anisotropy values are visualized in box plots.

### 5.4 Results

#### 5.4.1 Horizontal hydraulic conductivity

For the seasonal comparison, \( K_h \) values were recalculated to a common reference temperature of 20 °C. The lowest and highest \( K_h \) values are observed in the middle and near the depositional inner bank of the meander section, respectively in both December and August (Figure 5.3a, b). The \( K_h \) values vary between 0.23 and 53.1 md\(^{-1}\) and
Figure 5.3: Interpolated contour maps of $K_h$ (A,B), $K_v$ (C,D), VHG (E,F), and anisotropy in K (G,H) in December 2011 and August 2012 for the meander section (the location is shown on Figure 5.1d). $K_h$ and $K_v$ values are given in $\text{md}^{-1}$ and are recalculated for a common reference temperature of 20 $^\circ\text{C}$. On each map the thickness of the organic sediment layer, as surveyed in August 2012, is shown.
between 0.15 and 53.7 md$^{-1}$ in December and August, respectively, showing a lognormal distribution. The histograms of ln($K_h$) for the meander and straight sections are shown for both seasons in Figure 5.4a and b. Both in December and August, $K_h$ values are similar in the middle section and the erosional right bank of the stream but not at the depositional left bank (Figure 5.5a) where higher values are detected in December. In August the highest values are observed closer to the middle of the channel (Figure 5.5a). The Kruskal-Wallis test confirmed that $K_h$ values at the depositional left bank of the stream and in the middle of the channel are statistically different both in December and August. The test did not show any statistically significant differences between the meander and straight sections during either campaign. Nevertheless during both seasons, much higher $K_h$ values and more outliers can be observed in the meander than in the straight section (Figure 5.6a).

![Histograms of ln($K_h$) and ln($K_v$) values](image)

**Figure 5.4:** Histogram of ln($K_h$) and ln($K_v$) values pooled together from the meander and straight sections for the measurement campaign of December 2011 and August 2012.

### 5.4.2 Vertical hydraulic conductivity

For both seasons at each measurement location, the lowest $K_v$ estimates were obtained by substituting the calculated $K_h$ values at the test locations in Eq. 5.1. The Hvorslev
solution (Eq. 5.1) with $m=1$ gave the highest $K_v$ estimates, using this solution results in an increase of 2.9% compared to the minimum estimates. This difference is considered small and as $K_v$ values calculated by assuming $m=3$ are closest to the mean $K_v$ of all five methods, these values will be used further on.

For the seasonal comparison, $K_v$ values were recalculated to a common reference temperature of 20 °C. The contour plots of $K_v$ from December 2011 and August 2012 both show elevated $K_v$ values at the upstream part of the meander section, closer to the stream bend (Figure 5.3c and d), with much higher values in August. For both campaigns the calculated $K_v$ values show a lognormal distribution (Figure 5.4c and d). In both seasons, the highest $K_v$ values are observed at the depositional left bank and towards the middle of the stream (Figure 5.5b). The August dataset, however, shows a much greater spatial variability across the stream than the December dataset. The Kruskal-Wallis test did not indicate statistically significant difference in $K_v$ across the stream channel in either section or between the meander and straight sections, but for both seasons a larger range of values can be observed at the meander section (Figure 5.6b).
Figure 5.6: Box plot of $K_h$ (A), $K_v$ (B), VHG (C) and anisotropy in K (D) values from the December 2011 and August 2012 campaigns in the meander and straight sections. On panel D the scale of the y axis did not allow for the plotting of outliers, these are shown on the plot with a red cross and the corresponding value.

5.4.3 Streambed sediments

Based on the sediment cores removed after the August field campaign, streambed sediments mainly consist of medium and coarse grained sands with a layer of organic material (Figure 5.7a). From bottom to top; first a layer of medium and coarse grained sand with occasional layers of gravel can be observed, followed by fine organic sediments overlain by an upper layer of medium and coarse grained sand (Figure 5.7a).

Figure 5.7: Schematic layering of the streambed material (A). A typical core with continuous organic layer of constant thickness (B), a core where the organic layer is discontinuous (C) or of changing thickness (D). The cores shown on the pictures are A27 (B), D38 (C) and A39 (D).
The organic layer was surveyed in August 2012 and usually detected 0.12-0.3 m below the streambed surface with thicknesses varying between 0.02 and 0.26 m (Figure 5.3), forming a well-defined layer. In some cases the residues of the original material (branches, roots) were still discernible giving a more discontinuous layer of variable thickness (Figure 5.7c, d) or the layer was missing resulting in a continuous sand column. At the depositional left bank of the stream, an additional porous organic layer was also observed above the upper sand layer at a few locations.

When \( K_v \) is compared to the thickness of the organic sediment layers, as measured on the retrieved sediment columns after the August field campaign, three distinct groups can be identified; i) test locations with high \( K_v \) and thin or missing organic layers, ii) test locations with elevated \( K_v \) and only an upper organic layer on top of the sediment column, and iii) test locations with low \( K_v \) and changing thickness of organic layer (Figure 5.8a). In the latter group test locations show a inverse relationship between \( K_v \) and the thickness of the organic layer (Figure 5.8b).

---

**Figure 5.8:** Scatter plot of the thickness of the organic layer and the calculated vertical hydraulic conductivity (A). For a better visualization, the relationship of \( K_v \) and the thickness of the organic layer at low hydraulic conductivities is shown in B.

### 5.4.4 Hydraulic conductivity estimated from grain size analysis

Out of the ten samples taken from the upper sand layer, six samples were classified as medium-grained and four as coarse-grained sand. The 11 samples taken from the lower sand layer consisted of mainly gravel in one case and of coarse and medium-grained sand
in three and seven cases, respectively. Comparing calculated \( K_g \) values of the lower sand layer to the \( K_h \) values derived from slugtests, all six empirical methods overestimated the slugtest-based hydraulic conductivity \( (5.1) \). The Schlichter, Terzaghi, and USBR formula gave the lowest \( K_g \) values, but also the closest estimate to the values calculated from slugtests, while the Kozeny and Hazen formula gave the highest estimates (Figure 5.9a and Table 5.1). The estimated \( K_g \) for the upper sand layer gives higher values and larger variability than \( K_g \) estimates for the lower sand layer (Figure 5.9b).

### Table 5.1: The dimensionless coefficient \( (C) \), effective grain diameter \( (d_e) \) and porosity function \( (\varphi(n)) \) used by the empirical formulas to calculate hydraulic conductivity \( (K_g) \) based on grain size distribution (Vukovic and Soro, 1992). Also shown are the arithmetic mean \( (X_a) \), the geometric mean \( (X_g) \) and the median hydraulic conductivity \( (\tilde{X}) \) calculated by the methods. \( K \) values are given in \( \text{md}^{-1} \).

<table>
<thead>
<tr>
<th>Formula</th>
<th>( C )</th>
<th>( d_e )</th>
<th>( \varphi(n) )</th>
<th>( X_a )</th>
<th>( X_g )</th>
<th>( \tilde{X} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kozeny</td>
<td>8.3x10^{-3}</td>
<td>( d_{10} )</td>
<td>( \frac{n^3}{(1-n)^2} )</td>
<td>44.6</td>
<td>42.3</td>
<td>50.2</td>
</tr>
<tr>
<td>Hazen</td>
<td>6x10^{-4}</td>
<td>( d_{10} )</td>
<td>1 + 10(n-0.26)</td>
<td>37.0</td>
<td>35.5</td>
<td>35.9</td>
</tr>
<tr>
<td>Schlichter</td>
<td>10^{-2}</td>
<td>( d_{10} )</td>
<td>( n^{3.287} )</td>
<td>13.8</td>
<td>13.2</td>
<td>14.7</td>
</tr>
<tr>
<td>Terzaghi</td>
<td>6.1x10^{-3}</td>
<td>( d_{10} )</td>
<td>( \frac{n^{0.13}}{\sqrt[3]{1-n}} )</td>
<td>17.6</td>
<td>16.8</td>
<td>18.9</td>
</tr>
<tr>
<td>USBR</td>
<td>4.8x10^{-4} \times d_{20}^{0.3}</td>
<td>( d_{20} )</td>
<td>1</td>
<td>18.3</td>
<td>15.4</td>
<td>11.4</td>
</tr>
<tr>
<td>Shepherd</td>
<td>142</td>
<td>( d_{50} )</td>
<td></td>
<td>35.8</td>
<td>25.2</td>
<td>20.5</td>
</tr>
<tr>
<td>( K_h ) from slugtests</td>
<td></td>
<td></td>
<td></td>
<td>10</td>
<td>5.2</td>
<td>5.6</td>
</tr>
</tbody>
</table>

Figure 5.9: Hydraulic conductivity values calculated by different methods on the basis of grain size distributions for the lower sand layer (A, n=11) and the upper sand layer (B, n=10) of the streambed. For the lower sand layer the results of slugtests (n=40) carried out in August 2012 are also shown.
5.4.5 Vertical head gradient

During both seasons VHG values show large upward gradients at the downstream part of the meander section. In December lower upward gradients are observed at the upstream part of the section, closer to the bend (Figure 5.3e and f). In December 2011 all test sites were characterised by an upward hydraulic gradient (Figure 5.3e), while in August 2012, a change from upward to downward gradients occurs in the former low gradient zone (Figure 5.3f). During both campaigns there is a clear distinction between the depositional left and erosional right bank of the stream, with highest upward gradients observed at the right bank and in the middle of the channel. VHG values show a more homogeneous distribution across the stream than $K_h$ and $K_v$ values (Figure 5.5a, b, and c). According to the Kruskal-Wallis test, differences in VHG are not statistically significant across the stream channel in either section or between the meander and straight section. For both seasons, the meander section shows a wider range of VHG values with both the highest upward and downward gradients observed, while the straight section shows higher median upward gradients (Figure 5.6c).

5.4.6 Streambed anisotropy

<table>
<thead>
<tr>
<th>Location</th>
<th>December 2011</th>
<th>August 2012</th>
</tr>
</thead>
<tbody>
<tr>
<td>All data</td>
<td>55</td>
<td>33</td>
</tr>
<tr>
<td>Straight</td>
<td>99</td>
<td>13</td>
</tr>
<tr>
<td>Meander</td>
<td>37</td>
<td>44</td>
</tr>
<tr>
<td>Left1</td>
<td>76</td>
<td>39</td>
</tr>
<tr>
<td>Left2</td>
<td>69</td>
<td>31</td>
</tr>
<tr>
<td>Middle</td>
<td>58</td>
<td>29</td>
</tr>
<tr>
<td>Right2</td>
<td>54</td>
<td>34</td>
</tr>
<tr>
<td>Right1</td>
<td>56</td>
<td>25</td>
</tr>
</tbody>
</table>

Table 5.2: Anisotropy ratios calculated by the geometric mean of $K_h$ and $K_v$ given for the whole dataset, the meander and straight sections and different locations across the stream for the December 2011 and August 2012 datasets.

For both campaigns the contour plots of the anisotropy in $K$ show both large spatial and temporal variability (Figure 5.3g and h). In December 2011 high anisotropy ratios were calculated for the left depositional streambank of the meander section. In August 2012, the anisotropy ratio shows larger variability with the highest values observed between the depositional left streambank and the middle of the channel (Figure 5.5d). According to the Kruskal-Wallis test, differences in the anisotropy ratios are only statistically significant between the December and August datasets, but not between the
meander and straight sections, where both datasets have similar median values (Figure 5.6d).

Mean anisotropy ratios were also calculated based on the geometric mean of $K_h$ and $K_v$ datasets (Table 5.2). These ratios show a significant reduction from December to August. The values also represent a consistent spatial trend across the channel with high anisotropy ratios observed at the left depositional bank and slowly decreasing towards the erosional right bank (Table 5.2).

### 5.4.7 Principal Component Analysis

![Biplot of PC1 and PC2](image)

**Figure 5.10:** Biplot of the first (PC1) and second principal component (PC2). The notation of the test sites is by the month (D: December, A: August) and the serial number. The bold brackets around the name of the test sites indicate that the test site is located in the straight channel.

The first and second principal components explained 54% of the variability observed in the datasets. The variables $K_v$, VHG and $K_h$ showed the highest loadings; 0.79, 0.70 and 0.60, respectively along the first principal component. The anisotropy ratio and $K_h$ displayed the highest loadings along the second principal component with 0.79 and 0.61, respectively. In the biplot, most test locations group around the origin showing similar behaviour with regards to the variables (Figure 5.10). However, some test locations are grouped farther away from the origin. These locations correspond to one
of the following, anisotropy ratios higher than 430, $K_h$ values larger than 20 md$^{-1}$, VHG larger than -0.2 m, or $K_v$ values larger than 3md$^{-1}$ (Figure 5.10).

The PCA analysis shows that $K_h$ and the elevation of the piezometer screen variables are correlated. This indicates that $K_h$ is indirectly related to the layered streambed sediment structure and as the piezometer screens were located in the lower sand layer also to the heterogeneous material properties within this material. Being on opposing sides of the biplot, $K_v$ and VHG values are inversely related, the higher the $K_v$ the smaller the gradient. The PCA analysis also shows that except for test locations D6 and A4, most of the deviations occur in the meander section and not in the straight section.

5.5 Discussion

Each streambed attribute shows great spatial heterogeneity even in the small 5 x 6 m meander section. Most of the spatial variability can be related to heterogeneity in streambed sediments and morphology. $K_h$ values are highest at the depositional left bank of the stream during both campaigns and show the lowest values near the middle of the channel (Figures 5.5a and 5.3a,b). The location of high $K_h$ values correspond to the depositional inner bend of the meander, where, due to the freshly deposited sediments, the sediments are more mobile. Similarly Genereux et al. (2008) linked increase and drop in $K_v$ to the deposition and erosion of sediments, respectively, while Käser et al. (2009) also found lower permeabilities in zones adjacent to eroding banks.

High values of $K_v$ were mostly observed at the upstream part of the meander section (Figure 5.3c,d) at the thalweg of the stream. The spatial distribution of $K_v$ agrees with the results of Genereux et al. (2008), who attributed changes in $K_v$ across the channel to changes in grain size, thus indirectly to water velocity. Correspondingly, Flewelling et al. (2012) also measured high specific discharge in the thalweg of a stream. In our study the location of high $K_v$ values is shifted towards the erosional streambank where due to the increased water velocity even the lower organic layer can be partially or completely removed by deep scouring of streambed materials.

Based on the comparison of the streambed attributes of the meander and straight section, it can also be concluded that streambed attributes are more variable in the meander bend than in the straight channel (Figure 5.6). This is due to the more dynamic environment of the meander bends where constant changes in water velocity, depositional environment and consequently streambed elevation result in more variability in streambed materials. This is also confirmed by the PCA analysis, which shows that most of the deviations from the bulk values can be observed at the meander bend, while
test locations in the straight channel are close to the origin of the biplot (Figure 5.10). Due to the dynamic environment near the meander bends and the homogeneous depositional environments along the stream channel in the meander section, streambed attributes also show larger variability across than along the stream channel (Figure 5.5 and 5.3). For a thorough characterisation of streambed attributes, more measurements are therefore necessary in a channel bend than in a straight channel section and measurements should ideally be carried out in transects across the channel.

There are several sources of uncertainties in the $K_h$ and $K_v$ measurements presented in this study. As for the instrumentation, the slugtests in the piezometers were carried out close (approx. 0.1 m) to the in-situ permeameter tests as this was the only way to get information of both variables from the same test location with minimal sediment disturbance. Slugtests were performed while the permeameter test was carried out. It is assumed, however, that the measurements did not influence each other due to their different time span. While the fastest permeameter tests finished in three hours, they usually lasted for more than a day and the slugtests were carried out in a few minutes, taking an hour at maximum.

Another uncertainty arises from the removal and reinstallation of the instrumentation between the December 2011 and August 2012 campaigns. Due to the high stream discharge and consequently the scouring processes, a permanent installation of instruments in the streambed was impossible. Based on GPS data, the average difference between the positions of the corresponding test locations in December and August was 0.21 m, less than the average spacing of 0.56 m between the test locations. This spatial difference also means that the instruments in August 2012 were not installed exactly at the same locations where measurements from December 2011 have already disturbed the streambed sediments.

$K_v$ values presented in the study were calculated according to Eq.5.1 assuming an $m$ value of 3, corresponding to an anisotropy ratio of 9. Our results, however, show that an anisotropy ratio of 25-70 (Table 5.2) is more realistic. Calculating $K_v$ values with an $m$ value of 10 only results in a decrease of $K_v$ values by 2.5%, this difference is not considered as significant.

The range of $K_h$ values calculated in this study (0.15-53.69 md$^{-1}$) correspond to values defined as clean sand by Freeze and Cherry (1979) as also seen in the grain size analysis. $K_v$ values (7.27$x10^{-4}$-6.05 md$^{-1}$), however, can be as low as those reported for low permeability clay sediments of 0.6-2.5 md$^{-1}$ (Chen, 2004) and they are considerably lower than mean values for sand in other studies 18.8-43 md$^{-1}$ (Chen, 2000), 17-45 md$^{-1}$ (Cheng et al., 2011), and 16.6 md$^{-1}$ for slightly silty sand (Dong et al., 2012). Although the measurement scale and direction of the slug and permeameter tests are different, the reason for the low $K_v$ values and their large variability is the presence
of the lower organic sediment layer as observed in many of the sediment cores (Figure 5.7). Even a thin layer of this low conductivity material can reduce $K_v$ considerably (Figure 5.8b). In some cases this layer is discontinuous (Figure 5.7c) and much higher $K_v$ values can then be observed (Figure 5.8a). These higher $K_v$ values are due to the absence or discontinuity of the lower organic layer. In cores A16, A19, and A24 a continuous transition between the upper and lower sand layer can be observed, while in core A18 the organic layer is present but is thin and possibly also discontinuous as in case of D38 (Figure 5.7b).

When compared with the calculated $K_v$, the upper and lower organic layers display different characteristics. While a deep-lying, thin continuous layer of the lower organic material decreases $K_v$ considerably, the upper layer of organic material of the same thickness just slightly reduces $K_v$ (Figure 5.8a). The difference in the behaviour of these materials is probably the degree of compaction of the organic material.

Based on the retrieved streambed sediment cores, the piezometers were all screened in the lower sand layer (Figure 5.7), thus $K_h$ values reflect spatial heterogeneity within this layer. $K_v$ values on the other hand indicate heterogeneities of the lowest permeability layer in the sediment column. According to the sediment cores (Figure 5.7) and confirmed by the permeameter measurements, this is the lower organic sediment layer of the streambed. Thus, the $K_h$ and $K_v$ values used to calculate the anisotropy ratio represent the characteristics of the lower sand layer and the harmonic mean of the sediment column, respectively, at each test location. Therefore the anisotropy ratios are not representative of one material, but are rather determined by $K_v$, and therefore related to the heterogeneity of the low-permeability organic sediment layers. Consequently, despite a similar range of $K_h$ values, the calculated anisotropy ratios (0.5-1654) exceed by one order of magnitude the values reported for sandy stream channels, a mean of 23-69 (Chen, 2000) and 4.1 (Chen, 2004). Landon et al. (2001) and Lu et al. (2012b) even observed anisotropy ratios below one 0.1-1.3 and 0.87-2.37, respectively that Landon et al. (2001) explained by the different measurement scale of the tests and sediment disturbance. At test locations without or with discontinuous, thin organic sediments (A4, A16, A18, A19, A24), an anisotropy of 0.5-5, typical for streambed sediments, were also found in this study (Table 5.3).

Comparing results from the two campaigns in December 2011 and August 2012, $K_h$ is more stable than $K_v$ and VHG, which are both showing high temporal variability (Figure 5.3a-f). During the December campaign, high $K_v$ values were observed close to the bank at the upstream part of the channel bend, while in August, higher $K_v$ values were measured towards the middle of the channel (Figure 5.3b). The range of $K_v$ values also changes; by about a factor of two from a geometric mean of 0.079 md$^{-1}$ in December to a geometric mean of 0.160 md$^{-1}$ during the August campaign.
This increase can partly be explained by the elevated water temperature (decreased water viscosity) in August. But, similarly to Genereux et al. (2008), a more likely explanation is that sedimentation and scouring processes related to high discharge events reorganized the sediment structure (Sebok et al., in prep.). The large difference in temporal variability between $K_h$ and $K_v$ (Figure 5.4) can thus be related to dynamic scouring and sedimentation processes. $K_h$ is measured 0.5 m below streambed in a relatively stable environment, but $K_v$ measurements include the topmost streambed sediment layer, which is highly mobile. This topmost layer in streams is not only affected by bedform migration, sedimentation, and scouring processes (Sebok et al., in prep.) but can also undergo occasional clogging or deposition of a fine-grained veneer (Rosenberry and Pitlick, 2009b).

Due to the presence of the organic layer, $K_v$ and VHG are also related (Figure 5.10) and as also found by Käser et al. (2009) the spatial distribution of VHG inversely follows the distribution of $K_v$. While $K_v$ shows an increase in geometric mean, VHG displays a decrease from the median value of -0.090 in December to a median value of -0.068 in August, when also downward gradients could be observed. Similarly, temporal changes in the calculated anisotropy are inversely correlated to $K_v$ (Figure 5.10), the geometric mean of anisotropy decreases between December 2011 to August 2012 from 55 to 33 (Table 5.2). While the distribution of $K_h$ remains fairly constant between the measurement periods, the distribution of anisotropy follows the distribution of $K_v$. This is best displayed in August 2012, when test locations with the highest $K_v$ values also display the lowest anisotropy (Figure 5.3d, h).

The six methods used to calculate hydraulic conductivity from grain size distributions all yield different results (Figure 5.9a,b). This variation between methods was also noted in other studies (Vukovic and Soro, 1992; Lu et al., 2012a) and estimated by Vukovic and Soro (1992) to be of a factor of up to 100. The underestimation of hydraulic conductivity based on empirical formulas was observed by several studies (Landon et al., 2001; Song et al., 2009). Overestimation, on the other hand, like in our case is rarely encountered (Cheong et al., 2008). The reasons for this could be that $K$ estimates from grain size analysis are small-scale, non-directional measurements as the sediment structure is destroyed.

Despite the different $K_g$ estimates, the dynamic sedimentation processes are also reflected in the $K_g$ values when comparing the lower sand (Figure 5.9a) to the upper sand (Figure 5.9b). The upper sand layer not only shows higher estimated $K_g$ values, but also higher variability. Song et al. (2007) also observed a similar distribution in $K_v$ values and explained it by hyporheic processes where the constant water exchange leads to enlarged pore space and consequently unconsolidated sediment structure. In the soft-bedded Holtum stream, however, this distribution is probably related to the constant
mobilization and redistribution of streambed sediments (Sebok et al., in prep.). Such processes also prevent the formation of a colmation layer, which would typically reduce $K$ in the uppermost sediment layers (Rosenberry and Pitlick, 2009b).

The highest VHG values were observed at the most downstream, eastern end of the meander section during both measurement campaigns (Figure 5.3e,f). In this area both the observed $K_h$ and $K_v$ values are low thus suggesting the presence of a low permeability material possibly hindering groundwater discharge to the stream. The highest groundwater fluxes can be expected at locations with high $K_v$ and VHG values. These areas indicate the absence of the organic sediment layer, thus facilitating upward groundwater flux. Both in December and in August such an area can be found at the upstream end of the meander section at the depositional left streambank (Figure 5.3c-f).

The modelling study of Salehin et al. (2004) showed that streambed heterogeneity results in a shallower hyporheic zone and higher hyporheic exchange rates. In this study, next to the large spatial heterogeneity in streambed hydraulic conductivity, the detected low-permeability organic sediment layer also influences the hyporheic flow, thus this study site most probably has a shallow hyporheic zone which does not extend below the organic layer located at a depth of 0.1-0.3 m below the streambed. The high-permeability lower sand layer and the high VHG however, also make it likely that there is also lateral flow below the organic layer which will surface where the low-permeability organic layer is discontinuous and of smaller thickness.

### 5.6 Conclusions

Streambed attributes of horizontal hydraulic conductivity ($K_h$), vertical hydraulic conductivity ($K_v$), vertical head gradient (VHG) and anisotropy were observed at 40 locations in a channel bend and in a straight stream section in December 2011 and August 2012. All streambed attributes show great spatial variability related to streambed materials and stream morphological environment even on the small scale at a streambed section of 6 x 5 metres. The highest $K_h$ is observed at the depositional inner bend, while the calculated $K_v$ is related to the thickness of the organic streambed sediment layer and shows the most elevated values at the upstream part of the meander section. The high $K_v$ values in these areas are related to high discharge events, when the high water velocity is more likely to erode the organic sediment layer hindering groundwater discharge. Due to the dynamic sedimentation and scouring processes on the streambed, $K_v$ showed higher temporal variability than $K_h$ which was measured 0.5 m below the streambed surface in a relatively stable environment.
Large differences in $K_h$, $K_v$, VHГ, and anisotropy were also observed across the stream channel downstream of the meander bend, but only the differences in $K_h$ values between the depositional left bank and the middle of the channel was statistically significant. Although there were no statistically significant differences in $K_h$, $K_v$, VHГ, and anisotropy of K between the straight channel and downstream of the channel bend, principal component analysis confirmed that the channel bend showed higher spatial variability in streambed attributes than the straight channel. This indicates that in meandering streams more measurements are necessary for the thorough characterisation of hydraulic parameters than in a straight stream. Variability in streambed attributes in the meander section was also greater across the stream than along the channel.

The study also showed that streambed material properties, in this study particularly the deep-lying organic sediment layers, have a large influence on $K_v$ and consequently on groundwater discharge to the stream. Even a small thickness of a continuous layer of organic sediments reduced $K_v$ values considerably which would have a similar influence on vertical groundwater fluxes. This organic layer also indicates a shallow hyporheic zone with lateral flowpaths beneath the low-permeability layer.

**Acknowledgments**

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Table 5.3: Horizontal hydraulic conductivity ($K_h$), vertical hydraulic conductivity ($K_v$), vertical head gradient (VHG) and anisotropy ($K_h/K_v$) measured at the individual test locations during the December 2011 and August 2012 campaigns. Also shown are the numbers of slugtests ($K_h$ tests) carried out at the test locations. $K_h$ and $K_v$ values are given in md$^{-1}$.
References


Chapter 6

Paper II

High-resolution Distributed Temperature Sensing to detect seasonal groundwater discharge into Lake Væng, Denmark

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Abstract

Distributed Temperature Sensing (DTS) was used to map spatial and temporal changes in temperature on a 25 m by 6 m lakebed area in the winter (February), spring (May), and summer (August) of 2012. A constant and high discharge of groundwater with the average temperature of around 8 °C to the lake will result in either lower (summer) or higher (winter) daily temperatures and reduce temperature variability at the sediment-water interface (SWI). DTS data were used as a proxy for groundwater discharge using three metrics; daily minimum temperature, diel amplitude, and daily standard deviation of temperature. During the seasons, the daily minimum temperatures at the SWI indicate a discharge zone 4-6 m off-shore. From winter to summer the extent of this zone changes and the SWI temperatures also show a shift of discharge locations towards the shore. Fluxes estimated on the basis of vertical temperature profiles from the top 50 cm of the lakebed and seepage meters in August compare well with the locations of the high discharge zones detected by the DTS in the same period, giving confidence in the ability of the method to map both the areas and spatial variability of groundwater discharge to lakes. Compared to February, the DTS was able to detect new relatively cold temperature zones at the SWI in May and August indicating that groundwater discharge to the lake changes over time and that DTS can be used to monitor temporal variability in areas of discharge.

6.1 Introduction

The ecology of lakes is affected by internal and external fluxes of nutrients; internal fluxes being for example nutrients diffusively released from bottom sediments, while external fluxes are from sources outside the lake, such as inlets or groundwater discharge (Shaw et al., 1990). Leaching of agricultural pollutants to groundwater in a lake catchment may therefore cause groundwater-dominated lakes to become eutrophic. Restoration is especially challenging for such eutrophic lakes, because it involves an assessment of where, when, and how much groundwater discharges to the lake.

A characterization of groundwater-lake interaction at the scale of the lake involves a description of physiography, i.e., the setting of the lake in a catchment in terms of regional geology and landscape (Winter, 1981). System geometries and parameters
such as the ratio of the width of the lake to the thickness of the aquifer, slope of lake bed, regional aquifer heterogeneities, anisotropy in hydraulic conductivity and how deep the lake penetrates the aquifer affect how a lake interacts with groundwater (McBride and Pfannkuch, 1975; Pfannkuch and Winter, 1984; Winter and Pfannkuch, 1984; Cherkauer and Nader, 1989; Genereux and Bandopadhyay, 2001). In simple, homogeneous and small aquifer-lake systems, the spatial distribution of discharge to a lake has been found to decrease exponentially with distance from the shore (McBride and Pfannkuch, 1975). More commonly, heterogeneities in hydraulic conductivity play a significant role and can cause off-shore peaks in groundwater discharge (Cherkauer and Nader, 1989; Schneider et al., 2005) and irregular patterns in discharge near the lake shore (Kishel and Gerla, 2002; Kidmose et al., 2013). Temporal variability in discharge makes groundwater-lake interactions even more complex. For example, rapid and large changes in discharge have been observed to be triggered by rainfall events (Schneider et al., 2005) leading to a sudden increase in the flux of nutrients from groundwater into lakes (Downing and Peterka, 1978).

Accurate water and chemical budgets are needed when designing lake restoration schemes in order to assess the precise impact of a given measure. The groundwater component of a lake water budget is often evaluated as the residual of the water balance, estimated from flow net approaches and isotopic tracers (LaBaugh et al., 1997), or by integrating site-specific measurements (e.g. seepage meters) for the whole lake (Shaw et al., 1990). This leads to uncertainty in the water and chemical budgets (Shaw et al., 1990) and sometimes to conflicting results (LaBaugh et al., 1997). It is therefore desirable to supplement existing methods with approaches that can map discharge zones at high-resolution in both space and time so that well-designed field investigations can be carried out to obtain more accurate estimates of groundwater discharge into the lake.

Groundwater discharge can be measured using different methods, for example seepage meters have been often used to directly measure the vertical flux into a lake (Lee, 1977; Schneider et al., 2005; Rosenberry et al., 2010; Kidmose et al., 2011). Heat as a tracer has also been successfully applied in many studies of groundwater-surface water interactions although mostly with respect to streams (Anderson, 2005; Constantz, 2008). These heat-based studies have been used to not only locate discharge zones, but also to estimate the discharge flux from vertical streambed temperature profiles (Schmidt et al., 2007; Jensen and Engesgaard, 2011) or from time series analysis (Hatch et al., 2006; Keery et al., 2007). There are a few examples from lakes where temperature profiles have been used to compare (Kishel and Gerla, 2002) or to estimate (Kidmose et al., 2013) vertical groundwater fluxes.

Many of the existing methods only measure fluxes over a relatively small area, thus
multiple measurements are needed to represent the spatial distribution of discharge to a section of a lake. Distributed Temperature Sensing (DTS) is a quick and non-invasive approach for mapping of temperatures at the Sediment-Water-Interface (SWI) (Selker et al., 2006a). The fiber optic cable of a DTS system is capable of measuring temperature with a resolution of up to 0.01 °C and with a spatial-averaging interval of 0.25 m; detailed descriptions of the method are given in for example Selker et al. (2006a), Tyler et al. (2009), and Suarez et al. (2011).

DTS has been widely applied to detect groundwater discharge to streams (Selker et al., 2006b; Westhoff et al., 2007; Briggs et al., 2011; Krause et al., 2012), wetlands (Lowry et al., 2007) and to study the thermal variations of saltmarsh channels (Moffett et al., 2008). However, only a few studies have explored the use of DTS in lakes, for example Selker et al. (2006a) recorded temperatures at the bottom of Lake Geneva across the almost five km wide lake and Suarez et al. (2011) studied the thermal evolution of the water column in an experimental solar pond.

The use of DTS in lakes differs from stream applications in several ways. Due to the absence of stream flow, the installation of the cable in lakes is less complicated and for long-term measurements the effects of sedimentation and scouring are less prominent than can be the case in soft-bedded streams. A major challenge in lakes is the changing water depth in the littoral zone (where discharge typically occurs). The same discharge rate at a depth of 10 cm (near the shore line) and 100 cm (further off-shore) can result in two completely different temperature recordings due to the different effect of solar heating of the lake bottom. Due to this special environment that lakes represent for DTS applications it is therefore necessary to study how external conditions such as solar heating affect the DTS temperature signals and their interpretation.

Next to the mixing analysis (Selker et al., 2006b; Westhoff et al., 2007) usually carried out in DTS studies to quantify groundwater discharge, only Briggs et al. (2011) have compared DTS results in rivers with traditional methods like stream gauging and fluxes determined by geochemical mixing models and Lowry et al. (2007) used seepage meter measurements at three locations to verify concentrated groundwater discharge sites as detected by DTS. Nevertheless, to the knowledge of the authors the spatial variability in groundwater discharge indicated by DTS has never been compared to the spatial variability captured by traditional punctual measurements, for instance seepage meter measurements and fluxes estimated from vertical lakebed temperature profiles, methods widely used in lake studies to estimate groundwater fluxes. As the focus of previous DTS lake studies was not on detecting possible groundwater discharge areas, our main motivation was to see if DTS can be used in lakes to confidently map spatial and temporal variations in temperature signals thereby indirectly recording similar variations in discharge. The objectives of this study therefore have been to; (1)
explore the use of DTS in lakes by combining measurements at the SWI and at several depths in the water column, (2) perform high-resolution measurements at the SWI during three seasons (winter, spring, summer) to map spatial and temporal changes in discharge zones, and (3) to compare discharge areas detected by DTS with other proxies for groundwater discharge. These proxies are; maps of (i) indirectly estimated discharge from vertical lakebed temperature profiles using solutions to a 1D analytical heat transport equation, (ii) directly measured discharge from seepage meters, and (iii) ice thickness during winter seasons.

6.2 Field site

Lake Væng is located in the upper part of Gudenå River catchment in central Jutland, Denmark (Figure 6.1a). The topographic catchment area of the lake is 10.7 km², excluding the lake area of 0.16 km². The lake has mean and maximum water depths of 1.2 and 1.9 m, respectively (Jeppesen et al., 1998), and an average volume of 187,000 m³.

Lake Væng has the longest history of attempted lake restoration in Denmark. The main ecological problem is related to very high Total Phosphorous (TP) concentrations. TP concentrations in the lake water were recorded monthly from 2008 to 2009 and ranged between 36 to 196 μg TP l⁻¹. Kidmose et al. (2013) measured an average TP concentration of 162 μg l⁻¹ in discharging groundwater sampled from shallow boreholes around the lake, indicating a correlation between lake water and groundwater concentrations.

Precipitation, calculated by the Danish Meteorological Institute for a 10 by 10 km grid cell including Lake Væng, averages 787 mm yr⁻¹ over the period 2000 to 2010 (Scharling, 1999). Lake Væng has three minor inlets in the northern part and one major outlet in the southern end. The total discharge from the inlets was measured in 2011/2012 as part of this study and ranged between 10.8 to 47.0 ls⁻¹. The discharge at the outlet was continuously recorded from 2000 to 2008 and ranged between 135-165 ls⁻¹ (Nilsson et al., 2010). Kidmose et al. (2013) reported that groundwater represents 66%, the three surface inlets 30%, and precipitation 4%, respectively, of the total water input to Lake Væng in 2009. Hydraulic retention time in the lake was estimated to be approximately 20 days (Kidmose et al., 2013); Lake Væng is thus a groundwater-dominated lake with upward groundwater fluxes everywhere in the study section on the eastern shore (see below).

The lake is connected to an unconfined aquifer (Figure 6.2) consisting of medium to coarse sands and gravels of Pleistocene glacio-fluvial origin, deposited in a north-
west, southeast trending erosional valley (GEUS). The valley cuts through previously
deposited Pleistocene clays and sands, Miocene sand-rich fluvial and coal-bearing de-
posits, and Miocene marine clay-rich deposits with insets of fine sand (Rasmussen
et al., 2010). The low permeability deposits are not present as continuous layers and
the Miocene and Pleistocene sand aquifers are hydraulically connected. The lower-
lying parts of the Pleistocene valley, including the bottom of the lake, are topped by
organic sediments of up to 9 m thickness (Kidmose et al., 2013).

Figure 6.1: Map of the study area showing the location of Lake Væng in Denmark (A)
and its catchment with the topography and the isopotential map (both in meters above
sea level) with directions of groundwater flow (B). A map of the study area (C) shows
the location of the DTS temperature measurements at the SWI (1) and the location
of multi-level water column measurements (2). A sample layout of the SWI DTS loop
from February 2012 and the locations of vertical temperature profiles from July 2010
and February 2012 are shown on panel D.

Groundwater in the catchment flows mainly from west-southwest to east-northeast
(Figure 6.1b). Groundwater levels range between $\sim 110$ m above sea level (m.a.s.l) in
the southern part of the catchment to $\sim 26.4$ m a.s.l at Lake Væng. The lake surface
only changed a few centimetres in 2011/2012.

The eastern shoreline of this lake was chosen as the location of this study, because
seepage meter measurements and 2D groundwater flow modeling by Kidmose et al.
(2013) showed that discharge at this shore can vary in space by as much as a factor
of 50. Moreover, discharge may even increase exponentially with distance from the
shoreline, opposite to what is usually observed. This is due to organic sediments at the bottom of the lake creating a flow barrier (Kidmose et al., 2013), thus part of the groundwater flowing from west bypasses the lake and discharges near the eastern shore (Figure 6.2). In addition there is a water table divide east of the lake resulting in a small groundwater contribution from this side (Figure 6.1b and 6.2).

Figure 6.2: Geological W-E cross-section of the study area, see Figure 6.1b for location.

6.3 Methods

Distributed Temperature Sensing was used to locate areas of groundwater discharge in the littoral zone of the eastern shoreline (Figure 6.1d and 6.2). These temperature observations were compared with other methods used to estimate groundwater discharge; vertical temperature profiles in the upper 0.5 m of the lakebed to indirectly estimate fluxes, seepage meter measurements to directly measure groundwater discharge, and ice thickness measurements as a proxy for high groundwater discharge areas. Table 6.1 summarizes the different measurements carried out at the study site.

6.3.1 Distributed Temperature Sensing

DTS was applied in two ways; (1) in a looped layout at the SWI covering a horizontal area of approximately 25 m by 6 m (Figures 6.1c and 6.1d) and (2) in a looped layout in the water column covering a volume of the lake of 21 m by 2 m by 0.6 m slightly north of the area used for the DTS on the SWI (Figures 6.1c and 6.3). The spatial extent of the DTS study section at the SWI was largely determined by the ice thickness observations of February 2010 and 2012 (Section 6.3.4). The two DTS sites did not overlap in order to minimize sediment disturbance.
Figure 6.3: The layout of the DTS cable for the multi-level lake water temperature measurements. The location is shown on Figure 6.1d.

<table>
<thead>
<tr>
<th>Type of measurement</th>
<th>Time of measurements</th>
<th>Number of measurements</th>
<th>Mean flux (md⁻¹)</th>
<th>Minimum flux (md⁻¹)</th>
<th>Maximum flux (md⁻¹)</th>
</tr>
</thead>
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<tr>
<td>DTS at the SWI</td>
<td>February 2012</td>
<td>10</td>
<td>0.14</td>
<td>0.03</td>
<td>0.21</td>
</tr>
<tr>
<td></td>
<td>May 2012</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>August 2012</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Multi-level lake water DTS</td>
<td>May 2012</td>
<td>10</td>
<td>0.14</td>
<td>0.03</td>
<td>0.21</td>
</tr>
<tr>
<td>Temperature profiling</td>
<td>July 2010</td>
<td>4</td>
<td>1.21</td>
<td>0.35</td>
<td>3.46</td>
</tr>
<tr>
<td></td>
<td>February 2012</td>
<td>65</td>
<td>0.16</td>
<td>0.06</td>
<td>0.35</td>
</tr>
<tr>
<td></td>
<td>August 2012</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Seepage meter</td>
<td>January 2010</td>
<td>4</td>
<td>2.07</td>
<td>0.2</td>
<td>6.3</td>
</tr>
<tr>
<td></td>
<td>August 2012</td>
<td>39</td>
<td>0.15</td>
<td>0.0027</td>
<td>0.62</td>
</tr>
<tr>
<td>Ice thickness</td>
<td>February 2010</td>
<td>244</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>February 2012</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

a) Measurements at ice hole and ice free locations  
b) Measurements at ice hole locations  
c) Triplicate measurements  

Table 6.1: Measurements taken at the field site

DTS: Temperature measurements at the SWI

A Damsense fiber optic cable attached to a Sensornet DTS system (Oryx DTS-SR) was deployed on the lakebed between the 17-19 February, 21-22 May, and 3-5 August 2012 to map groundwater discharge zones at different times of the year with different contrasts in temperature between lake water and groundwater. However, in the analysis complete 24 h data sets from 0.00 am to 12 pm on 18 February, from 11.30 am on 21 May to 11.30 am on 22 May, and from 0.00 am to 12 pm on 4 August 2012 are compared. In the double-ended measurements, the spatial averaging interval was 1.01 meters along the cable with 0.12 °C precision, integrated over 20 minutes. The temperature offset calibration of the double-ended installations was carried out by having 30 meters of fiber optic cable in a calibration bath at the lakeshore (Figure 6.1d).

The fiber optic cable was fixed to the lakebed in a looped layout with iron staples.
(Figure 6.1d) and deployed at the same study section during the three campaigns, however the individual layouts were slightly different. Each setup contained 6-8 unevenly spaced cable rows with an average of 0.5 m separating the rows further off-shore and approximately 1 m spacing closer to the shore. Due to emergent vegetation no temperature measurements were taken directly at the shoreline. The DTS layout thus recorded an average temperature approximately at each meter along the 25 m study section and every 0.5-1 m cross-shore.

Figure 6.4: Panel A shows how the daily mean, minimum, maximum temperatures and the diel amplitude were calculated at a measurement location for a 24 hour period in May. Note that the low-discharge site has a lower diel amplitude than the high-discharge site. The special coloring scheme applied to the variables to emphasize anomalies is presented on the February 2012 dataset (B) where the measured daily minimum temperatures at each location along the fiber optic cable are shown. The value for delineating the discharge areas (dashed line) was calculated by adding to the spatial mean of daily minimum temperatures at each location along the fiber optic cable shown by the blue line) the standard deviation of this dataset. This value was plotted as a contour line on the interpolated temperature map (Figure 6.5a) and used to separate possible high discharge zones from low-discharge zones. Panel C shows an ice hole of 0.3-0.5 m in diameter with the sand rings in motion under the ice cover (D).

Temperature data were analyzed on the basis of three metrics; the diel amplitude (difference between daily minimum and maximum temperatures), the daily standard deviation, and the daily minimum temperature (Figure 6.4a). Deeper groundwater in Denmark has an average temperature around 8 °C; therefore high discharge zones should show relative warmer areas at the SWI in the winter time and vice versa in the
summer time. Furthermore, if discharge is stable then the changes in temperature at the SWI are reduced by the constant input of heat from groundwater. In other words, the diel amplitude or daily standard deviation of temperature should be lower in these areas. Spatial variability in groundwater discharge was visualized using interpolated maps of the metrics. A special coloring scheme was applied to emphasize the deviations from the spatial mean of the dataset (Figure 6.4b).

The discharge areas were delineated by taking the spatial mean of the daily minimum temperatures at the SWI plus one standard deviation of the daily minimum temperatures (February, shown on Figure 6.4b) or minus one standard deviation of the daily minimum temperatures (May and August). Discharge zones were named after the month they were detected and indexed based on their location in the study area (e.g. F_W is discharge in the Western zone in February).

**DTS: Multi-level lake water temperature measurements**

The same type of DTS system and fiber optic cable was fixed in a vertical looped layout (Figure 6.3) to the north of the SWI temperature measurements (Figure 6.1c). This setup allowed detection of any temperature stratifications or other variations arising from water circulation and enabled us to determine if temperature differences measured by the DTS at the SWI could be caused by changes in lake water temperature. The multi-level DTS system recorded temperatures between the 22-24 May 2012 with a spatial-averaging interval of 1.01 m along the cable with 5 min. integration times and 0.14 °C precision at 0.1, 0.3, and 0.6 m depth from the lake surface. The cable was placed 5, 6, and 7 m from the shore at these three different depth levels. The analysis of the data was carried out with the same metrics as the DTS temperature data from the SWI.

**6.3.2 Temperature profiling**

Vertical groundwater fluxes were calculated indirectly by fitting an analytical solution to measured vertical lakebed temperature profiles recorded at the SWI study area at 10 locations in July 2010, four locations in February 2012 (just prior to the first DTS campaign), and 65 locations in August, 2012 (immediately after the final DTS campaign). Thus, there was no data collection in the study section in-between DTS campaigns in order to minimize disturbance of the lakebed.

The probes were installed by the direct push method. In each profile the temperature sensors were placed at 0, 0.025, 0.05, 0.075, 0.1, 0.15, 0.2, 0.3, 0.4, and 0.5 m depths from the lakebed. The sensors were made of thermocouples that can measure
6.3.2 Parameters

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Unit</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Density of the fluid</td>
<td>( \rho_f )</td>
<td>(kg m(^{-3}))</td>
<td>999.73</td>
</tr>
<tr>
<td>Specific heat capacity of the fluid</td>
<td>( c_f )</td>
<td>(J kg(^{-1}) °C(^{-1}))</td>
<td>4192</td>
</tr>
<tr>
<td>Effective thermal conductivity</td>
<td>( \kappa_e )</td>
<td>(J m(^{-1}) s(^{-1}) °C(^{-1}))</td>
<td>1.84</td>
</tr>
<tr>
<td>Groundwater temperature</td>
<td>( T_g )</td>
<td>(°C)</td>
<td>8</td>
</tr>
<tr>
<td>Depth of constant groundwater temperature</td>
<td>( L )</td>
<td>(m)</td>
<td>5</td>
</tr>
</tbody>
</table>

Table 6.2: Parameters used in the steady-state analytical solution

with an accuracy of 0.2 °C.

Bredehoeft and Papadopulos (1965) gave the steady-state analytical solution to the one-dimensional conduction-convection equation often used to interpret temperature profiles (Schmidt et al., 2007; Jensen and Engesgaard, 2011);

\[
T(z) = T_s + (T_g - T_s) \frac{\exp[N_{pe}z/L - 1]}{\exp(N_{pe} - 1)}
\]  

(6.1)

where \( T_s \) is the surface water temperature (°C), \( T_g \) is the temperature of the groundwater (°C) at a given depth \( L \) (m), \( z \) is the depth (m) and \( N_{pe} \) is the Peclet number describing the ratio of convection to conduction:

\[
N_{pe} = \frac{q_z \rho_f c_f L}{\kappa_e}
\]  

(6.2)

where \( q_z \) is the vertical fluid flux in m s\(^{-1}\), \( \rho_f c_f \) the volumetric heat capacity of the fluid (J m\(^{-3}\) °C\(^{-1}\)) and \( \kappa_e \) is the effective thermal conductivity (J m\(^{-1}\) s\(^{-1}\) °C\(^{-1}\)).

Vertical fluxes were calculated by fitting the analytical solution (Eq.6.1) to the observed temperature data with the parameters given in Table 6.2. The values of the variables were chosen based on Jensen and Engesgaard (2011). The estimated fluxes were insensitive to reasonable choices of the depth of constant groundwater temperature (\( L \)). In all cases, the instantaneous temperature measured at the lakebed by the upper thermocouple was used as the upper boundary condition.

6.3.3 Seepage meter

Vertical fluxes were measured by seepage meters (Lee, 1977) at the four ice hole locations in January 2010 and again in August 2012 at 39 locations after the final DTS campaign. The seepage meters had a diameter of 57.5 cm and heights of 10, 15, or 25 cm. The top of the seepage meter was installed 2-5 cm above the lakebed and a plastic bag was pre-filled with approx. 0.5 l lake water. A plexi-glass or a plastic box was fitted above the bag connected to a valve to prevent velocity-head effects (Rosenberry, 2008). Measurements typically lasted from 20 min. up to 1-2 hours and were repeated...
three times. The mean of triplicate seepage meter measurements at each location at the DTS study section is presented. The seepage meter data was corrected with a factor of 1.1 because of friction loss in the seepage meter (Rosenberry, 2008).

### 6.3.4 Ice thickness measurements

Ice thickness data were used as a proxy for groundwater discharge with the assumption that high discharge of relatively warmer buoyant groundwater would decrease ice thickness or in some cases even prevent ice formation resulting in holes of 0.3-0.5 m diameter in the ice cover of the lake as observed in February 2010 and 2012 (Figure 6.4c). As an indicator of concentrated groundwater discharge, sand rings in motion of 0.05 m diameter (Figure 6.4d) were sometimes observed at the lake bottom under the ice holes.

Ice thickness data were collected in the study section in February 2010 at 244 locations aligned in four transects each running 30 meters perpendicular to the shoreline. In February 2012 only the locations of the ice holes were marked before the removal of the ice cover and deployment of the fiber optic cable on the lake bed (first DTS campaign).

### 6.4 Results

#### 6.4.1 DTS

The results are presented in terms of interpolated maps of temperature and daily variations in temperature (diel amplitude and daily standard variation). The daily minimum temperature maps represent the temperature observed at a specific location during a day. Although the whole dataset is used for the interpolation, the interpolated maps only show the overlapping section of the three campaigns as the locations of the fiber optic cable were different.

**Temperature measurements at the SWI**

Figure 6.5a shows the daily minimum temperatures at the SWI in February (recorded at 5 pm). The highest temperatures are observed in the Western ($F_W$) part of the study section compared with the spatial average of the whole data set (Table 6.3). The daily standard deviation (Figure 6.6a) and diel amplitudes (Figure 6.7a) in $F_W$ are lower when compared with the spatial average of the data set as well. All three
Figure 6.5: Daily minimum temperatures at the SWI in February 2012 (A), (B), and August 2012 (C) with indication of the location of possible groundwater discharge zones. Temperature values are given in °C. The black dots denote the approximate DTS measurement points where an average temperature along 1.01 m of cable is collected. Measurements are available from different areas and interpolation was made for the whole area. Only the section where the common areas of the three DTS campaigns overlap with the other flux measurements is shown.

indicators of discharge (temperature at the SWI, diel amplitude and daily standard deviation) suggest a spatial distribution of groundwater discharge, where the area at FW is a high discharge zone.

In May the DTS layout consisted of eight cable rows. Figure 6.5b shows the daily minimum temperatures (recorded at 5am). The lowest temperatures were observed at the Western (MW) edge of the study section, where the minimum temperatures at the SWI are much lower compared to the spatial average of the data set (Table 6.3). A similar low temperature zone can be detected closer to the shoreline in the middle of the section (MM). This zone is new compared to February. These two areas (MW and MM) have average daily standard deviations (Figure 6.6b) and diel amplitudes (Figure 6.7b) slightly higher than the data set average (Table 6.3); thus contradicting the results based on the daily minimum temperatures.

In August the DTS layout consisted of eight cable rows. Figure 6.5c shows the daily minimum temperatures recorded at 8 am. The results indicate small distinct cold areas at the Western (AW) and Northern (AN) end of the study section with a lower average minimum daily temperature compared to the whole data set (Table 6.3). However, the lowest minimum temperatures were observed in the Eastern end of the section (AE) (Table 6.3), this area has now emerged as a new possible discharge zone. The average standard deviation and diel amplitude of the whole data set were higher
Figure 6.6: Daily standard deviation of temperature at the SWI in February 2012 (A), in May (B), and August 2012 (C). Standard deviation values are given in °C. The solid black line indicates the areas where according to the daily minimum temperatures possible groundwater discharge zones are detected (Figure 6.5). The black dots denote the approximate DTS measurement points where an average temperature along 1.01 m of cable is collected. Measurements are available from different areas and interpolation was made for the whole area. Only the section where the common areas of the three DTS campaigns overlap with the other flux measurements is shown.

Figure 6.7: Diel amplitude of temperature at the SWI in February 2012 (A), in May (B), and August 2012 (C). Diel amplitude values are given in °C. The solid black line indicates the areas where according to the daily minimum temperatures possible groundwater discharge zones are detected (Figure 6.5). The black dots denote the approximate DTS measurement points where an average temperature along 1.01 m of cable is collected. Measurements are available from different areas and interpolation was made for the whole area. Only the section where the common areas of the three DTS campaigns overlap with the other flux measurements is shown.
than the standard deviation and diel amplitude observed in $A_W$ and $A_N$, but only slightly higher than in $A_E$ (Figure 6.6c, 6.7c and Table 6.3). All indicators agree on the location of $A_W$ and $A_N$ discharge zones, but the $A_E$ zone is not confirmed by the daily standard deviation and diel amplitude.

<table>
<thead>
<tr>
<th>Dataset location</th>
<th>Minimum daily temperature</th>
<th>Daily standard deviation</th>
<th>Diel amplitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>February 2012</td>
<td>Western ($F_W$)</td>
<td>4.89</td>
<td>0.29</td>
</tr>
<tr>
<td></td>
<td>Study site</td>
<td>4.19</td>
<td>0.46</td>
</tr>
<tr>
<td>May 2012</td>
<td>Western ($M_W$)</td>
<td>11.73</td>
<td>0.79</td>
</tr>
<tr>
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Table 6.3: Metrics for the February 2012, May 2012, and August 2012 dataset, all values represent a spatial mean and are given in °C.

Multi-level lake water measurements

![Multi-level lake water measurements](image)

Figure 6.8: Water column temperatures measured along the cable with distances shown on the x-axis and time on y-axis. Distance is shown from the cable row located 5 meters from the shoreline first at 0.1, then 0.3 and finally 0.6 m depth from the water surface, followed by the cable rows 6 and 7 meters from the shoreline in a similar way (see also Figure 6.3). The sections in between these cable rows with the highly variable temperature are the sections of cable placed in the air.

Figure 6.8 shows the DTS results over 1.5 days in May. During this period the air temperature varied by almost 13 °C ranging from 13 °C to 26 °C (Figure 6.4a).
During daytime, the DTS clearly detects a thermal layering in the water column. The water column warms up after sunrise, but with a time lag between the different depths. After sunset the temperature distribution in the water column rapidly becomes homogeneous, but with a slow cooling off period. Figure 6.8 also shows a homogeneous water temperature distribution irrespective of depth at nighttime implying that the daily minimum SWI temperatures recorded during the night (and thus used in this study to detect discharge zones in May and August) are not influenced by solar radiation. The DTS detected a possible discharge site 5 m offshore (Figure 6.8) recording consistently low temperatures during the measurement period.

### 6.4.2 Heat transport modeling

Vertical fluxes were first estimated on the basis of 10 vertical lakebed temperature profiles recorded in July 2010 (Figure 6.1d), 1.5 years before the first DTS campaign. The average flux is 0.14 m$^{-1}$ with a maximum flux of 0.21 m$^{-1}$ (Table 6.1).

In February 2012, immediately prior to the first DTS campaign, three temperature profiles D3-5 (Figure 6.1d) were measured close to the lake shore, while profile D2 was recorded in an ice hole (H3, see below), located in the middle of the DTS section. The highest flux of 3.46 m$^{-1}$ was measured at D2, with fluxes at the shoreline being an order of magnitude lower; 0.66 m$^{-1}$ at D3, 0.35 m$^{-1}$ at D4, and 0.39 m$^{-1}$ at D5.

![Figure 6.9: Vertical groundwater fluxes calculated on the basis of vertical lakebed temperature profiles (A) and fluxes obtained by seepage meter measurements (B). Note the logarithmic scale on the y axis, fluxes increase upwards.](image)

Figures 6.9a and 6.10c show the estimated fluxes in the study section in August.
2012 (65 profiles). Observed fluxes ranged between 0.06 and 0.35 md$^{-1}$ and show great spatial heterogeneity (Table 6.1 and Figure 6.10c), the highest fluxes were recorded at the western and northeastern end of the study section (spatial mean of 0.31 md$^{-1}$).

### 6.4.3 Seepage meter

In January 2010 seepage meter measurements in the ice holes showed fluxes of 0.2 md$^{-1}$ (HA), 6.3 md$^{-1}$ (HB), 3.2 md$^{-1}$ (HC), and 2.6 md$^{-1}$ (HD) (Table 6.1). Based on triplicate measurements at 39 locations, vertical fluxes in the DTS section ranged between 0.0027 and 0.62 md$^{-1}$ (Table 6.1 and Figure 6.9b) in August 2012. The maximum groundwater discharge of 0.62 md$^{-1}$ was measured at the western end of the study section (Figure 6.10b).

![Figure 6.10: Contour maps of the minimum temperatures at the SWI recorded by the DTS (A), the vertical fluxes measured directly by seepage meters (B) and the vertical fluxes calculated on the basis of temperature profiles (C) during the August 2012 field campaign. Temperature values are given in $^\circ$C and vertical fluxes in md$^{-1}$. On panel A the black dots denote the approximate DTS measurement points where an average temperature of 1.01 m along cable is collected. On panels B and C the black dots show the location of the seepage meter or vertical temperature profile measurements, respectively. Measurements are available from different areas and interpolation was made for the whole area. Only the section where the common areas of the three DTS campaigns overlap with the other flux measurements is shown.](image)

### 6.4.4 Ice observations

Figure 6.11 shows a map of the ice thickness and the location of ice holes as observed in February 2010 and in February 2012, two years later. The maximum ice thickness
in the DTS section in 2010 was 0.16 m with four ice-free locations; ice holes HB, HC, and HD grouped together in the northern part of the study section with HA located in the southern end. The locations of HB and HC match the locations of H4 and H3 recorded in 2012 with a small change of approximately 0.1 m and 0.4 m. The HA and HD ice holes were not found again in 2012, instead two new ice holes had appeared (H1 and H2).

Figure 6.11: Contour map of the ice thickness from February 2010 with the observed location of the ice holes in February 2010 and 2012 and the delineated discharge zones for each study period. Measurements are available from different areas and interpolation was made for the whole area. Only the section where the common areas of the three DTS campaigns overlap with the other flux measurements is shown.

6.5 Discussion

Compared to stream applications, the quasi-motionless water body of lakes represents a special environment for DTS applications. Due to the incoming shortwave radiation the heating of the water column and the SWI is inevitable. However, for the DTS applications a more challenging problem is the heating of the fiber optic cable which will record temperatures in excess of the bulk temperature of the SWI it is placed on, referred to as excess temperature (Neilson et al., 2010). The excess temperature depends mainly on the intensity of solar radiation, water depth and turbidity, the color
of the fiber optic cable and the lake bed substrate. Neilson et al. (2010) found that due to water flow and mixing of heat, solar radiation in a stream only resulted in an excess temperature of 0.01-0.17 °C. Due to the decreased water velocity the mixing of heat in the water column is less typical in lakes (Figure 6.8) thus the excess temperature can be higher and even more strongly related to water depth.

As in this study the North-South oriented lakebed section only had topographical shading, it is assumed that lake water temperatures measured at the same depth are exposed to the same solar heating. Indeed, the DTS shows the influence of water depth and solar radiation on water temperature resulting in a gradual thermal layering during the sunshine hours and its rapid disappearance during the night (Figure 6.8). Thus, only night measurements can reliably detect groundwater discharge during periods with high solar radiation (to avoid uncertainties arising from excess temperatures).

While in this study the effect of solar radiation is negligible in winter, this is not the case for the May and August campaigns. For the interpretation of the May and August data however, the daily minimum temperatures were used. These temperatures were recorded during the night when the multi-level lake water measurements revealed a relatively homogeneous temperature distribution in the water column (Figure 6.8), not showing the influence of solar radiation.

However, both during the day and night measurements the multi-level lake water DTS detected significant groundwater discharge at one location at 0.6 m depth (Figure 6.8) showing the cooling effect of groundwater despite the solar radiation. During the winter, the positively buoyant groundwater prevents the formation of ice cover at a concentrated groundwater discharge site, therefore a dense layout of multi-level water column DTS together with DTS at the SWI could be applied to examine such three-dimensional thermal plumes.

Lakebed temperature measured by DTS at the SWI as a proxy for groundwater discharge was evaluated using three metrics; (1) the minimum daily temperature, (2) the diel amplitude in temperature (maximum minus minimum daily temperature), and (3) the daily standard deviation of temperature.

The minimum temperature maps in February, May, and August (Figure 6.5) show high spatial variability indirectly reflecting a similar variability in discharge. The daily standard deviation and diel amplitude maps (Figure 6.6 and 6.7) also show great spatial and temporal variability, but also a distribution very similar to each other. In February the lowest daily standard deviation and diel amplitude (Figure 6.6a, 6.7a and Table 6.3) match the discharge zones indicated by warm temperatures at the SWI (Figure 6.5a). Similarly, in August the minimum SWI temperatures at the $A_W$ and $A_N$ areas agree well with low daily standard deviations and diel amplitudes (Figures 6.5c, 6.6c and 6.7c). There is however a disagreement between the three metrics in the case of $M_W$.
and M_M zones in May and the A_E zone in August. In May, the lowest temperatures were recorded in M_W and M_M, approx. 0.5 m apart from the area characterized by the highest daily standard deviations and diel amplitudes, while both the lowest daily standard deviations and lowest diel amplitude (Figure 6.6b, 6.7b and Table 6.3) were observed in the southern part of the area.

The results in May can possibly be explained by a combination of the almost motionless shallow water and the high solar radiation due to warm and clear weather (Figure 6.4a). The cooling influence of groundwater discharge at the high-discharge locations resulted in generally lower SWI temperatures, although due to solar radiation these sites were also heated during the day. But after sunset, as opposed to low-discharge sites, the cooling effect of groundwater at high-discharge sites caused a more rapid and intense cooling (Figure 6.4a), leading to increased daily standard deviation values and diel amplitudes as shown on Figure 6.4a and quantified in Table 6.3.

In August in the A_W and A_N zones the daily standard deviation and diel amplitude confirm the results of the minimum SWI temperatures. The difference in the agreement of the three metrics between May and August can possibly be due to differences in solar input. Although the lake water temperature is the same during the two campaigns, there are large differences in the air temperature and solar radiation. During May, clear skies and a gradual heating effect can be observed. In August, however, due to cloud cover, the air temperature was lower and fluctuated strongly after reaching a maximum at 10.00 am. Based on the air temperature data from both months it is assumed that the warming effect of solar radiation was not enough to overcome the cooling effect of groundwater in August, thus all metrics detect groundwater discharge in the A_W and A_N zones.

As decreased water velocity leads to less mixing of heat in lakes, a further challenge of DTS applications could be the entry of cold inlets or surface-runoff to the lake. During the August field campaign we noted seepage-face runoff from the riparian zone (groundwater), which might have affected the DTS results at the A_E zone. Although the daily minimum temperatures indicated a possible discharge zone here, the diel amplitude and daily standard deviation did not confirm these results. Moreover neither the seepage meter-based fluxes nor the temperature-based fluxes showed high values in the area. One reason can be the propagation of the negatively buoyant cold water on the SWI from the surface run-off to the lake.

These observations demonstrate the enhanced influence of external conditions on the DTS data recorded in lakes, also emphasizing the importance of using metrics with different temporal aspects to detect groundwater discharge. The daily standard deviation and diel amplitude both reflect daily temperature variations, including the
effects of solar radiation. On the other hand the daily minimum temperatures were measured during the night, free from solar input, thus considered more reliable in spring and summer conditions. Cold inlets entering the lake however, may greatly affect daily minimum temperatures and also slightly influence the daily standard deviation and diel amplitudes, thus a comprehensive interpretation of all metrics is necessary at all times. Next to the effect of solar radiation and cold inlets, future studies could possibly identify more factors influencing DTS applications in lakes.

Comparing data collected in different seasons (and years) enables the observation of short and long-term temporal and spatial changes in groundwater discharge. The DTS is able to identify a significant high discharge zone that is stable throughout the study period, i.e., F_W in February, A_W in August, and despite the disagreement of metrics possibly also M_W in May (Figure 6.5), which corresponds to the ice hole location HC/H4 observed both in 2010 and 2012 (Figure 6.11). This result agrees with the conclusion of the field experiments and modeling study of Kidmose et al. (2013) who found that discharge increases with distance from shore line. This is likely because groundwater mainly originates from the Western side of the lake and discharges to the lake as soon as the layer of organic sediments in the lake ends (Figure 6.2).

Spatial variations in groundwater discharge are also manifested in the appearance of new possible discharge zones in August (A_N) and most probably also in May (M_M). The location of the A_N zone matches the location of the HD ice hole found in 2010. The M_M zone could be related to the ice hole HB/H3 observed in February, with the assumption that by May this H3 discharge zone could have shifted slightly closer to the shoreline (approximately 0.5 m).

However, during the different seasons a temporal trend in the spatial distribution of groundwater discharge can also be observed. From winter to summer groundwater discharge locations shift slowly towards the lakeshore and are of smaller spatial extent (Figure 6.11). While in February rather the Southern part of the Western zone contributes to discharge, in May only the Northern part of the section (M_W) shows possible strong groundwater influence and by August the contributing zone (A_W) becomes spatially restricted. As the lake levels only changed a few centimeters in 2011/2012, one explanation of the temporal change in the distribution of discharge could be a decrease in the water table and thus the head-gradient on the Western side of the lake from where groundwater is mainly derived.

The high degree of spatial and temporal variability of discharge detected by the DTS was also confirmed by the traditional punctual measurements. The ice hole observations in 2010 and 2012 visually indicated the existence of a few scattered high discharge zones in the study section, implying that discharge was heterogeneous and temporally variable. The seepage meter and temperature-based fluxes in the same area from
August 2012 confirm the high degree of spatial variability shown both by the ice holes and the DTS campaigns, with fluxes ranging from very low to close to 0.6 md$^{-1}$ in isolated zones (Figures 6.9 and 6.10).

Even though both the seepage meter-based and temperature-based fluxes are associated with uncertainties in the measurements and in the calculation of vertical fluxes (installation, assumption of vertical flow), the two approaches give the same spatial mean in August 2012 (0.15-0.16 md$^{-1}$). Furthermore, the fluxes estimated from temperature profiles in the same area in July 2010 gave a similar mean flux of 0.14 md$^{-1}$ (but with a much lower variability than in August 2012). Compared to the temperature-based fluxes, seepage meter measurements in August 2012 gave a much wider range of fluxes (Table 6.1). Kidmose et al. (2013) found that seepage meter fluxes could vary in space by as much as a factor of 50. Our measurements suggest that the spatial variability in groundwater discharge is even higher (Table 6.1) which agrees with the results of Lee (1977) who found a range of fluxes between 8.64x10$^{-4}$ and 0.22 md$^{-1}$ around a lake and Rosenberry et al. (2010) also found fluxes between 0.019 and 1.37 md$^{-1}$ in undisturbed measurements along a lakeshore. The distribution of discharge as represented by Figures 6.9, 6.10b and 6.10c could therefore be even more heterogeneous.

These temperature- and seepage meter-based fluxes do not show the same high fluxes as measured in the ice holes in January 2010 and February 2012 (Table 6.3). Possible explanations could be; (i) the seepage meters and temperature probes were not exactly placed at the same location as the ice holes observed in February 2012; (ii) the zones with high discharge had moved between February and August or (iii) due to a decrease in head-gradient on the Western side of the lake, groundwater discharge also decreased. However, the temperature profile D2 measured directly in ice hole H3 in February (one of the ice holes present in both 2010 and 2012) gave a flux of 3.46 md$^{-1}$ very similar to the fluxes measured by seepage meters in January 2010.

In August 2012 the spatial distribution of discharge detected by DTS and the traditional punctual methods could also be compared. Only the seepage meter-based fluxes agreed with the DTS data at the location of the A_W discharge zone (Figure 6.10b), where in the proximity of the potential discharge zone detected by the DTS, fluxes up to 0.62 md$^{-1}$ were observed. Temperature-based measurements show fluxes of 0.3 md$^{-1}$ in the A_N area (Figure 6.10c), which also coincides with the ice hole HD detected in 2010. Similarly in February the DTS was only able to capture the H4 and H1 concentrated discharge zones, but not H2 and H3, although measurements from January 2010 and February 2012 confirmed fluxes up to 6.3 md$^{-1}$ at ice hole locations.

Thus the approximate location of all major discharge sites identified by the DTS was confirmed by seepage meter and temperature-based based fluxes, but the DTS did not detect all the major discharge sites and did not reflect the complex spatial pattern
that the traditional punctual measurements show. Contrary to our study, Lowry et al. (2007) could only confirm one out of two potential discharge sites detected by the DTS, where the groundwater spring was visible in the streambed sediments.

These discrepancies between the DTS, flux measurements and the ice holes as proxies for groundwater discharge can possibly be explained by either the magnitude of inflow that the DTS can detect or the spatial extent of the concentrated discharge sites. Based on the groundwater flux measurements of August the DTS was capable of detecting fluxes larger than approximately 0.3 md$^{-1}$. The spatial-averaging interval of the DTS installation used in this study was 1.01 m which can be compared to the size of the sand rings at the concentrated discharge sites (diameter of $\sim 0.05$ m). The spatial extent of the concentrated discharge sites is considerably smaller than the spatial-averaging interval. Thus, it appears that small, but concentrated discharge sites may not be visible in the SWI temperatures averaged over 1.01 m, unless the fiber optic cable is positioned directly on these sites and the contrast in temperatures is high. In stream applications, however, such concentrated discharge sites can be recorded with the gradual recovery in streambed temperatures detectable for several meters along a stream and cable. The spatial averaging of temperatures over 1.01 m distance in lakes may lead to even greater challenges in detecting concentrated discharge during the winter when positive buoyancy drives the warm groundwater away from the fiber optic cable.

Apart from a disturbance on the lakebed surface, the DTS method is still non-invasive compared to e.g. seepage meter measurements, thus it was also possible to preserve the original state of the field site. This way it was assured that the observed spatial variation in temperatures and indirectly in groundwater discharge is solely a result of changes in groundwater flow conditions and are not due to preferential flow paths created by the application of flux measurements methods modifying the deeper sediment structure. The installation of the cable is easy in lakes with gentle slopes, allows for more variation when deploying the fiber optic cable, and the effects of sedimentation and scouring are small. Thus, within two days an area of 150 m$^2$ (25 by 6 m) was mapped with a single fiber-optic cable. The DTS method could therefore be a promising option for long-term monitoring of major discharge sites.

### 6.6 Conclusions

Groundwater discharge zones were mapped at Lake Væng by measuring diurnal changes in temperatures at the Sediment-Water-Interface (SWI) using a looped DTS layout. Temperature data were evaluated by three different metrics; (1) the daily minimum
temperature, (2) the diel amplitude, and (3) the daily standard deviation in temperature. DTS was used in three seasons; February, May, and August of 2012 and compared with other types of measurements or proxies for discharge; seepage meter- and temperature-based fluxes, and ice thickness measurements. Although the approximate location of major discharge sites detected by the DTS was confirmed by seepage meter and temperature-based fluxes, the DTS does not reflect the spatial variability in discharge detected by seepage meter and vertical temperature profile measurements. For example, the DTS is not able to show the same areas where ice holes were observed, which may be explained by not having the cable precisely overlaying these areas of reduced size.

Nevertheless, the DTS was capable of detecting major discharge sites and showing not only a consistent pattern having highest discharge in the most off-shore western part of the study section, but also a shift of discharge areas towards the shoreline likely due to shifts in the head gradient. This is a result of the special groundwater-lake interaction, where groundwater originates primarily from the other side of the lake, flows underneath the lake, and discharges on the eastern side.

Based on the results of this study we observe the following; (i) DTS was able to capture spatial and temporal variations in temperatures at the SWI and could therefore be used in long-term installations to monitor groundwater discharge to lakes, (ii) DTS had difficulties detecting concentrated discharge zones of small size and spatial heterogeneity of fluxes, (iii) due to the quasi-motionless water body of the lakes the influence of external conditions (solar radiation, cold inlets) on SWI temperatures are enhanced, (iv) thus several metrics of different temporal aspect are necessary for a comprehensive interpretation of the SWI temperature data, (v) the method works best when the influence of solar heating is smallest, in the summer night measurements are required and the diel amplitude and daily standard deviation are not reliable metrics. The application of DTS with the layouts described in this work can represent a new methodology for studying lake-groundwater interactions in areas of greater spatial extent where temporal changes in discharge can occur.

**Acknowledgments**

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References


Chapter 7

Paper III

Application of Distributed Temperature Sensing for a coupled mapping of sedimentation processes and spatio-temporal variability of groundwater discharge in soft-bedded streams

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Abstract

The delineation of potential groundwater discharge areas based on Distributed Temperature Sensing (DTS) data of the streambed can be difficult in soft-bedded streams where sedimentation and scouring processes constantly change the position of the fiber optic cable relative to the streambed. Deposition-induced temperature anomalies thus resemble the signal of groundwater discharge and scouring will cause the cable to float in the water column and measure stream water temperatures. DTS applied in a looped layout with nine fiber optic cable rows in a 70 x 5 m section of a soft-bedded stream made it possible to detect variability in streambed temperatures between October 2011 and January 2012, while detailed, monthly streambed elevation surveys were carried out to monitor the position of the fiber optic cable relative to the streambed and to quantify the effect of sedimentation processes on streambed temperatures. Based on the simultaneous interpretation of streambed temperature and elevation data, a method is proposed to delineate potential high-discharge areas and identify deposition-induced temperature anomalies in soft-bedded streams. Potential high-discharge sites were detected using as metrics the daily minimum, maximum, and mean temperatures as well as the diel amplitude and daily standard deviation of temperatures. These potential high-discharge sites were mostly located near the channel banks also showing temporal variability in discharge which can be explained by the scouring and redistribution of streambed sediments, leading to the relocation of pool-riffle sequences. This study also shows that sediment deposits of 0.1 m thickness already resulted in an increase in daily minimum streambed temperatures and decrease in diel amplitude and daily standard deviation, while scouring sites showed lower daily minimum streambed temperatures and higher diel amplitude and daily standard deviation compared to areas without sedimentation and scouring.

7.1 Introduction

Due to its influence on stream ecology (Dahm et al., 1998) surface water-groundwater interactions have acquired more importance in environmental studies with special attention to groundwater discharge to streams. In spite of the wide variety of methods applied, it is still difficult to capture the spatial and temporal variability of groundwater
discharge. Among the methods applied to indirectly estimate groundwater discharge into streams, methods based on Darcy’s law, the mass balance approach and the use of heat as a tracer are the most widely used (Kalbus et al., 2006).

Heat as a natural tracer has been successfully applied for 50 years (Anderson, 2005). Making use of the temperature contrast between surface water and groundwater, it is possible to locate and delineate groundwater discharge zones. Vertical groundwater fluxes to the surface water bodies can be calculated based on an analytical solution of the steady heat transport equation (Bredehoeft and Papadopulos, 1965; Stallman, 1965; Schmidt et al., 2007; Jensen and Engesgaard, 2011) or based on time series analysis of the non-steady heat transport equation (Hatch et al., 2006; Keery et al., 2007).

Until the last decade the number of temperature measurements for using heat as a tracer of groundwater discharge was limited in space and time. As a new technique Distributed Temperature Sensing (DTS) is capable of measuring temperature continuously both in space and time along a fiber optic cable. Depending on the instrument specifications and the installations, temperature data can be averaged over each 0.25 meter with a precision of 0.01°C. The sensor of the method is a fiber optic cable which after deployment provides for quick and non-invasive temperature measurements. A detailed description about the theory can be found in Selker et al. (2006a) and Tyler et al. (2009).

Many studies used DTS to detect and quantify groundwater discharge to streams (Selker et al., 2006b; Lowry et al., 2007; Westhoff et al., 2007; Briggs et al., 2011; Karthikeyan et al., 2012; Krause et al., 2012) or to study the thermal variations of saltmarsh channels and lakes (Moffett et al., 2008; Sebok et al., 2013). The stream studies were mostly carried out in hard-bedded streams of low discharge. Westhoff et al. (2007) use temperature data recorded in a stream of an average discharge of 1.21 ls⁻¹ to calibrate a stream temperature model. Lowry et al. (2007) and Roth et al. (2010) carried out their studies in streams with discharges of 37-53 ls⁻¹ and 82.5 ls⁻¹, respectively. In these streams groundwater entering the stream forms plumes of anomalous temperature at the SWI and in the stream water. These anomalies are traceable along the cable for several meters usually followed by a gradual recovery of the original temperature as was observed in several studies (Selker et al., 2006a,b; Lowry et al., 2007; Westhoff et al., 2007; Briggs et al., 2011). Higher discharge streams, however, represent a special environment for DTS applications as groundwater plumes are immediately diluted and due to the lack of streambed temperature contrasts of several degrees, the delineation of groundwater discharge areas could present a problem. One single fiber optic cable may not be able to capture variability in streambed temperatures and thus groundwater discharge across the stream.

The fiber optic cable used in previous DTS studies was generally deployed on the
streambed at the Sediment-Water Interface (SWI). Only Lowry et al. (2007) installed the fiber optic cable just below the SWI and Krause et al. (2012) buried it 0.05 m deep in streambed sediments in order to avoid signal loss due to advective heat transport caused by river flow at the streambed surface. At the same time due to the soft streambed material and the dynamic environment, redistribution of streambed sediments is a naturally occurring phenomenon in soft-bedded streams. This may result in scouring of streambed material from under the fiber optic cable or the burial of the fiber optic cable. These processes occur simultaneously in a stream and change in space and time. This issue was not yet addressed in previous studies, where the fiber optic cable was installed in low-discharge, hard-bedded streams and fixed with rocks found in the channel. However, sedimentation processes in soft-bedded streams will modify streambed temperatures recorded by the DTS and thus also affect the data interpretation regarding groundwater discharge locations. The development of new monitoring methodologies and DTS data interpretation in soft-bedded streams is therefore necessary.

A DTS field study was carried out in a lowland, soft-bedded stream in Western Denmark based on a looped fiber optic cable layout and a new monitoring methodology using long-term DTS and monthly bed morphology surveys. The objectives were to; (i) quantify the effects of changing streambed morphology on DTS streambed temperature recordings, (ii) investigate the capability of DTS to monitor sedimentation and scouring processes and to (iii) delineate potential high-discharge zones based on the detected spatial variability in streambed temperatures.

### 7.2 Field site

The study site is located along Holtum stream (Figure 7.1b), a lowland stream in Jutland, Western Denmark (Figure 7.1a). This glacial floodplain valley is found in the upper part of Skjern River Basin with a catchment area of 70.4 km². The topmost sediment layer in the area is characterised by glacial sand and silt from the Weichsel glacial period (Houmark-Nielsen, 1989). The stream is groundwater-dominated with groundwater discharge fluxes varying along the streambed (Karan et al., 2013; Poulsen et al., in prep.). The flow direction of the stream at the 70 meter long and 3.5-5 m wide study section is approximately to the West with a slight bend at the middle of the section (Figure 7.1c). Stream water temperature ranges between 1 and 16°C during the year and between 3.3 and 9.8°C from October 2011 to January 2012; the study period. The mean annual discharge of the stream was 1068 ls⁻¹ in 2010 and 1267 ls⁻¹ in 2011, measured 2 km downstream from the study site at Hygild station (Figure 7.1b).
Between the study site and the discharge measurement station a tributary of a mean annual discharge of 280 l s\(^{-1}\) enters the stream. In the study section the average water depth in 2011 varied between 0.54 m in May and 0.71 m in December with maximum water depths of 0.95 m and 1.47 m, respectively. The stream has a soft streambed consisting mostly of medium and coarse-grained sand. The DTS survey was carried out after the annual weed cutting in August, thus emergent vegetation was scarcely present and did not influence the positioning of the cable.

Figure 7.1: Location of the catchment in Denmark (A), with the position of the study site within the catchment and the location of the precipitation measurements at Voulund and discharge measurements at Hygild sites (B). The setup of the looped DTS system, the location of the artificial obstruction (D) and the experimental section (Figure 7.7) are shown on panel C.

7.3 Methods

7.3.1 Distributed Temperature Sensing

Streambed temperatures were measured by a fiber optic cable (BruSteel multimode, Brugg Cables) using the Sensornet Distributed Temperature Sensing system (Oryx DTS-SR). The fiber optic cable was fixed to the streambed bottom on 12 October
In the double-ended measurements streambed temperatures were averaged for each 1.01 meter along the cable and integrated over 20 minutes. The installation was calibrated by having 30 meters of fiber optic cable for calibration both at the beginning and end of the installation (Figure 7.1c, precision of the measurements is shown in Table 7.1).

In previous studies the fiber optic cable was laid out longitudinally along the stream (Selker et al., 2006b; Westhoff et al., 2007; Lowry et al., 2007; Roth et al., 2010; Briggs et al., 2011) or with two cable loops (Krause et al., 2012). In the present study the cable was instead fixed to the streambed in a looped pattern by the aid of iron staples. In the 70 meters long study section the cable was positioned along the flow direction of the stream with wide cable bends at the end of each row (Figure 7.1c). Nine parallel cable rows were laid in this fashion by the aid of equidistant poles dividing the width of the stream into sections of equal distance of ∼ 0.4 m between the cable rows. Laying the cable in nine rows along the stream in a looped pattern made it possible to enhance the measurement grid across the width of the stream. Therefore, during each measurement 584 average temperature values were collected along 1.01 m cable sections. The streambed temperature measurements were later used to make interpolated temperature maps of the streambed.

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</tr>
<tr>
<td>13 December 00:00 am - 12:00 pm</td>
<td>0.04</td>
<td>12 December 2011</td>
</tr>
<tr>
<td>17 January 00:00 am - 12:00 pm</td>
<td>0.05</td>
<td>17 January 2012</td>
</tr>
</tbody>
</table>

Table 7.1: Precision of the 24 hour DTS datasets analyzed and the corresponding streambed elevation surveys.

### 7.3.2 Streambed elevation surveys

Monthly streambed elevation surveys were carried out by the aid of a Trimble R8 5800 GPS with a mean horizontal precision of 1 cm ± 0.3 cm (maximum of 3.7 cm) and a mean vertical precision of 1.4 cm ± 0.4 cm (maximum of 5.7 cm). The first survey was made after fixing the cable to the streambed on 13 October. The same survey was repeated on 13 November, 12 December 2011, and as a follow-up on 17 January 2012. Streambed elevation measurements were taken at each meter along the stream and five times across the 3.5-5 m wide stream. In case of sudden elevation changes more measurements were necessary. The relative elevation change compared to the
initial elevation recorded on 13 October was then calculated for each data point and interpolated contour maps were used to visualize the effects of deposition and scouring on the recorded streambed temperatures. As it was possible to study groundwater discharge after a significant rain event on 19 October (Figure 7.2), the 24 hour dataset from 20 October was also analyzed for groundwater discharge with the methods presented below, although no simultaneous streambed elevation survey is available. A later survey of 13 November represented streambed morphology on 20 October with the assumption that streambed morphology did not change during this low-flow period (Figure 7.2).

### 7.3.3 Precipitation and discharge measurements

Stream discharge data (Figure 7.2) was collected at Hygild station (Figure 7.1b) 3 km downstream from the study area with two Acoustic Velocity Meter (AVD, Teledyne RD Instruments) located at different levels above the streambed. Precipitation data is available from Voulund station (Figure 7.1b) 7 km in the Northwestern direction from the field site. Air temperature was also measured at the field site by PT100 probe fixed to a climate station.

![Figure 7.2: Precipitation and air temperature at the study site and stream discharge measured 2 km downstream of the study site during the study period. Dates on the figure refer to the days when the discharge peaks were registered. The black arrows indicate the days when DTS data was analyzed.](image)

Figure 7.2: Precipitation and air temperature at the study site and stream discharge measured 2 km downstream of the study site during the study period. Dates on the figure refer to the days when the discharge peaks were registered. The black arrows indicate the days when DTS data was analyzed.
7.3.4 Characterising sedimentation processes

Effect of sedimentation processes on streambed temperatures

In order to quantify how sedimentation processes affect streambed temperatures, DTS data from 14 November and 13 December 2011 were divided into groups based on the streambed elevation changes relative to the time of deployment of the cable. Areas with more than 0.05 m sediment deposition were classified as sedimentation sites. Areas where more than 0.05 m sediment was eroded were classified as scouring sites. Sites with less than ± 0.05 m change in elevation were classified as stable sites. This arbitrary threshold of ± 0.05 m was determined considering the vertical precision of the GPS measurements and the soft sandy streambed material which could be slightly compressed by the measurement device. Detailed classification of the groups and the number of data points in each group are shown in Table 7.2.

<table>
<thead>
<tr>
<th>Group</th>
<th>Elevation change</th>
<th>14 November 2011</th>
<th>13 December 2011</th>
<th>17 January 2012</th>
</tr>
</thead>
<tbody>
<tr>
<td>Potential discharge sites a)</td>
<td>Dataset</td>
<td>9 (1.5%)</td>
<td>18 (3%)</td>
<td>584</td>
</tr>
<tr>
<td>Sedimentation</td>
<td>&lt; 0.6 m</td>
<td>584</td>
<td>584</td>
<td>584</td>
</tr>
<tr>
<td></td>
<td>0.4-0.6 m</td>
<td>19 (3.2%)</td>
<td>7 (1.3%)</td>
<td>10 (1.7%)</td>
</tr>
<tr>
<td></td>
<td>0.3-0.4 m</td>
<td>7 (1.3%)</td>
<td>10 (1.7%)</td>
<td>14 (2.4%)</td>
</tr>
<tr>
<td></td>
<td>0.2-0.3 m</td>
<td>28 (4.8%)</td>
<td>14 (2.4%)</td>
<td>16 (2.7%)</td>
</tr>
<tr>
<td></td>
<td>0.1-0.2 m</td>
<td>28 (4.8%)</td>
<td>14 (2.4%)</td>
<td>16 (2.7%)</td>
</tr>
<tr>
<td></td>
<td>0.05-0.1 m</td>
<td>79 (13.5%)</td>
<td>82 (14%)</td>
<td>40 (6.8%)</td>
</tr>
<tr>
<td>Stable a)</td>
<td>± 0.05 m elevation change</td>
<td>305 (52.2%)</td>
<td>309 (52.9%)</td>
<td>299 (51.2%)</td>
</tr>
<tr>
<td>Scouring</td>
<td>0.05-0.1 m</td>
<td>66 (11.3%)</td>
<td>90 (15.4%)</td>
<td>120 (20.5%)</td>
</tr>
<tr>
<td></td>
<td>0.1-0.2 m</td>
<td>59 (10.1%)</td>
<td>36 (6.2%)</td>
<td>52 (8.9%)</td>
</tr>
<tr>
<td></td>
<td>0.2-0.3 m</td>
<td>33 (5.6%)</td>
<td>10 (1.7%)</td>
<td>14 (2.4%)</td>
</tr>
<tr>
<td></td>
<td>0.3-0.4 m</td>
<td>14 (2.4%)</td>
<td>8 (1.3%)</td>
<td>2 (0.3%)</td>
</tr>
<tr>
<td></td>
<td>0.4-0.6 m</td>
<td>13 (2.2%)</td>
<td>14 (2.4%)</td>
<td>2 (0.3%)</td>
</tr>
<tr>
<td></td>
<td>0.6-0.8 m</td>
<td>8 (1.3%)</td>
<td>14 (2.4%)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>&gt; 0.8 m</td>
<td>8 (1.3%)</td>
<td>14 (2.4%)</td>
<td></td>
</tr>
</tbody>
</table>

a) Potential high-discharge sites are also included in the group with ± 0.05 m sedimentation
b) In the 17 January dataset no potential groundwater inflow sites were detected due to extreme changes in streambed elevation and consequently the increased uncertainty of the location of the cable. Data presented here reflect elevation changes between the 13 December 2011 and 17 January 2012 survey

Table 7.2: Sedimentation groups defined on the basis of streambed elevation changes with the number of data in each group

The effect of sedimentation processes on streambed temperatures was analyzed by comparing the daily minimum, maximum and mean temperatures, diel amplitude (difference of daily maximum and minimum temperature) and daily standard deviation of the different sedimentation groups. As the temperature conditions for each mea-
measurement period were slightly different and the variability of metrics was therefore large, the values of the individual groups were normalized. This way the variations both within the metrics, the sedimentation groups and different measurement periods became directly comparable. The normalization of values was carried out as follows:

$$\hat{X}_i = \frac{X_i - X_{\text{stable}}}{\sigma}$$  \hfill (7.1)

where $\hat{X}_i$ is the normalized value of the variable in group $i$, $X_i$ is the arithmetic mean of the variable in group $i$, $X_{\text{stable}}$ is the arithmetic mean of the variable in the stable group with less than $\pm 0.05$ m change in elevation, and $\sigma$ is the standard deviation of the variable of the whole dataset.

To study the effects of scouring and sedimentation on long-term streambed temperature records, on 12 October 2011 a steel piezometer pipe of 0.03 m diameter and 3 m length, from now on referred to as artificial obstruction, was installed in the thalweg 1.5 m deep in the streambed (Figure 7.1d). The effects of scouring on DTS temperature records were studied on the 13 December dataset when the scouring created by the pipe due to the current was maximal. The obstruction was removed on 14 December 2011.

The effect of sedimentation on streambed temperatures was then studied after the removal of the artificial obstruction, using the streambed dataset recorded on 17 January when the scour hole was already filled up by freshly deposited sediments. In the analysis, the same sedimentation groups and normalized metrics will be used as described above. Due to the extreme changes in streambed elevation, and thus the increased uncertainty of the location of the cable relative to the streambed, the 17 January dataset was not analyzed to detect potential discharge sites. The temperature characteristics of groundwater discharge sites will be compared to the sedimentation groups by plotting the normalized metrics of the discharge sites detected on 14 November and 13 December 2011 together with the normalized metrics of the sedimentation groups of 17 January 2012.

### Dynamic streambed processes monitored by DTS

The monthly streambed elevation surveys only provide a snapshot of the streambed morphology formed by the dynamic sedimentation/scouring processes. Long streambed temperature time series from 13 November until 31 December 2011 were therefore used to monitor sedimentation processes and to directly relate changes in streambed morphology to hydrometeorological or artificial events. Sediment migration speed along
the stream was also calculated by dividing the distance the temperature anomaly moved by the time elapsed. Similarly it was also possible to quantify the spatial extension of the freshly deposited sediments in the scour hole created during the sedimentation experiment.

### 7.3.5 Identification of potential groundwater discharge sites

Deeper groundwater in Denmark has an average temperature of around 8 °C. Groundwater discharge to streams is thus manifested by cold temperature anomalies during summer when the stream temperatures are higher (9-16 °C) and as warm temperature anomalies during winter when stream temperatures are lower (1-7 °C). In the range of stream discharge at the field site, the groundwater heat flux is quickly dissipated and streambed temperature contrasts are expected to be rather low. It may therefore be difficult to identify groundwater discharge sites based on the streambed temperature time series only. The streambed temperature time series recorded by the DTS were therefore analyzed by the aid of basic metrics described below.

Assuming that discharging groundwater has a constant temperature during a measurement campaign, potential discharge sites can possibly be identified using metrics such as the daily minimum, maximum, mean streambed temperatures, the diel amplitude (calculated as the difference between the daily maximum and minimum temperatures), and the daily standard deviation of temperature. To qualify as a potential groundwater discharge site, three criteria were used. First, all basic metrics at the high-discharge sites, such as the daily minimum, maximum and mean streambed temperature, must exceed the spatial mean of the whole dataset by two times the standard deviation of the whole dataset as in winter these sites will be closer to the average groundwater temperature of 8 °C. Secondly, both the daily standard deviation and diel amplitude at the high-discharge site must be lower than the mean of the corresponding metrics of the dataset by 2 standard deviations as groundwater has a more stable temperature than surface waters. Groundwater discharge thus dampens diurnal temperature oscillations reducing the diel amplitude and the daily standard deviation at high-discharge sites. The arbitrary cutoff limit of two standard deviations was chosen to give a limited number of potential high-discharge sites.

High stream discharge related to rainfall or anthropogenic events such as river restoration or channel modifications are likely to mobilize and redistribute sediments in soft-bedded streams. Sediment deposition on the fiber optic cable isolates the cable from the stream resulting in dampened diurnal oscillations. Depending on the depth of burial, the isolation of the cable can also create a temperature anomaly resembling a groundwater signal. Thus as a third criteria, streambed elevation changes at the
potential high-discharge locations could not exceed ± 0.05 m compared to the initial elevation of 13 October 2011 to exclude the effects of sedimentation and scouring in the data interpretation.

As the temperature signal of groundwater and sedimentation-induced warm temperature anomalies are very similar, their discrimination presents several difficulties. While the metrics used in this study are sufficient to delineate potential high-discharge sites within the stream, it is not possible to reliably predict if the warm temperature anomaly at sedimentation sites is solely a result of sediment deposition or the coupled effect of sediment deposition and significant groundwater discharge at the same location. Similarly groundwater discharge can also occur at scouring sites. Sites with more than ± 0.05 m change in elevation will not be considered as potential high-discharge sites in this study, although groundwater discharge is possible at these locations.

Being a groundwater-fed stream, groundwater discharge is likely to occur everywhere within the study section (Poulsen et al., in prep.). The detected potential discharge sites therefore likely represent discharge sites with high groundwater flux. Groundwater discharge patterns will be studied in this way based on complete 24 hour datasets from 13 October 13:00 until 14 October 13:00 and from 00:00 am until 12:00 pm on 20 October, 14 November and 13 December 2011 (Table 7.1).

### 7.4 Results

#### 7.4.1 Effect of sedimentation processes on streambed temperatures

The simultaneous analysis of streambed temperatures and elevation data indicates that areas with sediment deposition display higher normalized daily minimum, maximum and mean temperatures and lower diel amplitude and daily standard deviation than the stable group (Figure 7.3 and Table 7.3), except for the daily maximum temperature values in the December dataset (Figure 7.4). The daily minimum, maximum and mean temperatures show an increase with increasing deposition thickness.

In the 14 November dataset, areas with scouring do not present consistent changes in metrics when related to the depth of scouring. Areas with 0.3-0.4 m of scouring, however, display similar behaviour as the sedimentation group with higher normalized daily minimum, maximum and mean temperatures and lower normalized diel amplitudes and standard deviations than the stable group (Figure 7.3). On the other hand, the 13 December dataset shows consistent changes in the metrics when related to the
depth of scouring (Figure 7.4). Following a gradual decrease in normalized daily minimum, mean and maximum daily temperatures and reaching the minimum values in the group with 0.3-0.4 m scouring, a gradual increase can be observed. The normalized diel amplitudes and daily standard deviations show a similar but reverse trend.

![Graph showing temperature and diel amplitude for different sedimentation groups on 14 November 2011](image)

Figure 7.3: Daily minimum, maximum, mean temperatures, daily standard deviation and diel amplitude calculated for different sedimentation groups on 14 November 2011 and normalized to the stable group with less than ±0.05 m change in streambed elevation. Positive results indicate higher values, negative results lower values than the stable group.

After the removal of the obstruction from the stream, the 17 January streambed elevation survey showed that the scour hole created by the obstruction was filled up with sediments of up to 0.82 m thickness. Sites with sediment deposition display higher normalized minimum, mean and maximum temperatures and lower normalized diel amplitudes and daily standard deviation than the stable group (Figure 7.5). An increase in normalized temperatures and decrease in normalized diel amplitudes and daily standard deviations can be observed in the group with 0.05-0.1 m and 0.1-0.2 m deposition. The normalized temperatures do not increase further with increasing sediment thickness above 0.2 m (Figure 7.5). On 17 January sites with more than 0.1 m sedimentation showed minimum, maximum and mean streambed temperatures 18, 13, and 15% higher respectively than the stable group, and both diel amplitude and daily standard deviation were considerably lower, 75% than that of the stable group (Table 7.3).

Based on the three criteria described earlier, several potential groundwater discharge sites were delineated both on 14 November and on 13 December 2011 (see Section 7.4.3). These potential high-discharge locations display higher normalized daily
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<table>
<thead>
<tr>
<th>Group</th>
<th>Daily minimum</th>
<th>Daily maximum</th>
<th>Daily mean</th>
<th>Diel amplitude</th>
<th>Daily standard deviation</th>
</tr>
</thead>
<tbody>
<tr>
<td>14 Nov. Stable</td>
<td>6.12</td>
<td>6.73</td>
<td>6.50</td>
<td>0.61</td>
<td>0.17</td>
</tr>
<tr>
<td>Discharge</td>
<td>6.67 (+9%)</td>
<td>7.15 (+6%)</td>
<td>6.97 (+7%)</td>
<td>0.44 (-18%)</td>
<td>0.10 (-39%)</td>
</tr>
<tr>
<td>Sedimentation</td>
<td>6.27 (+2%)</td>
<td>6.81 (+1%)</td>
<td>6.60 (+1%)</td>
<td>0.53 (-11%)</td>
<td>0.14 (-12%)</td>
</tr>
<tr>
<td>Scouring</td>
<td>6.13 (0%)</td>
<td>6.75 (0%)</td>
<td>6.52 (0%)</td>
<td>0.62 (0%)</td>
<td>0.17 (0%)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Group</th>
<th>Daily minimum</th>
<th>Daily maximum</th>
<th>Daily mean</th>
<th>Diel amplitude</th>
<th>Daily standard deviation</th>
</tr>
</thead>
<tbody>
<tr>
<td>13 Dec. Stable</td>
<td>5.10</td>
<td>6.03</td>
<td>5.55</td>
<td>0.92</td>
<td>0.26</td>
</tr>
<tr>
<td>Sedimentation</td>
<td>6.17 (+20%)</td>
<td>6.83 (+13%)</td>
<td>6.48 (+16%)</td>
<td>0.46 (-51%)</td>
<td>0.12 (-53%)</td>
</tr>
<tr>
<td>Scouring</td>
<td>5.22 (+2%)</td>
<td>6.02 (-1%)</td>
<td>5.61 (+1%)</td>
<td>0.80 (-14%)</td>
<td>0.23 (-14%)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Group</th>
<th>Daily minimum</th>
<th>Daily maximum</th>
<th>Daily mean</th>
<th>Diel amplitude</th>
<th>Daily standard deviation</th>
</tr>
</thead>
<tbody>
<tr>
<td>17 Jan. Stable</td>
<td>5.14</td>
<td>5.91</td>
<td>5.56</td>
<td>0.77</td>
<td>0.19</td>
</tr>
<tr>
<td>Sedimentation</td>
<td>6.11 (+18%)</td>
<td>6.69 (+13%)</td>
<td>6.48 (+15%)</td>
<td>0.58 (-25%)</td>
<td>0.14 (-25%)</td>
</tr>
<tr>
<td>Scouring</td>
<td>5.18 (0%)</td>
<td>5.98 (+1%)</td>
<td>5.62 (+1%)</td>
<td>0.79 (+3%)</td>
<td>0.20 (+4%)</td>
</tr>
</tbody>
</table>

Table 7.3: Effect of sedimentation on the daily minimum, maximum, mean temperatures, diel amplitude and daily standard deviation in the 14 November, 13 December 2011 and 17 January 2012 dataset. Values are given in °C, the percentages show the values relative to the stable group.

- a) Streambed elevation change less than 0.05 m
- b) More than 0.1 m sediment deposition
- c) More than 0.1 m scouring

![13 December 2011](image)

Figure 7.4: Daily minimum, maximum, mean temperatures, daily standard deviation and diel amplitude calculated for different sedimentation groups on 13 December 2011 and normalized to the stable group with less than ±0.05 m change in streambed elevation. Positive results indicate higher values, negative results lower values than the stable group.
minimum, mean and maximum temperatures and lower normalized diel amplitudes and daily standard deviations than any other groups for each survey period (Table 7.3). The differences between the potential discharge locations and the other groups are greater in the December than in the November survey (Figures 7.3 and 7.4). Nevertheless, the diel amplitude and daily standard deviation values of the potential high-discharge group show less deviation from the other groups than that of the normalized daily minimum, mean and maximum temperatures.

Figure 7.5: Daily minimum, maximum, mean temperatures and daily standard deviation and diel amplitude calculated for different sedimentation groups on 17 January 2012. Elevation differences between December and January are used to separate the categories. Groundwater discharge zones were not detected on 17 January as the extreme changes in streambed elevation prevented the localization of the relative position of the cable compared to the initial 13 October 2011 survey. Normalized values of the potential groundwater discharge sites are thus given on the basis of the 14 November and 13 December 2011 data, where the data was compared to the relevant metrics in the stable group at that time.

7.4.2 Sedimentation processes directly monitored by DTS

Sedimentation experiment

The sedimentation process will be examined along one cable row (Figure 7.6) located at the depositional inner bend of the stream where, due to the soft sediments, the highest scouring rates were observed. After the removal of the obstruction on 14 December the first warm temperature anomalies caused by sedimentation appear on 16 December (Figure 7.6a). Based on this anomaly, the longitudinal extension, the rate of movement
and the slope of the sediment deposit was determined with DTS data. It was found that by 22 December sediments were deposited over a reach stretching 5 m downstream the removed obstruction (Figure 7.6a) indicating a longitudinal growth rate of 0.75 m d\(^{-1}\). An abrupt growth from 5 meters to 8 meters in the longitudinal extent of the deposition can be observed between 22 and 24 December following a high discharge event on 24 December (Figure 7.2). The longitudinal growth rate of freshly deposited sediments was 1.5 m d\(^{-1}\) within these two days. After this event the depositional area maintains its size and seems to have reached equilibrium.

Figure 7.6: Sedimentation induced temperature anomaly observed after the removal of the steel pipe on 14 December 2011 along one row of fiber optic cable in the study section (A). The x axis shows the distance downstream along the cable (note that the flow direction is from left to right) and the y axis shows time elapsed. The location of this section is shown in Figure 7.8a. Stream and air temperatures between 14 and 31 December 2011 (B).

The asymmetrical pattern of the temperature anomaly (Figure 7.6a) conforms to the flow direction as the flow will form a steep slope at the scour hole upstream from the obstruction and a less steep morphology downstream from the obstruction. This new depositional environment may also have affected other areas in the stream. For instance, an area with warm temperature anomaly at 50 meters was reduced in its extension possibly due to scouring (Figure 7.6a).

The effect of sediment deposition on streambed temperature is also reflected by all metrics calculated in a 10 m long section along the scour hole (Figure 7.1c). When the obstruction is removed, the cable is quickly buried under freshly deposited sediments 0.5 and 1.5 m downstream from the location of the removed obstruction as illustrated by the gradually increasing minimum streambed temperatures (Figure 7.7), forming an anomaly similar to that induced by groundwater discharge. As sedimentation extends
in the downstream direction from 21 December, the highest temperatures are observed 2.5 meters downstream from the location of the removed obstruction. Accordingly, the highest recorded streambed temperatures on 17 January are observed just downstream the location of the sedimentation experiment where the cable got buried under sediments (Figure 7.8a,b).

Figure 7.7: Daily minimum temperatures recorded in the experimental section after the removal of the artificial obstruction on 13 December. The location of the obstruction and the section presented is shown on Figures 7.1 and 7.11.

Long-term dynamic streambed processes

From 14 November until 3 December the dataset represents conditions under homogeneous stream discharge (Figure 7.2). During this period one significant shift in the location of a temperature anomaly (Figure 7.8a) can be observed. The temperature anomaly starts to migrate in the downstream direction from its initial position, represented by the black dashed line (Figure 7.8a), after the 12.6 mm rain event occurring on 25 November and followed by a discharge event on 26 November (Figure 7.2). The end of the migration can also be attributed to a rain event on 4 December which induces yet another redistribution of already deposited sediments. Based on the time elapsed between these two rain events and the downstream distance covered by the anomaly, a high downstream migration speed of 0.7 m d$^{-1}$ can be calculated.

From 4 December 2011 the temperature recordings are considerably influenced by dynamic sedimentation/scouring processes. The first rain event on 4 December with 14.8 mm precipitation was accompanied by a 3.3 °C reduction in air temperature (Figure 7.2) corresponding also to a reduction of 2 °C in streambed temperature within a day (Figure 7.8a). The rain event on 9 December with 18.6 mm precipitation leads to a significant increase in stream discharge and is present on the temperature readings
as a sudden disappearance or weakening in the warm temperature anomalies caused by previous deposits (Figure 7.8a). Warm temperature anomalies appear again later at approximately the same locations but with a smaller spatial extension and with less temperature differences from the surrounding area. A decrease in the extension of the warm temperature anomalies indicated the effects of scouring caused by the 4-5 January rain event of 18 and 24 mm respectively (Figure 7.8a).

Figure 7.8: Streambed temperatures recorded from 13 November to 31 December 2011 (y axis) along the fiber optic cable (A) and the corresponding streambed elevation from 14 December 2011 (B). The black rectangle represents the cable section along the study section where the effects of the sedimentation experiment on temperatures recorded by the DTS was analysed (Figure 7.6). The red arrows indicate the location of the artificial obstruction at each cable row.

7.4.3 Delineation of potential groundwater discharge sites

The deployment of the cable on the SWI was followed by a 24h measurement period on 13 October. Based on the previously described metrics potential high-discharge sites are located in the middle of the study section close to the channel bend (daily minimum temperatures shown on Figure 7.9a and metrics given in Table 7.4). Despite low-flow conditions (Figure 7.2) at the end of the 24 hour survey, the cable was already slightly
buried under sediments near the inner bend, where the highest channel elevation was observed (Figure 7.9b). Based on this visual observation this location was excluded from among the potential high-discharge sites (Figure 7.10), although the temperature metrics fulfilled the criteria of a potential high-discharge site.

Figure 7.9: Daily minimum streambed temperatures on 14 October (10:30 am) (A), streambed elevation measured on 13 October 2011 (B), and daily minimum streambed temperatures on 20 October (07:00 pm) (C). The solid black lines on the streambed temperature maps and the streambed elevation map represent the spatial mean of the minimum streambed temperature dataset plus 2 standard deviations of the dataset.

Based on the 20 October dataset only one distinct warmer temperature zone could be identified (Figure 7.9c) further downstream of a similar zone near the outer bend on 13 October (Figure 7.10). Although the 13 November streambed elevation data was used for the interpretation of the 20 October dataset, this warm zone is considered to be a potential discharge zone as it is located in the erosional outer bend of the stream where sediment deposition on the cable is not likely.

On 14 November potential warm streambed zones can be identified at several scattered locations (Figure 7.11a). However, streambed elevation changes (Figure 7.11b) show that some of these warm anomalies are located in zones with sediment deposition, thus next to groundwater discharge, sedimentation is a more likely explanation for the warm temperature anomaly. Discarding these zones, a potential high-discharge site is located even further downstream a similar zone on 20 October and at several scattered locations upstream the channel bend (Figure 7.10).
### Table 7.4: Daily mean, minimum, maximum temperature and diel amplitude and daily standard deviation of all the data and potential discharge locations from the 13 October 2011 streambed temperature dataset. Values are given in °C.

<table>
<thead>
<tr>
<th></th>
<th>Daily mean</th>
<th>Daily minimum</th>
<th>Daily maximum</th>
<th>Diel amplitude</th>
<th>Daily standard deviation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dataset mean</td>
<td>7.17</td>
<td>5.97</td>
<td>8.67</td>
<td>2.69</td>
<td>0.89</td>
</tr>
<tr>
<td>Standard deviation</td>
<td>0.04</td>
<td>0.07</td>
<td>0.03</td>
<td>0.09</td>
<td>0.03</td>
</tr>
<tr>
<td>Mean of potential discharge locations</td>
<td>7.31</td>
<td>6.21</td>
<td>8.59</td>
<td>2.40</td>
<td>0.78</td>
</tr>
</tbody>
</table>

On 13 December the daily minimum streambed temperature is the highest further downstream of the potential discharge site near the left bank detected on 14 November (Figure 7.11c). This area is partially exposed to streambed elevation changes above ± 0.05m (Figure 7.11d), but potential high-discharge sites follow the previously observed location shift in the downstream direction (Figure 7.10).

![Figure 7.10: Location of potential high-discharge zones as a result of the simultaneous interpretation of DTS temperature and streambed morphology data.](image)

To illustrate the effect of sediment deposition on streambed temperatures, daily minimum temperatures recorded on 17 January are shown on Figure 7.11e. Comparing it with the streambed elevation changes between 14 December and 17 January 2011 (Figure 7.11f), warm temperature anomalies show an almost perfect coincidence with deposition sites.

### 7.5 Discussion

Heat inflow with groundwater to a high-discharge and fairly deep stream is diluted without causing a change in temperature traceable at the SWI for longer distances along a stream. Moreover, in soft-bedded streams the position of the instruments relative
Figure 7.11: Daily minimum streambed temperatures on 14 November (07:00 pm) (A), 13 December 2011 (07:00 pm) (C) and 17 January 2012 (06:00 pm) (E) given in °C with the dataset average contoured with green. Changes in streambed morphology between 13 October and 14 November 2011 (B), between 13 October and 13 December 2011 (D) and elevation changes between 13 December 2011 and 17 January 2012 (F). Elevation changes are given in meters with the positive values representing sedimentation and the negative values scouring. The solid black lines on the streambed temperature plots and streambed elevation change plots represent the spatial mean of the minimum streambed temperature dataset plus 2 standard deviations of the dataset.
to the streambed may change due to sedimentation and scouring, thus the measured temperatures may be indicative of different depth conditions and therefore not comparable. The looped DTS layout used in this study, however, proved to be an efficient way for monitoring spatial variability of streambed temperatures, with detected differences of up to 1.5 °C in this soft-bedded stream (mean daily discharge 1267 ls−1 in 2011), although these were not traceable in great distances in the downstream direction. In order to account for the sedimentation processes, streambed elevation changes were also monitored during the measurement period to avoid e.g. misinterpreting the deposition of stream sediments on the cable as areas of significant groundwater discharge.

Based on the simultaneous analysis of streambed elevation and temperature data, this study showed that the dynamic streambed processes influence the temperature measurements taken by DTS. The co-location of warm temperature anomalies and depositional areas is clearest at the location of the sedimentation experiment (Figure 7.8a also visible on Figure 7.11e, f). The same pattern, however, can also be observed in the dataset of 14 November 2011, closer to natural conditions where some warm temperature anomalies are depositional areas. Here, some areas of higher streambed temperature partially correspond to locations with more than 0.05 m deposition (Figure 7.11a, b), thus reflecting the difficulties of interpreting DTS data in soft-bedded streams.

Our results confirm that even a thin sediment layer of 0.05-0.1 m can cause streambed temperature anomalies with a deviation from the mean mimicking the characteristics of a groundwater signal (Figure 7.3, 7.4 and 7.5). This can be explained by the burial of the fiber optic cable, which isolated from the diurnal signal of the stream, records dampened streambed temperatures leading to lower diel amplitudes and daily standard deviations. Such anomalies were already apparent within 24 hours after the deployment of the cable where, on 13 October, elevated streambed temperatures were observed at the depositional inner bend of the stream (Figure 7.9a) emphasizing the need for careful interpretation of SWI data and streambed morphology in soft-bedded streams. For other streams this period may be shorter or longer as the speed of the burial process strongly depends on stream discharge and streambed sediment characteristics.

This study shows that the burial depth of 0.05 m used by Krause et al. (2012) to avoid signal loss does not alter the actual streambed temperatures significantly. Sediment thicknesses of more than 0.1 m resulted in 2%, 2%, and 18% increase in daily minimum temperatures; 11%, 14%, and 25 % decrease in diel amplitudes; and 12%, 14%, and 25% decrease in daily standard deviation in the November, December and January datasets, respectively, compared to the stable sedimentation group (Table 7.3). These results indicate that the diel amplitude and daily standard deviation are more sensitive to changes in streambed morphology. Sebok et al. (2013) also found that diel
amplitudes and daily standard deviations of temperatures at a lakebed as metrics for groundwater discharge are more sensitive to external conditions than daily minimum or maximum temperatures. However, with the exception of the extreme sedimentation conditions of the January dataset, the November and December temperatures recorded at the sedimentation sites show less differences from the dataset average than expected. It is assumed that during the summer months, when the contrast between the deep groundwater (8 °C) and the surface water temperature (up to 16 °C) is greater, this influence is even stronger than shown in this study with a maximal temperature contrast of 2.6 °C.

Scouring is not as apparent on temperature readings as the deposition-induced anomalies. These sites also show a deviation from the mean streambed temperature, but a consistent trend was only observed in the December dataset when scouring rates were at the maximum (Figure 7.4). It is assumed that at scouring sites the fiber optic cable either follows the changes of the streambed or floats reflecting the temperature conditions at the bottom of the water column. Thus, it is consequently more influenced by the diurnal temperature variation of the stream water. Hence, the elevated diel amplitude and daily standard deviation were 14% and 3%; and 13% and 4% above the mean values of the stable sites in December and January, respectively (Table 7.3).

The nine fiber optic cable rows also showed that streambed temperatures across the stream are far from uniform. Groundwater discharge areas in high-discharge streams, with small temperature contrast between surface water and groundwater, can therefore easily be overlooked when using only one or two cable rows, which were sufficient in previous studies of low-discharge streams (Westhoff et al., 2007; Lowry et al., 2007). Based on the streambed temperature data and its implications on groundwater discharge we can conclude that discharge in this small 70 x 5 m section is not uniform, neither in space nor in time.

Streambed temperature datasets showed that the location of the potential high groundwater discharge zones changed over time. On 13 October high discharge zones were identified downstream the depositional inner bend at the middle of the section and at the outer bend (Figure 7.10). By 20 October high discharge is concentrated at the right bank of the stream at the outer bend (Figure 7.10). Streambed temperature data from 14 November and 13 December shows potential high discharge even further downstream. Most of these potential discharge sites are located downstream of the artificial obstruction (Figure 7.10) and also show a tendency of downstream migration as scouring around the obstruction progresses (Figure 7.11b, d). This temporal variability in potential high discharge sites may be attributed to changes in head gradients and in streambed morphology.

Changes in streambed morphology can affect groundwater discharge not only by
the scouring of the low-permeability sediment layers but also by the redistribution of sediments and thus the relocation of pool-riffle sequences. During low-flow periods, fine-grained streambed material can be deposited on the streambed surface and thus reduce streambed hydraulic conductivity. High-discharge events remove this fine veneer, thus increasing hydraulic conductivity at the SWI enhancing groundwater discharge. In streams however, where the streambed is mobilized on a regular basis, like in the soft-bedded Holtum stream, this is not a likely explanation as individual events tend to constantly flush away the low-conductivity veneer (Rosenberry and Pitlick, 2009). In our study the observed shift in the location of potential discharge sites is therefore not related to the removal of the upper low-conductivity sediment layer during high stream discharge events.

The observed speed of change in the location of discharge sites between the survey periods is approximately 0.6-0.8 md$^{-1}$, calculated on the basis of the distance the warm temperature anomaly covered along the cable and the time elapsed. This agrees well with the observed migration speed of 0.7 md$^{-1}$ of the sedimentation-induced temperature anomaly between 25 November and 3 December (Figure 7.8a). As these velocities are similar and the sedimentation experiment mobilized a large amount of streambed sediments (Figure 7.8b) a possible explanation of the shift in the location of the discharge sites may be the change in streambed morphology.

It is assumed that sediments eroded from the inner bend and deposited further downstream change the location of small pools thus inducing upwelling following the redistribution of sediments. Potential discharge locations on 13 October are found downstream from the depositional inner bend in the deeper areas (Figure 7.10) suggesting upwelling due to a change in the slope of the streambed at the tail of the bedform agreeing with experimental results (Evans and Petts, 1997; Hannah et al., 2009) and modelling studies (Harvey and Bencala, 1993; Storey et al., 2003). The installation of the artificial obstruction leads to scouring at the inner bend and sediment deposition further downstream by 14 November (Figure 7.11b). Likewise, results for 13 December indicate upwelling induced by similar streambed morphology changes further downstream (Figure 7.11d). Thus, temporal variability in the location of discharge between 13 October, 14 November, and 13 December is most likely related to changes in streambed morphology rather than to the removal of a low-conductivity surface layer.

Large rain events can cause changes in hydraulic heads, thus also in the distribution of groundwater discharge. The significant rain event with 25 mm of rain on 19 October resulted in a large increase in stream discharge (Figure 7.2). Visual observations indicate, that during and after rain events in this period, shallow groundwater springs appear on the right streambank next to the discharge site detected on 20 October. Due
to the close proximity of the surface springs, it is assumed that groundwater discharge to the stream occurs through a high permeability lens that by attracting subsurface flow creates a preferential flow path. Thus the sift of potential discharge zones on 20 October is believed to be a result of the rain event.

With the exception of the 13 December dataset, when potential groundwater discharge was detected approximately in the middle of the stream, discharge zones are usually located close to the streambanks. This was also observed in the modelling study of Storey et al. (2003) who found that in a gravel-bedded stream both the groundwater discharge and surface water down-welling was concentrated near the streambanks. Storey et al. (2003) also compared this to the findings of McBride and Pfannkuch (1975) who showed that in case of lakes groundwater discharge decreases exponentially with distance from the shore. This could also mean that DTS applications with only one cable row in the middle of the channel may overlook the strongest groundwater signal. On the other hand fiber optic cable deployed close to the channel bank in meandering streams is more likely to be exposed to the dynamic sedimentation/scouring processes as described in this study.

The proposed method can delineate potential high-discharge zones in soft-bedded streams based on SWI temperatures and streambed elevation surveys, thus making the fast mapping of surface water-groundwater exchange possible over large areas. Groundwater discharge at these potential zones can later be quantified by for instance vertical streambed temperature profiles. As a limitation of the method, however, groundwater discharge can occur at deposition and scouring areas as well. In this study elevation changes above \( \pm 0.05 \text{ m} \) (used as the third criterion for the delineation groundwater discharge sites) were observed at half of the study area (Table 7.2). This large area excluded from the analysis can be attributed to the sedimentation experiment, thus among natural conditions a larger proportion of the studied area is expected to fulfill the conditions also reducing the extent of unmapped areas.

### 7.6 Conclusions

The looped DTS layout was able to detect spatial and temporal variability in streambed temperatures on the spatial scale of 70 x 5 m in a soft-bedded stream even though the high discharge rapidly changes streambed temperatures whereby anomalies cannot be traced at long distances downstream. Streambed temperature measurements across the stream showed great variability with potential high groundwater discharge occurring close to the streambanks during each measurement period. While using a single fiber optic cable at the middle of a high discharge-stream, this spatial heterogeneity and the
potential discharge sites close to the banks can be overlooked, the looped DTS layout provides a more accurate characterisation of streambed temperature heterogeneities.

Sedimentation and scouring processes in this soft-bedded stream occur simultaneously, influencing the DTS recordings. Sediment deposition of more than 0.1 m on the fiber optic cable was shown to cause temperature anomalies similar to, but less strong, than those caused by groundwater discharge. Scouring on the other hand increases the diel amplitude and daily standard deviation considerably. Also depending on the mobility of the streambed and the length of the measurements, in high-discharge, soft-bedded streams it is therefore recommended to: (i) to monitor the changes in the position of the fiber optic cable relative to the streambed; (ii) use both the daily minimum/maximum temperatures (depending on the season) and the diel amplitude and daily standard deviation as metrics for detecting potential groundwater discharge sites.

Making use of the dynamic nature of sedimentation processes and consequently the deposition-induced temperature anomalies, DTS was also used in this study to monitor sedimentation processes occurring both in space and time along the study site and surrounding an artificial obstruction. This gave an insight into the lateral extension of the deposits, the steepness of morphology, and bedform migration speed.

In order to detect spatial and temporal variations in streambed temperatures and to differentiate between deposition-induced temperature anomalies and potential high groundwater discharge zones in soft-bedded streams, the combined use of DTS and streambed elevation surveys is proposed during both short-term and long-term DTS applications. Based on our experience with a soft, mobile streambed, the burial of the cable at the depositional inner bends is imminent. Thus, in meandering channels, even during short surveys, basic information about the streambed morphology is needed. For long-term measurements in soft-bedded channels we suggest detailed streambed elevation surveys as a method to monitor the relative position of the fiber optic cable and to approximate the thickness of sediment deposits.

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References


Chapter 8

Paper IV

Assessing variability of surface water - groundwater dynamics using combined hydraulic and tracer methods

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\textit{manuscript in preparation}
Abstract

Detection and quantification of groundwater discharge to streams are crucial for the assessment of nutrient and possible contaminant exchange at the surface water-groundwater interface. In lowland agricultural catchments with significant groundwater dominance this is of special interest, as the risk of excess supply of nutrients to streams is high in these zones. The aim of this study was to test the ability of different hydraulic and tracer methods combined from catchment to point scales to obtain an understanding of the groundwater-surface water dynamics controlling stream discharge in a lowland gaining headwater stream in Denmark. At catchment scale (26 km²-114 km²) the sub-catchment responses to main rain events were investigated by hydrograph separations based on Electrical Conductivity (EC) and stable isotopes $^2$H/$^1$H. At medium scale (150 m-2 km) the groundwater discharge to the stream was quantified by the use of high precision differential gauging with an Acoustic Doppler Current Profiler (ADCP), and the spatial distribution of groundwater discharge zones was investigated by the use of Distributed Temperature Sensing (DTS). At point scale groundwater fluxes to the stream were quantified by Vertical streambed Temperature Profiles (VTP). Differences in sub-catchment responses to events were detected, ranging from approximately 65% event water for the most responsive sub-catchment and less than 10% event water for the least responsive sub-catchment. The spatial pattern of groundwater discharge was detected in detail by the DTS measurements and quantified by ADCP measurements. The DTS and ADCP measurements indicated high spatial variability in the groundwater discharge, where a significant part of groundwater discharge was concentrated in few major inflow zones. VTP measurements suggested high groundwater fluxes in areas detected by DTS and ADCP, and this coupling of ADCP, DTS and VTP proposes a novel method to detect areas of concentrated groundwater discharge in detail. Furthermore, the use of environmental tracers for hydrograph separations proved useful for obtaining information about sub-catchment functioning during events despite the damping from the high groundwater signal.
8.1 Introduction

Understanding of groundwater and surface water exchange dynamics is of great importance for a broad range of disciplines within the field of hydrology. Both catchment and point scale spatial and temporal variability in groundwater-surface water interactions play a decisive role in sustainable river basin management (Boulton et al., 2010). For instance, the interaction between groundwater and surface water in the hyporheic zone is a prerequisite for maintenance and improvement of stream biodiversity (Malcolm et al., 2003; Hayashi and Rosenberry, 2002). Also, detection of losing and gaining stream reaches is crucial for appropriate and effective management of water resources (Nyholm et al., 2003). Furthermore, the implementation of the EU Water Framework Directive has entailed an increased focus on the protection of existing, as well as newly established, wetlands, floodplains and riparian zones in order to lower nutrient loads to rivers, lakes, and seas (DME; Hoffmann and Baattrup-Pedersen, 2007; Griffith et al., 2006; Kronvang et al., 2005). In this context, the zones of interaction between surface water and groundwater govern the transfer of solutes and nutrients (Dahl et al., 2007; Gooseff, 2010; Kasahara and Hill, 2008; Krause et al., 2008).

Interactions between groundwater and surface water are controlled by a range of complex temporal and spatial processes governed by catchment geology, physiography, hydrology and hydrometeorology (Brunke and Gonser, 1997; Sophocleous, 2002; Winter, 1999). Due to the resulting complexity in groundwater-surface water dynamics, there is an urgent need to improve our understanding of these processes at several scales and develop new approaches as well as simple tools for mapping them. Consequently, a significant number of studies on groundwater-surface water interactions have already been conducted at different spatial and temporal scales, in different hydrological settings and by the use of a range of different methods, as summarised by Sophocleous (2002) and Kalbus et al. (2006).

At catchment scale (kilometres) transit time models have been used to infer internal catchment flow processes and flow paths by the use of environmental tracers (McGuire and McDonell, 2006; Tetzlaff et al., 2007). Numerical groundwater models have been applied to describe groundwater flow paths (Pint et al., 2003) and to assist in the detection of zones of significant groundwater discharge. Catchment responses to large precipitation events have been investigated by stream hydrograph separations, where stable isotopes and chemical tracers have been used to explain runoff formation processes (Sklash and Farvolden, 1979; Uhlenbrook and Hoeg, 2003).

At medium scale (10 m - 1 km) differential flow gauging and tracer dilution experiments are commonly applied tools for measuring net differences in stream discharge caused by groundwater recharge and discharge (McCallum et al., 2012; Briggs et al.,
Recent advances of ADCP instruments for discharge measurements have made it possible to obtain discharge data with short measurement periods and with a high precision (Mueller and Wagner, 2009), which opens up new possibilities for high precision differential flow gauging.

Groundwater fluxes at small spatial scale (<1 km) have been studied by the use of seepage meters (Landon et al., 2001; Langhoff et al., 2006; Rosenberry, 2008) and temperature as a natural tracer (Conant, 2004). Groundwater fluxes can be quantified both from vertical temperature profiles (VTP) (Schmidt et al., 2007; Jensen and Engesgaard, 2011) and from temperature time series measured in the streambed (Hatch et al., 2006; Keery et al., 2007). Recently, Distributed Temperature Sensing (DTS) has become a widely used method for monitoring temperatures at the sediment-water interface with a spatial and temporal resolution down to 0.25 m and 1 min, respectively (Selker et al., 2006a; Tyler et al., 2009), thus detecting groundwater discharge over longer stream sections and bridging the gap between <1 km and 1-10 km scales (Lowry et al., 2007).

The necessity of combining the different hydraulic and tracer methods across different scales in order to obtain useful descriptions of groundwater-surface water dynamics is widely recognised (Bencala et al., 2011; Kalbus et al., 2006; Lischeid, 2008; Scanlon et al., 2002). Therefore, studies have focussed on integrated approaches where groundwater and surface water dynamics are studied at different spatial scales with different methods. In the study by Briggs et al. (2011) DTS, differential flow gauging and chemical tracers were combined to quantify groundwater inflow to streams. McCullum et al. (2012) combined chemical tracers and differential flow gauging and Krause et al. (2012) combined DTS data with measurements of hydraulic head gradients to detect discharge zones in a lowland river in the UK. Recently, Schmadel et al. (2013) conducted a comprehensive study combining a range of methods from point to reach scale. They emphasise the importance of combining measurements at different spatial scales to avoid wrong inferences regarding exchange processes based on interactions at one spatial scale only.

Accurate characterisation and quantification of groundwater-surface water interactions present special challenges in highly groundwater-influenced lowland catchments (Krause et al., 2012). This is due to the fact that groundwater discharge can occur continuously along the stream, interchanging with concentrated discharge sites that contribute significantly to base flow. Furthermore, the catchment scale response to rain events is only moderate, providing only very general information about groundwater discharge dynamics. Medium scale measurements, like differential flow gauging, can quantify groundwater discharge along the stream, but without locating the concentrated high discharge zones. DTS measurements can detect the locations of significant
discharge areas, but they may not be capable of quantifying fluxes since the temperature signal of groundwater gets immediately diluted when entering the stream (Sebok et al., in prep.). Point scale measurements permit both detection of these zones and quantification of fluxes, but thorough mapping of large stream sections is excessively time-consuming.

Here, we report on a study where hydraulic (ADCP, hydraulic heads) and tracer (hydrograph separations from EC and $^2$H/$^1$H, DTS and VTP) methods were combined to assess the temporal and spatial variability in groundwater-surface water interactions from catchment scale to point scale in a lowland gaining headwater stream in Denmark. The aim of the study was to test the ability of combining hydraulic and tracer methods across different scales to develop an understanding of the overall groundwater-surface water dynamics controlling stream discharge in a highly groundwater dominated stream network. The specific objectives were to; (i) describe and compare variability in subcatchment (42 km$^2$-114 km$^2$) responses to rainfall events through hydrograph separations; (ii) assess the spatial variability in groundwater-surface water interactions along a 2 km stretch of the stream by combining high precision ADCP differential flow gauging (intervals of 150-200 m) with a novel coupling of DTS (spatial resolution of 1 m) and VTP (point measurements); and (iii) compare the results obtained at the different spatial scales.

8.2 Field site

The study was carried out in the gaining lowland Holtum stream, located in the Skjern river catchment in Jutland, Western Denmark (Figure 8.1a). This glacial floodplain valley is characterised by sediments of sand and silt deposited during the latest glaciation, the Weichsel glacial period (Houmark-Nielsen, 1989). The mean annual precipitation in the catchment is 950-1000 mm with an actual evapotranspiration of 460-480 mmyr$^{-1}$ (Ringgaard et al., 2011). Average annual air temperature in the catchment was 7.5 °C in 2012 with stream temperatures between 1 and 16 °C during the year. The average annual discharge at the catchment outlet was 1.2 m$^3$s$^{-1}$ and the 5th and 95th percentiles were 0.7 and 2.1 m$^3$s$^{-1}$ respectively, for the period 1994-2012.

Data collection was carried out at three stations along the main stream network (stations 1, 2 and 4) and at one station (station 3) located in a tributary (Figure 8.1b, c) confluencing with the main stream between stations 2 and 4. Between stations 1 and 4 the stream flows from east to west with a mean gradient of 0.001. Four main tributaries flow into Holtum stream between the two stations (Figure 8.1b). Between stations 2 and 4 there is a small inlet from a fishery, constantly carrying 0.07 m$^3$s$^{-1}$. Beyond
Table 8.1: Catchment characteristics and land use for each sub-catchment, with mean annual discharge, catchment size, specific discharge, distance from the source and land use.

<table>
<thead>
<tr>
<th></th>
<th>Mean annual discharge (m² s⁻¹)</th>
<th>Catchment size (km²)</th>
<th>Specific discharge (l s⁻¹ km⁻²)</th>
<th>Distance from the source (km)</th>
<th>Urban (%)</th>
<th>Agriculture (%)</th>
<th>Forest (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Station 1</td>
<td>0.17</td>
<td>26</td>
<td>7</td>
<td>0.6</td>
<td>27</td>
<td>51</td>
<td>20</td>
</tr>
<tr>
<td>Station 2</td>
<td>0.8</td>
<td>70</td>
<td>11</td>
<td>12.7</td>
<td>21</td>
<td>56</td>
<td>22</td>
</tr>
<tr>
<td>Station 3 (tributary)</td>
<td>0.28</td>
<td>42</td>
<td>7</td>
<td>11.6</td>
<td>16</td>
<td>41</td>
<td>41</td>
</tr>
<tr>
<td>Station 4</td>
<td>1.2</td>
<td>114</td>
<td>11</td>
<td>14.7</td>
<td>13</td>
<td>53</td>
<td>34</td>
</tr>
</tbody>
</table>

a) For station 3 it is distance to the source of the tributary.

a riparian zone of approximately 5 m, station 1 is surrounded by agricultural fields, whereas the near-stream areas at stations 2, 3 and 4 are wetlands. The mean annual discharge, the corresponding topographical catchment and land use of each station are summarised in Table 8.1. Hourly precipitation data was available from Voulund field site, located 6 km in the north-western direction from station 4 (not shown here).

Figure 8.1: Map of the study area and sampling sites. (a) The location of the study area in Denmark. (b) Locations of the stations of event samplings and their corresponding catchments. (c) The campaign measurements conducted between station 2 and 4.
8.3 Methods

The study was conducted through spring, summer and autumn event sampling of stream water at stations 1-4 during three larger rainfall events in 2012. Furthermore, a one week campaign was carried out between stations 2 and 4 during a low-flow period in June 2012 where DTS, ADCP and VTP measurements were conducted (Figure 8.1c). In addition, water samples and hydraulic heads were collected from piezometers installed in riparian zones/wetlands at stations 1, 2 and 4. The different types of measurements are summarised in Table 8.2 and precipitation and stream discharge values during the event samplings and the campaign are shown in Figure 8.2. Details on data collection procedures and analysis are given in the coming sections.

<table>
<thead>
<tr>
<th>Scale</th>
<th>Measurement</th>
<th>Time of measurement</th>
</tr>
</thead>
<tbody>
<tr>
<td>Catchment</td>
<td>EC and δ^2H</td>
<td>20-30 Apr 2012, 8-14 May 2012, 21-30 Sept 2012</td>
</tr>
<tr>
<td>Medium/Campaign</td>
<td>ADCP</td>
<td>9-13 Jun 2012</td>
</tr>
<tr>
<td></td>
<td>DTS</td>
<td></td>
</tr>
<tr>
<td>Point</td>
<td>VTP</td>
<td>9-13 Jun 2012</td>
</tr>
<tr>
<td></td>
<td>Piezometer water sampling</td>
<td>Mar 2012 and Feb 2013</td>
</tr>
<tr>
<td></td>
<td>Piezometer hydraulic heads</td>
<td>Aug 2012, Feb 2013, May 2013 (station 1)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Dec 2011, Mar 2012, Feb 2013, Jun 2013 (station 2)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>March 2012 (station 4)</td>
</tr>
</tbody>
</table>

Table 8.2: Summary of sampling periods and data collection methods, with the scale covered by the method, the method/instrument and time of measurement.

8.3.1 Water sample collection and analysis

The catchment responses during events were studied with hydrograph separation based on analysis of event stream water EC and the stable isotopes \(^2\text{H}/^1\text{H}\). The stream water samples were collected during three different precipitation events 20-30 April, 8-14 May and 21-30 September 2012. The September event consisted of one big and two smaller events (Figure 8.2) and has been split into September 1 (21-22 Sept.), September 2 (25 Sept.), and September 3 (28 Sept.) for the purpose of hydrograph separation.

Stream water samples were collected with two Teledyne ISCO 6712 and two 3700 portable samplers. During the April event the sampling interval was 5 hours and during the May event it was 4 hours for the first half of the event and 3 hours for the second half of the event. During the September event the sampling interval was 3 hours. Precipitation was collected in a classical Hellmann Rain Gauge and bulk water samples for isotope analysis were collected manually. During the September event there was an instrument failure at station 1, implying that no stream water data is available from the middle until the end of the 25 September event.
The collected stream water and precipitation samples were sealed in 200 ml plastic bottles and later analysed for $\delta^2$H on a PICARRO L2120-i Isotopic Water spectrometer. Isotope fractions are given in per mille relative to Vienna Standard Mean Ocean Water (VSMOW). The precision of the measurements was 0.3‰ for $\delta^2$H.

![Graph showing precipitation and discharge during 2011/2012](image)

**Figure 8.2**: Precipitation and discharge during 2011/2012. (a) Hourly precipitation measured 6 km northwest of station 4. (b) The measured discharge at the catchment outlet (station 4) with event sampling periods (green) and the medium scale campaign measurement period (red).

### 8.3.2 Hydrograph separation

To detect the variability of sub-catchment responses to rainfall events a one-tracer two-component hydrograph separation was conducted (Sklash and Farvolden, 1979), based on the three monitored events. The stream water was separated into pre-event and event water fractions on the basis of the measured $\delta^2$H concentrations. “Pre-event” water refers to water present in the catchment before the event and “event water” refers to the water that enters the catchment during the event. The mixing equation used to estimate the pre-event fractions is given by Genereux and Hooper (1998):

$$f_{pe} = \frac{C_T - C_e}{C_{pe} - C_e} \quad (8.1)$$
where $C_T$ represents the isotopic concentration in the stream water. $C_e$ represents the isotopic composition of the event water and $C_{pe}$ represents the isotopic composition in the pre-event water. The concentration in the stream water immediately prior to the events was used as $C_{pe}$, based on the assumption that the influence from event water at that time is negligible (Pinder and Jones, 1969; Sklash and Farvolden, 1979). $C_e$ represents the isotopic concentrations in the rainfall during the events. For the April and May events sparse precipitation samples were available and $C_e$ was calculated as a weighted mean and a bulk value, respectively. For the September event $C_e$ was calculated as an incremental weighted mean value of the precipitation compositions (McDonnell et al., 1990).

In order to test, whether the assumption of a strictly two component system is applicable, a simple approach is to conduct an additional one-tracer two-component hydrograph separation based on a chemical tracer. The two-component hydrograph separation based on a chemical tracer gives the fractions of sub-surface and surface water, where “sub-surface” water refers to the water which has passed through the mineral soil, and “surface” water refers to the water which has not infiltrated the mineral soil (Genereux and Hooper, 1998). In the ideal case of a completely groundwater dominated stream network, the sub-surface component would be equal to the groundwater component, and hence be equal to the pre-event component. Furthermore, the event and surface fractions would also both correspond to the rain component, since the rain which contributes to the stream water during the event, would not have passed through the mineral soil due to the high ground water table (Rodhe, 1998). However, any discrepancies between the pre-event and sub-surface fractions can indicate the presence of a third component, for instance a shallow groundwater/soil component (Wels et al., 1991). Standing alone, this approach will reveal neither the source nor the flow path of the third component, just indicate the likely presence of it, and can therefore only be useful as a first-hand indication or as a supplement to other methods.

Thus in addition to the isotope hydrograph separation, a one-tracer two-component hydrograph separation was conducted based on EC. The same mixing equation (Eq. 8.1) was applied. Instead of $C_e$ and $C_{pe}$ the EC values of the surface runoff component $C_S$ and the subsurface component $C_G$, respectively, were used. It is assumed that $C_S$ has a similar EC value as the rainfall, from the reasoning of no contact with mineral soils. Furthermore, it is assumed that the EC value of the stream water prior to the event is similar to the EC value of the subsurface component, and consequently $C_G$ can be estimated. $C_T$ represents the EC value in the collected stream water samples during the event. EC values of the rainfall were calculated as bulk, weighted mean and incremental weighted mean as already described for the $\delta^2$H values.

Uncertainties in the pre-event water fractions inherent from uncertainties in deter-
mination of the concentrations used in Eq. 8.1 were calculated based on the procedure by Genereux (1998). This method is based on an uncertainty propagation technique using Gaussian error estimators, and was calculated at the 0.05 confidence level. Uncertainties in EC and $\delta^2$H values in stream water prior to events were used to determine the uncertainty in $C_{pe}$ and $C_G$. Uncertainties in the rainfall and stream water during events were calculated based on the measurement precisions ($\pm 0.3\%$ for $\delta^2$H and 0.5% of measured EC value) since only one sample per time interval was available. This uncertainty estimation procedure only estimates the minimum total uncertainty in pre-event fractions inherent from determination of each of the tracer concentrations. Hence, it does not account for uncertainties possibly introduced by violations of any of the assumptions behind the calculated hydrograph separations.

The stream water sampling during the September event was initiated at the early stage of the stream discharge response. Consequently, no stream water samples were collected prior to the event. However, the May event had similar conditions prior to the event as the September event, and the May concentrations were therefore used as September pre-event and subsurface concentrations of EC and $\delta^2$H, respectively.

### 8.3.3 Groundwater sampling

Samples of shallow and deep groundwater were collected at stations 1, 2 and 4 (Table 8.2). Groundwater samples were taken from steel piezometers with a 3/4 inch outer diameter and 10 cm long screens installed with the direct push method. At station 1 water samples were taken from 11 different piezometers with screen depths ranging from 1 to 5 m below the surface, at station 2 from 20 different piezometers with screen depths ranging from 1.5 to 14 m, and at station 4 from 14 different piezometers with screen depths from 1 to 8 m. At station 2 water samples were taken from several depths from the two deepest piezometers. Three volumes of water were purged from the piezometers before the water samples were taken. The EC of the groundwater samples was measured on site. The groundwater samples were stored and analysed for $\delta^2$H in the same manner as the stream water and precipitation samples. In addition, the hydraulic heads in the piezometers were measured (Table 8.2).

### 8.3.4 Discharge measurements

A permanent gauging station for continuous discharge estimation is located at the catchment outlet (station 4). Continuous discharge estimates are calculated based on the stage-discharge relation, continuous water stage measurements (OTT Thalimedes
pressure transducer) and monthly current meter control measurements of discharge (Rantz, 1982; Herschy, 1999).

In the 2450 m long stream section between stations 2 and 4 (Figure 8.1c), differential gauging of discharge was carried out during the June 2012 campaign for detection of the medium scale pattern of groundwater inflow. Stream discharge was measured with an ADCP Streampro manufactured by Teledyne RD Instruments. The ADCP Streampro has a 4-beam 2 MHz transducer and a sampling frequency of 1 Hz. The ADCP measures discharge based on water velocities and cross-sectional area. The ADCP is operated from a platform and tethered across the stream, perpendicular to the main flow direction. ADCP discharge measurements were conducted for each 200 m in layout A and C and for each 150 m in layout B (Figure 8.1c). The ADCP measurement procedure was optimised according to recent recommendations (Mueller and Wagner, 2009; Muste et al., 2004a,b), and minimum ten discharge measurements with an average deviation less than or equal to 5% were made at each location in order to lower the uncertainty of the discharge estimates.

### 8.3.5 Campaign sampling: Distributed Temperature Sensing

During the June 2012 campaign DTS was used to locate zones of concentrated groundwater discharge. A BruSteel fiber optic cable connected to a Sensornet Oryx-SR system was deployed along the middle of the stream on the sediment-water interface in three layouts, A, B and C (Figure 8.1c) to cover the whole length of the stream section. To avoid damage of the fiber optic cable, no measurements were made between 1366 and 1530 m in the downstream direction from station 2 (Figure 8.1c) due to remnants of a weir.

<table>
<thead>
<tr>
<th>Layout</th>
<th>Time of measurement</th>
<th>Length (m)</th>
<th>Precision ($^\circ$C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>11 June 13:20-12 June 11:50</td>
<td>0-905</td>
<td>0.05</td>
</tr>
<tr>
<td>B</td>
<td>12 June 17:20-13 June 16:00</td>
<td>906-1366</td>
<td>0.21</td>
</tr>
<tr>
<td>C</td>
<td>9 June 18:00-10 June 17:20</td>
<td>1530-2452</td>
<td>0.04</td>
</tr>
</tbody>
</table>

Table 8.3: Time of DTS stream bed temperature measurements with the length and precision of each layout.

For each layout streambed temperature data was collected with double-ended measurements of 10 minute integration times and a 1.01 m spatial averaging interval. Each installation was calibrated by running approximately 30 m fiber optic cable through a calibration bath. The precision of the installations is shown in Table 8.3. During each layout streambed temperature time series of 22-23 hours were collected with different starting times (Table 8.3), but results are presented by aligning the measurements.
relative to time of day.

Deeper groundwater in Denmark has an average temperature of \( \sim 8^\circ C \) and the average stream temperature was \( 13^\circ C \) during the campaign. Therefore, potential groundwater discharge sites should show relatively cold streambed temperatures during the field campaign. However, due to different daily temperatures, the decrease in streambed temperatures at the potential discharge sites was not directly comparable between the layouts. In order to compare streambed temperatures measured at different days at different locations, the strength of the groundwater temperature signal for each measurement location was calculated as:

\[
S_i = \frac{T_l}{T_i} \quad (8.2)
\]

where \( S_i \) is the strength of the groundwater temperature signal at location \( i \), \( T_l \) is the mean temperature measured at the corresponding layout \( l \) during the measurement period, and \( T_i \) is the mean temperature at location \( i \) during the measurement period. Thus, \( S_i \) values above one represent colder streambed temperatures than the mean of the layout.

### 8.3.6 Campaign sampling: Vertical streambed Temperature Profiles (VTP)

Vertical groundwater fluxes were estimated with VTPs in the areas of relatively low streambed temperatures, as indicated by the DTS surveys. Such low temperature zones were measured at 18, 9, and 15 locations in Layout A, B and C, respectively. At these locations streambed temperatures were collected after 10 minutes of stabilisation at 0, 0.025, 0.05, 0.075, 0.1, 0.15, 0.2, 0.3, 0.4 and 0.5 m below the streambed by thermocouples with an accuracy of 0.2°C. Due to the firm streambed material and the long measurement time needed, VTP measurements were only conducted at the most pronounced discharge sites.

Based on the VTP measurements vertical groundwater fluxes were estimated by fitting the steady-state analytical solution of the one dimensional conduction-convection equation (Bredehoeft and Papadopulos, 1965) to the measured temperature data as described by Schmidt et al. (2007) and Jensen and Engesgaard (2011):

\[
T(z) = T_s + (T_g - T_s) \frac{\exp[N_p z/L - 1]}{\exp(N_p - 1)} \quad (8.3)
\]

where \( T(z) \) is the streambed temperature (°C) measured at depth \( z \) (m), \( T_s \) is the stream
water temperature ($^\circ$C), $T_g$ is the groundwater temperature ($^\circ$C) at a given depth $L$ (m), and $N_{pe}$ is the Peclet number giving the ratio of convection to conduction:

$$N_{pe} = \frac{q_z \rho_f c_f L}{\kappa_e}$$  \hspace{1cm} (8.4)

where $q_z$ (ms$^{-1}$) is the vertical fluid flux, $\rho_f c_f$ is the volumetric heat capacity of the fluid (Jm$^{-3}$ $^\circ$C$^{-1}$) and $\kappa_e$ is the effective thermal conductivity (Jm$^{-1}$s$^{-1}$ $^\circ$C$^{-1}$).

For each VTP, $T_s$ was given as the temperature measured by the uppermost sensor, and the constant groundwater temperature of $8^\circ$C ($T_g$) was assumed at a depth of 5 metres ($L$). A volumetric heat capacity of $4.19 \times 10^{-6}$ Jm$^{-3}$ $^\circ$C$^{-1}$ was used for the water, and effective thermal conductivity of 1.8 Jm$^{-1}$s$^{-1}$ $^\circ$C$^{-1}$ was assumed for the sandy streambed.

8.4 Results

8.4.1 Variability in tracer responses

The variability in stream water $\delta^2$H and EC showed an overall tendency of being damped downstream in the catchment with decreasing standard deviations in the downstream direction (Figure 8.3). During all three events station 1 had the largest variability in both EC and $\delta^2$H values, whereas station 3 had the smallest variability during all three events. The April event exhibited the smallest variability in $\delta^2$H response at all stations, whereas the September event showed the largest response (Figure 8.3a). The same trend was seen for EC values, except that there was large variability in the response at station 1 during the April event as well (Figure 8.3b).

For $\delta^2$H values, the stations showed similar temporal trends between the events, but the May event showed slightly higher average $\delta^2$H values than the April and September events (Figure 8.3a). Generally, the average $\delta^2$H values for all stations and all events showed only modest differences, with the lowest average $\delta^2$H (-55.3 $\%_o$) occurring during the September event at station 1 and the highest average $\delta^2$H (-51.2 $\%_o$) during the May event at station 2 (Figure 8.3a). Also, for EC the stations showed similar temporal behaviour with generally lower average EC values during the September event as compared to the May and April events (Figure 8.3b). For EC, the average values ranged from 238 $\mu$Scm$^{-1}$ at station 2 during the September event to 307 $\mu$Scm$^{-1}$ at station 1 during the May event (Figure 8.3b).
Figure 8.3: Variability in tracer values measured during event sampling of stream water. (a) $\delta^2$H and (b) EC values measured in the stream water during the events at the four stations. The data points are shown in blue together with the mean, STD and 95% confidence intervals of data from each event at each station.

8.4.2 Hydrograph separation

The conditions for the three monitored events are summarised in Table 8.4, and the concentrations used for calculating the hydrograph separations are shown in Table 8.5. The September 1 event showed the largest increase in discharge of all the monitored events (Table 8.4). The intensity of the September 2 event was similar to the May and April events. The September 3 event intensity was similar to September 1 but a shorter duration; thus, the resulting rise in stream discharge (35%) was more similar to the April and September 2 events (Table 8.4). The May event was slightly more intense than the April and the September 2 events and resulted in a larger increase in discharge (70%).

The largest event response was detected during the September 1 event. The $\delta^2$H-based pre-event fractions and uncertainties during the September event for all four stations are depicted in Figure 8.4 together with the hourly precipitation and hourly discharge at the catchment outlet (station 4). Uncertainties deriving from uncertainties in determining the tracer concentrations for the mixing Eq. (8.1) were generally less than 10% during the September events (Figure 8.4c-f).

Generally, station 1 showed the quickest and most significant event response with the pre-event fraction reaching a minimum of 35% during the September 1 event (Fig-
Table 8.4: Summary of rainfall and runoff characteristics with rainfall intensity and duration, peak discharge, maximum discharge increase and number of rain samples for each precipitation event. September is divided into three sub-events.

Pre-event fractions during peaks of the April, May and September events showed similar trends between stations (Figure 8.5a). Generally, station 1 showed the largest event responses, station 2 and 4 reacted similarly, though with a tendency for station 4 to have slightly larger pre-event fractions, except for the May event. The September 1 event showed a distinctly larger response at all stations compared to the other events. During all events station 3 only showed modest response with consistently more than 70% pre-event fractions (Figure 8.5a). From the uncertainty calculation, uncertainties at the peaks of the event responses depicted in Figure 8.5 were all below 10% (not shown).
Figure 8.4: Runoff and precipitation characteristics and pre-event fractions for the September event. (a) Hourly precipitation measured 6 km northwest of station 4. (b) Catchment runoff measured at station 4. (c-f) Pre-event fractions for stations 1, 2, 3 and 4, respectively. Lag times between peak responses are indicated with red lines and uncertainties of the pre-event fractions are given by the blue band.
The subsurface fractions showed similar trends between stations as the pre-event fractions, with station 1 being most responsive and station 3 least responsive. However, except for the September 1 event, no difference was observed between sub-surface fractions at station 2 and 4, contrary to what was observed for pre-event fractions. Also, the sub-surface fractions for station 1, 2 and 4 varied less than the pre-event fractions between events (Figure 8.5). For instance, at station 4 for the May, April, September 2 and September 3 events the sub-surface fractions varied only between 80 and 90%, whereas the pre-event fractions varied between 65 and 95% (Figure 8.5). At all stations the September 1 event showed the most pronounced surface response at all stations and the sub-surface fractions at station 1, 2 and 4 were smaller than the pre-event fractions.

Figure 8.5: (a) Pre-event fractions and (b) sub-surface fractions for all events. The September event was subdivided into three sub-events and trend lines are added between pre-event and subsurface fractions calculated for the main stream. Uncertainties were less than 10% for all fractions (not shown).

8.4.3 Cross sectional variability in hydraulic heads and tracer values

Based on the analysed water samples from the piezometers, deep groundwater was found to have an average $\delta^2$H value of $\sim 56 \%e$. Average, minimum and maximum values of $\delta^2$H and EC for stations 1, 2 and 4 are summarised in Table 8.6. In the majority of the piezometers installed at stations 1, 2 and 4, the groundwater table is constantly less than 2 m below ground. Iso-potential maps based on hydraulic head observations were used to interpret and draw general groundwater flow paths in cross sections at stations 1, 2 and 4 (Figure 8.6) with the assumption that flow
paths are perpendicular to iso-potential lines. Due to the limited fluctuations observed in groundwater levels at stations 1, 2 and 4 (<15 cm), it is assumed that the flow conditions depicted in Figure 8.6 are representative of the general flow patterns for the whole study period. The measured δ²H values in the groundwater were used to interpolate contour lines in Figure 8.6.

Figure 8.6: Interpolated cross-sectional contour maps of δ²H values and groundwater flow paths. (a) Cross-section at station 1 based on data from February 2013. (b) Cross-section at station 2 based on data from February 2013. (c) Cross-section at station 4 based on data from March 2012. The horizontal red lines represent the screen depth of the piezometers where the water samples were taken. The red arrows show the groundwater flow paths based on interpolated hydraulic head data which were measured when the water samples were taken, except for station 2 where they were recorded in June 2013. The thickness of the arrows is proportional to the hydraulic gradients.

At station 1, head gradients (not shown here) were significantly smaller than at station 2 and 4. To the right of the stream flow paths were directed upwards with no systematic trends of flow towards the stream. Right beneath the stream a distinct
upward flow was observed (Figure 8.6a) which coincided with lower $\delta^2$H values. $\delta^2$H values were generally less negative closer to the surface. However, no consistent pattern between $\delta^2$H and groundwater flow paths could be observed deeper down in the profile.

At station 2, groundwater flow paths indicated upward groundwater flow to the left of the stream (Figure 8.6b). This coincided with relatively homogeneous $\delta^2$H values with increasing values towards the surface. To the right of the stream a more lateral upward flow towards the stream channel was observed with a lower variability in $\delta^2$H. At the far right of the stream a consistent horizontal stratification in $\delta^2$H was observed in the deep borehole, with values decreasing downwards.

At Station 4 to the left of the stream, groundwater flow paths indicated an upward flow towards the wetland. A high gradient in $\delta^2$H values occurred at the left bank of the stream, with increasing $\delta^2$H values at increasing distance from the stream (Figure 8.6c). However, directly beneath the stream $\delta^2$H values of -55 $/\%_o$ and -54 $/\%_o$ were measured at the same location where the groundwater flow paths indicated a strong upward flow.

### 8.4.4 Campaign sampling of medium and small-scale groundwater discharge

The results from the June 2012 campaign are summarised in Figure 8.7. There was a difference of 1.2 cm in water stage between the start and end of the June 2012 campaign and a 2.5 cm difference between the measured highest and the lowest water stages. From the start to the end of the campaign there was an overall decrease of 9% in discharge at the catchment outlet (station 4).

In general, the ACDP measurements showed an increase in discharge from 0.7 to $\sim$ 1.2 m$^3$s$^{-1}$ along the 2450 m long study stretch (Figure 8.7b). Measurements in Layout
A showed a steady gradual increase in discharge until the last 200 m section where a sudden increase in discharge occurred, coinciding with the inflow of the tributary with an average discharge of 0.23 m$^3$s$^{-1}$ during the measurement period. Stream discharge decreased in Layout B between 906 and 1117 metres, followed by a gradual recovery to a slightly higher discharge than the initial value at the beginning of the layout.

![Diagram](image)

Figure 8.7: DTS, VTP and ADCP measurements from the campaign sampling in June 2012. (a) DTS temperatures measured in Layouts A, B and C between station 2 and 4. The locations of stations 2 and 4 are shown as well as the location of the tributary inflow. (b) ADCP discharge measurements are shown in blue, red and orange, combined by trend lines. Dashed and solid trend lines indicate distinctions between separate measurement rounds. Green dots show VTP-based fluxes.

Stream discharge in Layout C showed the steadiest increase of all layouts, but also the highest difference between the two rounds of discharge measurements along the layout. The mean difference in discharge of 0.073 m$^3$s$^{-1}$ between the two measurements in Layout C was due to a rain event of 20.4 mm on 9-10 June after which the stream discharge rose by approximately 8%. However, the increase mainly occurred during the night, and the ADCP discharge measurements conducted along Layout C are therefore not expected to be significantly affected.

Based on streambed temperature data recorded by the DTS, potential high groundwater discharge sites were detected at 600, 705, 735, 800 and 825 m downstream along
the cable at Layout A (Figure 8.7a). At Layout B a potential high discharge site was observed at 1205 m. The cold temperature anomaly at 1300 m is expected to be a result of the water input from the fishery to the stream and is thus not considered a potential high discharge site. At Layout C the most pronounced groundwater discharge sites were detected at the most downstream end at 1900, 1980, 2285, 2380 and 2415 m. Due to the rain event on 9-10 June, the air temperature decreased and the lowest streambed temperatures were, therefore, measured in Layout C.

8.4.5 Vertical streambed temperature profiles

VTP measurements were carried out at the potential discharge sites as indicated by the DTS. In Layout A estimated vertical groundwater fluxes ranged from 0.09 to 1.3 md\(^{-1}\) with a mean of 0.44 md\(^{-1}\) (Figure 8.7b), showing high spatial variability of fluxes within short distances. In Layout B a minimum and maximum flux of 0.07 md\(^{-1}\) and 0.52 md\(^{-1}\) were estimated, the lowest flux occurring at the decreasing limb of the discharge curve. In Layout C fluxes from 0.06 md\(^{-1}\) to 0.86 md\(^{-1}\) were observed with a mean of 0.29 md\(^{-1}\).

![Figure 8.8: Correlation of the strength of the groundwater signal as recorded by the DTS with upward groundwater fluxes estimated from the vertical temperature profiles.](image)

For Layout A and B the comparison between estimated upward groundwater fluxes and the strength of the groundwater signal (Eq. 8.2) at the corresponding DTS locations showed a linear correlation (correlation coefficients of 0.64 and 0.52, respectively) (Figure 8.8). This indicated that higher upward fluxes usually coincided with a stronger groundwater signal. In Layout C, however, a linear correlation was not detected.
8.5 Discussion

8.5.1 Differences between sub-catchment behaviour

The pre-event fractions revealed that the sub-catchment of station 1 was consistently the most responsive and that of station 3 the least responsive (Figure 8.5a). Although the sub-catchment drained by station 3 is twice the size of that drained by station 1, their sub-catchments are comparable regarding land use and specific discharges (Table 8.1). However, compared to the wetland surrounding station 3, station 1 has a larger proportion of agricultural fields with tile drains in the near stream area, which may explain some of the quick response.

Based on the small and the slow, delayed event responses observed in pre-event fractions at station 3 (see Figure 8.4e, for September event, 8.5a), the tributary is expected to be highly groundwater-dominated. Additionally, no significant differences were seen between pre-event fractions and sub-surface fractions at station 3 (Figure 8.5). Thus, most likely the assumption of two end-members in the hydrograph separation was met, i.e. subsurface water consisted solely of groundwater, causing the pre-event and sub-surface fractions to represent the same water. That indicates, that most of the precipitation reaches the groundwater and no stored shallow/soil component is released during the event. This mechanism has previously been observed in areas of shallow groundwater tables, where the saturated zone only builds up in connection with the groundwater table (Rodhe, 1998). However, to further elaborate this explanation, additional measurements would be preferable of, for instance, variations in groundwater gradients during events.

Stations 2 and 4 revealed similar event responses, intermediate of those observed at station 1 and 3. There was a tendency for station 4 to be damped in the pre-event responses as compared to station 2 (Figure 8.4d, f and 8.5a). This is expected to be partly due to the inflow from the groundwater-dominated tributary between station 2 and 4 and partly due to the significant groundwater discharge to the stream as detected during the field campaign. From the specific discharges (Table 8.1) it is seen that the topographical areas drained by station 2 and 4 are smaller relative to the average stream flow than at station 1. Hence, indicating that the groundwater contribution is less pronounced at station 1, compared to station 2 and 4. Together with the mentioned influence from the tile drains at station 1, this is expected to explain the relatively quicker and larger event response as compared to stations 2 and 4. This is in accordance with observed groundwater gradients (indicated with arrow thickness in Figure 8.6), which reveal significantly smaller gradients towards the stream at station 1 compared to stations 2 and 4.
At stations 1, 2 and 4 discrepancies were observed between subsurface and pre-event fractions (Figure 8.6). These discrepancies could indicate the occurrence of a component which is accounted for by neither of the two hydrograph separation techniques (Wels et al., 1991; Hooper and Shoemaker, 1986). Consequently, a near surface water contribution from this zone may exist, which is a mix of rainfall, shallow groundwater and perhaps “old” event water as discussed by Rodhe (1998). Thus, due to the contribution of this zone, the composition of the water pushed out of the riparian zone during high rainfall events might be highly variable over time. Stations 2 and 4 are surrounded by wetlands and in addition the groundwater flow lines suggest groundwater upwelling to the wetlands (Figure 8.6). Furthermore, shallow groundwater tables were observed at stations 1, 2 and 4 (<2 m at all times) as well as field observations of frequent standing water during events in the wetlands surrounding stations 2 and 4 and partly station 1. This also supports the possible presence of a mixing zone between groundwater and precipitation, which is released during events due to saturation by the raising groundwater table.

The presence of a shallow mixing/storage zone at stations 1, 2 and 4 are somewhat supported by the relatively large variability of $\delta^2$H of the groundwater in the upper 1-5 m, especially left of the stream at stations 2 and 4 (Figure 8.6, Table 8.6). Hence, indications are that the stream flow components during events could be divided into a deep groundwater component discharging right beneath the stream channel (Figure 8.6), a shallow mixed component of groundwater/soil/old event water and a surface/event water component. This is in agreement with Karan et al. (2013) who found a shallow relatively young groundwater component discharging to the stream at station 4. However, based on the available data, there was no distinct difference between the average EC and $\delta^2$H of the shallow soil/groundwater and the deep groundwater. For instance, a separation of piezometers into shallow (0-3.5 m) and deeper screen depths (3.5-13 m) at stations 2 and 4 yielded $\delta^2$H values of -53.4±2‰ and -52.8±2‰ and EC values of 260±110 $\mu$Scm$^{-1}$ and 266±103 $\mu$Scm$^{-1}$, respectively. Consequently, the prerequisite of distinct differences in end members for a two-tracer three-component hydrograph separation is not met with the given dataset (Genereux and Hooper, 1998).

The exact pre-event fractions as well as sub-surface fractions calculated should thus be interpreted with caution especially for stations 1, 2 and 4. Furthermore, it should be noted, that no measurements of soil EC values were available during the study period. Hence, the assumption of EC acting as a conservative tracer might not be fully met. However, based on the complementary information from well data and field observations, the general pattern of sub-catchment behaviour during events is expected to be plausible. For further investigation of the sources constituting the event hydrograph, sampling of shallow ground/soil water prior to events would be required
8.5.2 Threshold for event responses to be triggered

The size of the hydrograph responses in terms of pre-event were greatly influenced by
the rainfall intensity. For instance, during the September 3 event, the discharge increase
(at station 4) was approximately 10% less than during the April and September 2 events
(Table 8.4). However, the intensity of the rain event was twice that observed during
both the April and September events, which is expected to be the reason for the larger
event fraction observed at all stations. In contrast, for the sub-surface fractions only
the September 1 event resulted in a significantly different response compared to the
other events. However, both the pre-event fractions as well as the subsurface fractions
suggest that an event as in April, with 15 mm rain and a resulting discharge increase
of 30-50%, constitutes a threshold below which the stream flow components are not
altered.

8.5.3 Detection of spatial variability in groundwater discharge
by campaign measurements

Detailed, medium to small-scale spatial variation in groundwater discharge to the
stream was characterised by DTS, ADCP and VTP measurements. DTS measurements
have previously been used to detect and also to quantify concentrated groundwater
discharge to streams (Selker et al., 2006b; Briggs et al., 2011) based on a tempera-
ture mixing approach combined with differential gauging upstream and downstream of
indicated discharge sites. DTS results from June 2012 showed drops in streambed tem-
peratures of 0.5-1 °C due to groundwater discharge (Figure 8.7a), but instead of large
step changes in temperatures (Selker et al., 2006b; Briggs et al., 2011) the groundwater
discharge did not alter the downstream temperatures as also observed in a wetland
stream (Lowry et al., 2007) and in a Danish stream with a much lower mean discharge
of 0.25 m³s⁻¹ (Karthikeyan et al., 2012). Thus, quantification of discharge using a
mixing analysis was not possible due to the small temperature contrast between sur-
face water and groundwater and with the high stream discharge rapidly diluting the
temperature signal of the upwelling groundwater. ADCP measurements, on the other
hand, gave a good estimation of net groundwater discharge between measurement sec-
tions. However, due to the spacing of these sections the information was spatially
course compared to the metre-scale spatial resolution of the DTS and the point-scale
flux estimation by VTPs. The ADCP and DTS methods combined can provide a detailed picture of groundwater-surface water interactions, but great logistical effort is required in order to map stream stretches longer than a few kilometres.

The discrepancy between the spatial scales which each of the methods is capable of resolving is illustrated when comparing the ADCP measurements to the DTS and VTP data. Since ADCP is expected to measure discharge within an uncertainty of 5%, there exists a lower limit for measurement intervals during differential gauging. For this study, intervals of approximately 150-200 m were close to the lower limit, especially for Layout A and C, which demonstrated the most gradual changes in discharge increase. Consequently, the ADCP method was not capable of showing the same spatial variability in groundwater discharge as the DTS metre-scale and the VTP point measurements. For this reason, it was also possible to still detect cold temperature anomalies indicating groundwater discharge and relatively high upward fluxes of 0.43 \( \text{md}^{-1} \) in a generally losing stream section (Figure 8.7b). Schmadel et al. (2013) found similar discrepancies between methods mapping medium to small-scale groundwater inflow. These findings emphasise the importance of combining methods covering different scales to avoid ambiguity or wrong inferences due to extrapolation of results between scales.

### 8.5.4 Coupling of VTP and DTS measurements

So far, only a few studies have endeavoured to confirm the groundwater discharge sites indicated by the DTS with estimates of discharge based on either seepage meter data or vertical temperature profiles (Lowry et al., 2007; Sebok et al., 2013). This study shows that VTPs reflect the same spatial variability in groundwater discharge as the DTS, detecting both very high and very low upward fluxes within metres (Figure 8.7). Moreover, there was a good linear correlation between upward fluxes and the strength of the groundwater signal detected by the DTS for Layout A and B (Figure 8.8). A similar good correlation was not seen for Layout C. This section showed, however, that even if the streambed temperatures were higher than the mean, thus no high discharge was expected, upward fluxes up to 0.15 \( \text{md}^{-1} \) could still be measured, indicating that the stream is completely groundwater dominated. In spite of the discrepancy of estimated fluxes and groundwater signal strength in case of Layout C, the DTS and VTPs still complement each other (Figure 8.7), yielding an estimate of upward flux for a given signal strength. This confirms that cold streambed temperature anomalies correspond to locations of high upward groundwater fluxes during summer periods.
8.5.5 Small-scale spatial variability in groundwater discharge

Both the catchment-scale and the medium-scale measurements confirm that this part of the stream is groundwater-dominated. By correcting the net stream discharge measurements between station 2 and station 4 for the tributary and the inlet from the fishery, the groundwater discharge to the stream resulted in approximately 30% increase in total stream discharge. However, similarly to the DTS observations of Lowry et al. (2007); Briggs et al. (2011) and the VTP-based flux estimates of Schmidt et al. (2007) and (Anibas et al., 2011), the spatial distribution of groundwater discharge was not homogeneous along the stream section. The most gradual groundwater inflow was observed along Layout C, where ADCP revealed a steady increase in discharge, while along Layouts A and B larger variability in streambed temperatures, ADCP discharges and VTP fluxes was observed (Figure 8.7). For instance, discharge increased with approximately 16% in the second half of Layout A (450 m), excluding the contribution of the tributary. This zone coincided with the most distinct zone of discharge as detected by the DTS and also with the strongest groundwater signals and highest VTP fluxes. In the last section of Layout B (350 m), a 10% increase in discharge coincides with the second largest zone of inflow, as detected by the DTS.

Consequently, our results suggest that a significant part of the groundwater inflow along the studied reach is concentrated in relatively few zones. These concentrated discharge sites indicate that most of the upwelling groundwater reaches the streams via preferential flow paths governed by differences in streambed hydraulic conductivity and hydraulic head conditions (Sophocleous, 2002). Since these zones of high groundwater inflow will also carry the largest amounts of, for instance, nutrients or potential contaminants, their detection and quantification are of great importance. This is of special interest for gaining lowland streams in agricultural areas, including a significant part of the Danish streams due to the potential of high nutrient loads, as discussed by Krause et al. (2012).

8.6 Conclusions

We studied the groundwater-surface water dynamics from catchment scale to point scale in a gaining lowland headwater stream in Denmark. The aim of the study was to test the ability of combining hydraulic and tracer methods across different scales to obtain an understanding of the overall groundwater-surface water dynamics controlling stream discharge in a highly groundwater-dominated catchment. The study illustrated the strength of combining different hydraulic and tracer methods across different spatial
scales to assess the groundwater flow dynamics in a groundwater-dominated catchment; assuming that this approach may also be applicable in other lowland catchments.

The hydrograph separations conducted for the three sampled rain events at the four different stations revealed distinct differences in event responses between the four sub-catchments. The most upstream sub-catchment drained by station 1 generally showed the largest event responses, with a minimum pre-event of 35% occurring during the September event. Station 3, representing a tributary to Holtum stream, was the least responsive to events, and only showed a significant event response during the largest September event, with approximately 70% pre-event water at the peak of the response. Stations 2 and 4 at the more downstream part of the catchment showed similar event responses intermediate of those observed at station 1 and 3. Discrepancies between pre-event and subsurface fractions were found at station 1, 2 and 4 but not at station 3. This is explained by the likely presence of a shallow groundwater/unsaturated zone component contributing to event runoff at stations 1, 2 and 4, which is not accounted for in the two-component hydrograph separation.

The event responses were damped downstream indicating an increasing groundwater influence. This was supported by the medium-scale investigations of the spatial pattern of groundwater inflow, detected by the DTS measurements and quantified by ADCP measurements. Significant groundwater discharge was observed, resulting in a stream discharge increase of approximately 30% between stations 2 and 4. The groundwater discharge was found to be primarily confined in few distinct zones, suggesting the presence of preferential flow paths. The major zones of groundwater discharge detected by DTS and ADCP measurements were supported by point scale VTP measurements indicating high groundwater fluxes. This coupling of ADCP, DTS and VTP proposes a new method to detect areas of concentrated groundwater discharge in detail. Since these zones of high groundwater inflow will also carry the largest amounts of for instance nutrients or potentially contaminants, the detection and quantification of them is of great importance. This is of special interest especially for lowland gaining streams in agricultural areas such as a significant part of the Danish streams, due to the potential of high nutrient loads.

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References


Chapter 9

Paper V

Air/water/sediment temperature contrasts in small streams to determine groundwater discharge

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Abstract

The concept of linear regression analysis of stream temperature versus air temperature is advanced to include additional temperature measurements in the streambed-water interface to more precisely identify groundwater-stream interaction. Temperature variations were investigated in a 15 km stream and showed increased influence of groundwater discharge at four locations out of ten. Parameters from the regression model together with mean diel amplitude are efficient for identifying areas of increased groundwater discharge. We show that the differing correlations between the temperature of air/stream and air/streambed-water interface ($P < 0.0001$) measured at four locations are strongly affected by groundwater discharge. Mean diel amplitudes $\leq 0.6 \, ^\circ C$ correlate with observed increase in flow accretion and vice versa. Due to the simplicity of both the implementation and analysis the method offers a cost-effective alternative for known methods of identifying groundwater discharge areas.

9.1 Introduction

The identification of groundwater discharge to streams is desirable for example for managing stream water quality since specific stream reaches can act as focal points for discharge of groundwater and nutrients (Winter et al., 1998). Temperature measurements have in recent years been used in a large number of groundwater-stream studies to identify discharge zones with applications becoming more sophisticated with the technological advancement in sensors being able to measure temperature more accurately and their easy deployment in streams or wells (Constantz, 2008). Distributed Temperature Sensing (DTS) is a new promising technology capable of covering large stream reaches from the meter to kilometer range with high resolution (Lowry et al., 2007). The DTS equipment is expensive and to use DTS at the scale of a stream network and over long periods requires many resources. To meet the needs of identifying groundwater discharge zones in a cost-effective manner within several kilometers of stream network we expand on the concept of using linear regression models of air/stream temperature relationships (Stefan and Preud’homme, 1993; Erickson and Stefan, 2000).
A linear regression model of stream temperature versus air temperature is used as a proxy for groundwater-stream interaction based on the premise that a regression line close to unity reflects that stream temperature is governed by air temperature. Deviation from unity indicates that other processes impact stream temperature. For instance, significant groundwater discharge or shading would decrease the effects of air temperature on stream temperature lowering the slope of the regression line and increasing the intercept (Caissie, 2006; Erickson and Stefan, 2000; Webb and Nobilis, 1997). Stefan and Preud’homme (1993) showed that weekly and monthly mean temperatures correlated better than daily means. Furthermore, small head-water streams influenced by groundwater discharge may show lower correlation between air and stream temperature because of little influence of the atmosphere on the stream temperature (Erickson and Stefan, 2000). The method has limitations; (i) when air temperatures are higher than 25 °C due to increase in evaporative heat loss reducing increase in stream temperature with increased air temperature or (ii) when air temperatures fall below freezing point since stream temperatures may not fall below 0 °C due to release of latent heat from ice formation (Webb et al., 2003).

Previous investigations using linear regression analysis have been used mainly to predict stream temperatures and estimate future climatic impacts on streams (Caissie, 2006; Jeppesen and Iversen, 1987; Pedersen and Sand-Jensen, 2007). These studies speculated on the effects of groundwater discharge, but without any quantitative analyses. O’Driscoll and DeWalle (2006) used stream-air temperatures in several streams within a karst setting and showed that lower slope and higher intercept were governed by groundwater discharge. Due to very different discharge patterns within a karst setting (e.g. reaches either with focused discharge and losing conditions were related to conduits and fractures) they found the air-stream temperature relation to be a good indicator for groundwater affected areas. However, in other settings where groundwater discharge/recharge may be more diffuse large differences in the slopes and intercepts of a regression model cannot be expected. The method of analyzing stream-air temperature relations is simple but relative little interest has been seen in the literature. Our focus is therefore on identifying groundwater discharge areas to a small 15 km long head-water stream within a post-glacial fluvial setting using and expanding the linear regression approach by now also applying diel amplitude as an indicator of discharge. Furthermore, in addition to the common application of linear regression analysis based on stream temperature monitored at a single depth within the water column, we base it on stream temperatures measured at two depths; in the Streambed-Water Interface (SWI) and near the air/water interface. A dual-depth temperature monitoring to more accurately determine the streamflow frequency and durations in ephemeral channels was suggested by Constantz et al. (2001), but here a dual-depth approach is
used to identify groundwater affected areas. Thus, two quantitative metrics are used for both depths; (i) slope, intercept and coefficient of determination, $r^2$, of the regression between air/water temperatures also used by others (Caissie, 2006; Erickson and Stefan, 2000; Webb and Nobilis, 1997) and now also (ii) the average daily damping in temperature.

### 9.2 Methods

![Map of the study area](image)

Figure 9.1: The study area in the western part of Denmark with a topographical catchment of 126 km$^2$. The grey points depict the location of the data loggers along Holtum stream. The streamflow is from east to west.

Holtum stream in western Denmark is a lowland, perennial and gaining head-water stream (Fig. 9.1). At ten locations (in average $\sim 1.5$ km apart, Fig. 9.1) TidbiT v2 temperature data loggers, operating at an accuracy of $\pm 0.2^\circ$ C, were installed in the water column $\sim 10$ cm below the stream stage and right in the SWI. Data loggers were strapped to a dipstick and temperatures were logged every 15 min. from Sep. 2008 to Jun. 2010. Half-hourly air temperature data were obtained from a weather station within $\sim 6$ km from the investigation area (Ringgaard et al., 2011) assumed to be representative of the air temperature at the ten locations (Mohseni et al., 1998).
By installing an additional temperature logger in the SWI we compare and analyze results of the linear regressions of both the upper stream- and lower SWI temperatures versus air temperature. We use average daily mean temperatures from the two depths in evaluating two metrics; (i) parameters of the linear regression model (slope, intercept and $r^2$) and (ii) the average daily damping. Damping is here defined as the average diel amplitude and it is based on the hypothesis that groundwater discharge will lead to less fluctuations in stream temperature. The results are compared to other measures of discharge (temperature-based Darcy fluxes, synoptic measurements of stream discharge used to approximate the baseflow component as flow accretion ($m^3 s^{-1} km^{-1}$)). A Principal Component Analysis (PCA) was carried out to differentiate between locations with different groundwater influence using the basic statistical variables calculated from the recorded temperatures (mean, median, standard deviation, standard error, minimum, maximum, sample variance, mean diel amplitude, and range) and parameters measured in the field (flow accretion, vertical flux, differential head between stream and groundwater).

9.3 Results and Discussion

The suffix “u” and “l” denote “upper” and “lower” data logger in the water column and at the SWI, respectively. The number of daily mean temperature values at X1u, X1l and 1718l are much lower (34-56) than at the remaining locations (256-628) and thus omitted in the analysis. Fig. 9.2A shows the variation in daily mean temperatures at the remaining locations from seven consecutive months in 2009 (Jan. 17 to Aug. 18). Large fluctuations in daily mean temperatures due to seasonal effects are discernible in all data loggers except three; 1516l, 1314l and 0708l, all placed in the SWI. Differences in the daily mean temperatures within the interquartile range in the upper and lower data logger are clearly distinguished at the three locations showing more damping in the SWI. Here, the daily means within the interquartile range are clustered closer to 8 °C (equivalent to mean groundwater temperature in Denmark) indicating a persisting effect of higher groundwater discharge dampening the daily temperature differences in the SWI. In contrast, the lower data loggers at the remaining locations show larger variation and thus a higher interquartile range close to that of their corresponding upper data logger.

The pattern with areas showing possible lower and higher groundwater discharge is clearer when comparing the daily differences between March and April 2008 (Fig. 9.2B). The interquartile range for the air temperature changes from $\sim 3.5$ °C in March to $\sim 8.5-11.5$ °C in April with the median temperature increasing $\sim 5.5$ °C. At most
Figure 9.2: Box and whisker plots of daily mean temperatures. (A) Variation in daily mean temperatures for seven consecutive months. (B) Variation in mean daily temperatures for March and April 2008. The grey shaded and white boxes represent the month of March and April, respectively.

locations the increase in median temperature from March to April is $\sim 2 \, ^\circ C$, which shows the effect of a gaining stream. However, at the same locations as above (1516l, 1314l, and 0708l) there is likely a higher groundwater discharge as the median is only raised by $\sim 1 \, ^\circ C$ with data also showing a smaller variation in daily mean temperatures. So even though the temperature contrasts between air and groundwater (in March and April) are not the highest during a season it is still possible to identify groundwater-affected areas.

Fig. 9.3 shows regression lines for stream/SWI temperatures vs air temperatures. The coefficient of determination, $r^2$, are all highly significant ($P<0.0001$), except for 0506u ($P<0.05$). At four locations, 1901l, 1516l, 1314l and 0708l the slope deviates clearly from unity with the regression model explaining only 43-75 % of the variation. At the remaining locations the regression model explains more than 91 % of the variation (Fig. 9.3). At 1516l and 0708l there are two trends in the scatter plots for SWI-air temperatures around 8 $^\circ C$. The two trend lines are likely caused by streambed scouring altering the sedimentation around the data logger at the SWI. Sedimentation on top of the logger dampens the stream temperature appearing as groundwater signal and vice-versa for a data logger being removed relatively from the SWI due to scouring. Nevertheless, both of the trends in the lower data loggers differ notably from the slope.
in the upper data logger. At 0304 regression lines are highly correlated but showing clearly higher intercepts and lower maximum stream temperatures compared to other locations. Here, the stream is shallow and both loggers are close to each other and groundwater thus likely causing the effects (O’Driscoll and DeWalle, 2006).

Figure 9.3: Regression models for all locations based on daily temperature means. The coefficient of determination, $r^2$, for both upper and lower data loggers is specified for each location.

The slope and intercept of the regression lines are inversely related at 1901l, 1516l, 1314l and 0708l opposite to what is found at the remaining locations (Fig. 9.4A and 4B). Thus, lower slopes and higher intercept indicate higher groundwater discharge (O’Driscoll and DeWalle, 2006). Mean diel amplitudes $\leq 0.6$ °C are found at 1901l, 1314l and 0708l (Fig. 9.4C), where the regression models also show clear deviation from
a slope of unity. Therefore, such a threshold may indicate areas with high groundwater discharge. The linear regression model at 1516l identified this area as a discharge zone, but the mean diel amplitude is, however, slightly larger (1.0 °C) than this threshold.

Figure 9.4: Results of the mean slope, intercept, and diel amplitude and the 95 percent confidence interval at all locations for the Complete Data set (CD). The vertical bars show the confidence interval. The “xxd” refer to the number of randomly selected days of temperature data.

The spatial patterns in average diel amplitude in the lower data loggers are compared with mean flow accretions in Fig. 9.5 (six synchronous campaigns with 58 discharge measurements were carried out in the period Nov. 2008 to Nov. 2010; Karan et al. (in prep)). The flow accretions are synoptic measurements and the relative mag-
nitudes are not expected to match the relative magnitude of the mean diel amplitude exactly. The highest measured flow accretions of 0.08-0.24 m$^3$ s$^{-1}$ km$^{-1}$ occur at locations with diel amplitudes $\leq$ 0.6 °C. There is a pattern of increasing flow accretion with decreasing mean diel amplitude from the upstream- to the downstream part. Flow accretions $\geq$ 0.07 m$^3$ s$^{-1}$ km$^{-1}$ are found at four locations (1901l, 1516, 1314l, and 0708l), where the slope and intercept in the regression models indicate groundwater discharge. At these locations the increase in flow accretions and decrease in mean diel amplitudes (except in 1516l) are consistent with lower slopes and higher intercepts in the regression models. Likewise, at the remaining locations the decrease in flow accretions and increase in mean diel amplitudes are consistent with higher slopes and lower intercepts of the regression models. The diel amplitude at 1901l is notably lower compared to the trend line clearly depicting a strongly affected groundwater discharge area compared to the up- and downstream location (Fig. 9.5). The flow accretion at 0102l is lower than expected compared to the mean diel amplitude. We are unsure why this is but groundwater abstraction from the local water work (~ 500 m south of the location) could potentially lower the flow accretion. However, overall, low mean diel amplitudes, low slopes and large intercepts in the regression models agree well and identify areas with higher groundwater discharge.

Figure 9.5: Comparison of the mean diel amplitude in lower data loggers with flow accretions. The solid lines represents the trend in the data depicting an inverse relation.

At 1901l, 1314l, and 0708l, fluxes were estimated from measured temperature profiles and a steady-state solution to the 1D heat transport equation (Karan et al., in
Giving relatively high fluxes of 0.6-1.2 m d$^{-1}$. This agrees with the high flow accretions in the same areas. At 1516l the flux is lower (0.2 m d$^{-1}$) although there is a relatively high flow accretion and the regression model also indicate relatively high discharge. The flux is, however, higher than at the remaining locations (0102l, X4l, 1718l) with fluxes of 0.1-0.2 m d$^{-1}$. These lower fluxes agree with an observed increase in diel amplitudes and decrease in flow accretions (Fig. 9.5).

A Principal Component Analysis (PCA) was used to test the relationship of lower mean diel amplitude to higher groundwater discharge. The PCA was carried out using basic statistical variables calculated from the daily temperatures (mean, median, standard deviation, standard error, minimum, maximum, sample variance, mean diel amplitude and range) and parameters measured in the field (flow accretion, vertical flux, differential head between stream and groundwater). Based on the PCA loading plots (not shown), the sample variance and the range of the measured water temperature govern the separation of the upper and lower data logger. Also, the observed flow accretion and the calculated vertical flux are inversely correlated to the mean diel amplitude corresponding to our initial hypothesis.

In this catchment a threshold in diel amplitude around 0.6 $^\circ$C agrees with the parameters of the regression models indicating groundwater-affected areas. Depending on the climatic characteristics of other post-glacial fluvial settings with gaining reaches a similar threshold could be established. For losing reaches, the absence of significant temperature damping may not necessarily indicate recharge conditions. However, areas affected of flow reversals (e.g. due to new groundwater abstraction causing stream flow depletion), could potentially be detected with the dual-depth monitoring due to changes in temperature damping in the SWI temperature sensor analogous to streamflow-affected temperature changes in ephemeral streams (Constantz et al., 2001).

### 9.4 Practical aspects

Data shows that groundwater effects were discernible between two months, even at a time where the contrasts in temperature between groundwater and air is not the greatest during a season. Therefore, to test how long a data set would be needed to reproduce the observed pattern in the complete data set from each data logger we randomly selected daily means of stream/SWI temperature and the corresponding air temperature equivalent to a sampling period of 30, 60, 90, 120, 150, and 180 days. The slopes and intercepts in the regression models and the calculated mean diel amplitudes from the randomly selected data were compared to those based on the complete data set (Fig. 9.4). For both metrics, data sets between 30-90 days most frequently fell outside
the 95 percent confidence interval of the complete data set. For data set lengths \( \geq 120 \) days this is only the case for two out of the seventeen data loggers. Thus, a minimum data set of 120 days is recommended.

This study was carried out with relatively few resources; installing 20 data loggers along a 15 km reach took just one day, data loggers are in-expensive (around $2000), and the analysis is simple. This approach could prove cost-effective in identifying areas of interest complementing other experiments or designing new field investigations. Prior to implementation some considerations need to be taken into account. In a loose streambed, scouring/sedimentation could alter the placement of the lower data logger (probably affecting 1516l). Therefore, depending on the streambed characteristics loggers in the SWI may need to be adjusted during the monitoring period.

9.5 Conclusions

We show that by monitoring the stream temperature in both the stream and in the SWI, areas of increased groundwater discharge are identified with a simple analysis comprising of two metrics; (i) parameters in the linear regression model of stream water-air temperatures (slope, intercept and \( r^2 \)) and (ii) the mean diel amplitude. Spatial changes in flow accretions and temperature-based fluxes agree with those identified as high groundwater discharge areas from the two metrics using data from the SWI logger. Data from the SWI with increased groundwater discharge should therefore show: (1) decrease in slope of the regression line, (2) increase intercept of the regression line, (3) lower mean diel amplitudes, and (4) less variation in daily mean temperatures. Our analysis show that 120 days of monitoring in this catchment is sufficient to reproduce the observed patterns. Therefore, due to the simplicity of the proposed method (both in terms of implementation and analysis) this could be a first cost-effective approach to identify areas of significant groundwater discharge to small streams in a post-glacial fluvial setting.

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