

RESEARCH ARTICLE

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Key Points:

- Headwater CO₂ evasion strongly contribute to C loss from a small lake catchment
- The gas transfer velocity mainly regulates aquatic CO₂ evasion
- Spatial and temporal dynamics of C flux can be large at small spatial scales

Correspondence to:

J. Kokic,
jovana.kokic@ebc.uu.se

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Carbon dioxide evasion from headwater systems strongly contributes to the total export of carbon from a small boreal lake catchment

Jovana Kokic¹, Marcus B. Wallin¹, Hannah E. Chmiel¹, Blaize A. Denfeld¹, and Sebastian Sobek¹
¹Department of Ecology and Genetics/Limnology, Uppsala University, Uppsala, Sweden

Abstract Inland waters are hotspots for carbon (C) cycling and therefore important for landscape C budgets. Small streams and lakes are particularly important; however, quantifying C fluxes is difficult and has rarely been done for the entire aquatic continuum, composed of connected streams and lakes within the same catchment. We investigated carbon dioxide (CO₂) evasion and fluvial fluxes of dissolved inorganic carbon and dissolved organic carbon (DIC and DOC) in stream and lake systems within the 2.3 km² catchment of a small boreal lake. Our results show pronounced spatial and temporal variability in C fluxes even at a small spatial scale. C loss from the catchment through CO₂ evasion from headwaters for the total open water-sampling period was 9.7 g C m⁻² catchment, dominating the total catchment C loss (including CO₂ evasion, DIC, and DOC export from the lake, which were 2.7, 0.2, and 5.2 g C m⁻² catchment, respectively). Aquatic CO₂ evasion was dominated by headwater streams that occupy ~0.1% of the catchment but contributed 65% to the total aquatic CO₂ evasion from the catchment. The importance of streams was mainly an effect of the higher gas transfer velocities than compared to lakes (median, 67 and 2.2 cm h⁻¹, respectively). Accurately estimating the contribution of C fluxes from headwater streams, particularly the temporal and spatial dynamics in their gas transfer velocity, is key to landscape-scale C budgets. This study demonstrates that CO₂ evasion from headwaters can be the major pathway of C loss from boreal catchments, even at a small spatial scale.

1. Introduction

Inland waters have been termed as sentinels, integrators, and regulators of climate change emphasizing the role they play in global biogeochemical cycles [Williamson *et al.*, 2009]. The knowledge about their importance for the global carbon (C) cycle is well established, although to what extent is still under debate [Aufdenkampe *et al.*, 2011; Cole *et al.*, 2007; Tranvik *et al.*, 2009]. As a result of these research efforts inland waters have recently been considered in global C models [Intergovernmental Panel on Climate Change (IPCC), 2013]. Inland waters can modulate the overall carbon balance of entire landscapes, by not only acting both as important sources of C to the atmosphere through evasion of greenhouse gases, such as carbon dioxide (CO₂) and methane (CH₄), but also as sinks through C burial in sediments of lakes and reservoirs [Algesten *et al.*, 2004; Bastviken *et al.*, 2011; Molot and Dillon, 1996; Richey *et al.*, 2002]. In a recent estimate of global CO₂ evasion from inland waters Raymond *et al.* [2013] concluded that 2.1 Pg C yr⁻¹ is evaded from streams, rivers, lakes, and reservoirs, which is of similar magnitude as the estimated net uptake of ~2.6 Pg C yr⁻¹ by terrestrial ecosystems [IPCC, 2013]. In addition, by covering only 2% of the total surface of the Earth, inland waters also bury more C than the world's oceans per unit area [Dean and Gorham, 1998], recently estimated to 0.6 Pg C yr⁻¹ [Tranvik *et al.*, 2009].

CO₂ evasion from inland waters is driven by supersaturation of CO₂ in the water and controlled by the gas transfer velocity (*k*) [Cole *et al.*, 1994]. In lakes, the supersaturation of CO₂ is partly caused by mineralization of organic matter by microbial and photochemical processes. These processes outweigh the photosynthetic consumption and hence produce a net heterotrophic system [Pace *et al.*, 2004]. Particularly in boreal regions, net heterotrophy in lakes is subsidized by a large amount of terrestrial organic matter (humus) flushed into lakes [Sobek *et al.*, 2003]. In addition, the input of dissolved inorganic carbon (DIC) via groundwater and stream inflow have been shown to be important for the supersaturation of CO₂ in lakes [Dillon and Molot, 1997; McDonald *et al.*, 2013; Stets *et al.*, 2009; Striegl and Michmerhuizen, 1998].

Fluvial stream networks have been deemed important for CO₂ evasion as shown by *Butman and Raymond* [2011] for the U.S. and especially emphasized for small-order streams [*Koprivnjak et al.*, 2010; *Raymond et al.*, 2013]. *Wallin et al.* [2013] showed that evasion of CO₂ from streams exceeded the fluvial downstream transport of dissolved organic carbon (DOC) and dissolved inorganic carbon (DIC) from a 67 km² boreal catchment. For a boreal stream, *Öquist et al.* [2009] showed that stream water was highly supersaturated in CO₂ (on average 5–17 times the equilibrium with the atmosphere) as a result of terrestrially respired C entering via groundwater input. They found that most of this CO₂ was emitted to the atmosphere within 200 m downstream after entering the stream. Furthermore, CO₂ supersaturation in streams has also been connected to the wetland coverage in the catchment [*Dawson et al.*, 2002; *Wallin et al.*, 2014, 2010]. A recent study by *Peter et al.* [2014] has also shown that in-stream respiration contributes to the observed supersaturation.

Traditionally, lakes and streams have been studied separately, partly due to the lack of previous integrative research efforts but also since the drivers for CO₂ supersaturation and the physical gas exchange with the atmosphere are often different between the systems. The lack of connectivity between studies hampers a more integrative understanding of C dynamics in inland waters at landscape scales. Focus has in the past been on lakes since they cover a larger surface area [*Algesten et al.*, 2004], but recent studies have provided evidence that streams have a larger impact on a landscape scale [*Campeau et al.*, 2014; *Crawford et al.*, 2014; *Huotari et al.*, 2013; *Lundin et al.*, 2013]. *Aufdenkampe et al.* [2011], *Humborg et al.* [2010], and *Buffam et al.* [2011] have presented exemplary studies of such integrations of inland waters, but all these estimates, with few exceptions [*Juutinen et al.*, 2013] operate at quite large spatial scales, both global and regional. However, both the dynamics and magnitudes of C fluxes can be large within a system, as shown by *Kortelainen et al.* [2013] for small lakes and *Teodoru et al.* [2009] for small streams. Thus, average estimates for the smallest aquatic systems remain difficult to accomplish which introduces high uncertainty when extrapolating to larger scales. Accordingly, it has been proposed by *Cole et al.* [2007] that the smallest low-order streams are difficult to include in global estimates of CO₂ evasion as the knowledge of their global area and *k* is scarce. To overcome these shortcomings, integrative and comparative studies investigating C dynamics, especially CO₂ evasion, for both lakes and streams at small spatial scales need to be conducted.

In this study we investigated spatial and temporal patterns in aquatic C concentrations and fluxes within a small boreal catchment (226 ha) in Sweden. We hypothesize that (1) aquatic C concentrations and fluxes within the catchment vary significantly in space and time, (2) that streams dominate the total aquatic CO₂ evasion of the catchment, and (3) that the C loss from a small boreal lake is smaller than the CO₂ evasion from the headwater systems in its catchment.

2. Methods

2.1. Study and Site Description

The study was conducted in the 226 ha catchment of Lake Gäddtjärn, situated in the southern boreal region of Sweden (59.86°N, 15.18°E) (Figure 1). The lake itself has a surface area of 6.8 ha and a mean depth of 3.4 m. The elevation range in the catchment is 254 to 340 m above sea level. The catchment is mainly forested (84%) with Norway spruce (*Picea abies*) and Scots pine (*Pinus sylvestris*) as dominating tree species. The remaining catchment is covered by wetlands (14%) and surface waters (4%). The area is typical for this region of Sweden with high density of aquatic systems that emanate from the last glaciation period some 9800 to 9600 years ago [*Lundqvist*, 1986]. Much of the stream network has been man-made ditched during the last 100–150 years to improve forest productivity.

Annual average temperature is 4.5°C, and annual average precipitation is 900 mm for the years 1961–1990 (data obtained from the Swedish Meteorological and Hydrological Institute, SMHI, www.smhi.se). The catchment has three very small headwater lakes (Kringeltjärn 0.86 ha, Prästtjärn 0.73 ha, and Svintjärn 0.33 ha) situated on higher altitudes (288–330 m above sea level) that drain into the main inlet stream (GD inlet 1) into Lake Gäddtjärn. Lake Gäddtjärn also has a second smaller inlet stream (GD inlet 2) that drains a wetland, and one outlet (GD outlet) that further drains into a larger lake, Lake Kölsjön (not part of this study). Sample sites (*n* = 9) are illustrated in Figure 1 and were chosen to represent the variety of surface water ecosystems present in the catchment.

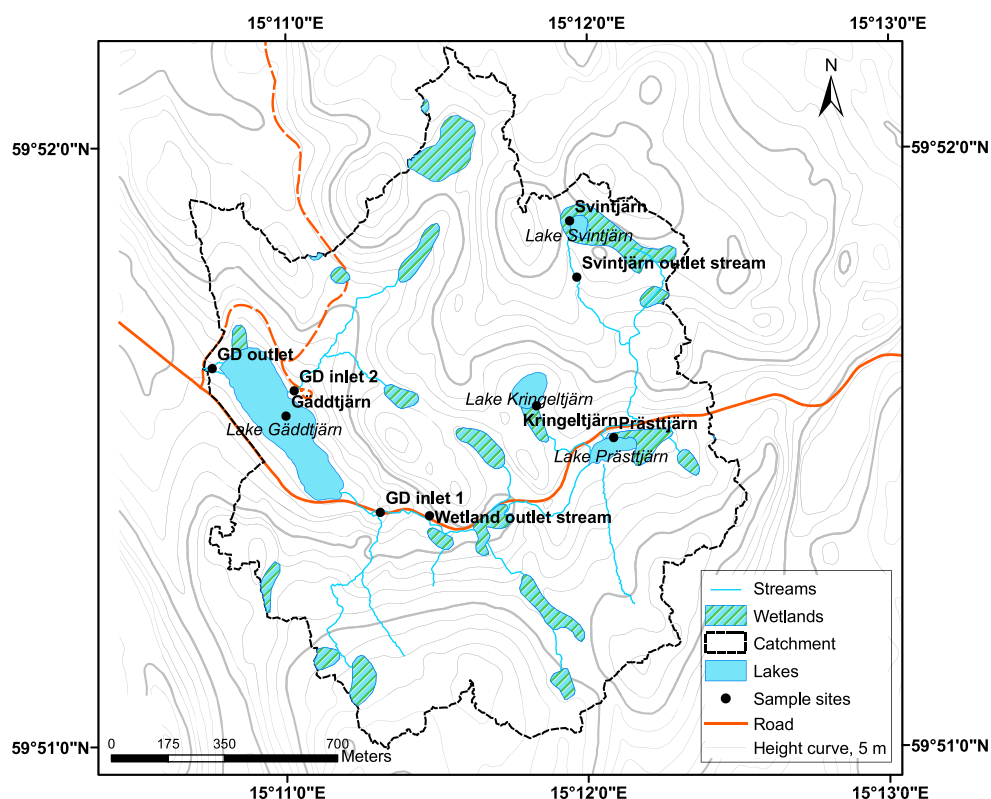


Figure 1. Map over the catchment of Lake Gäddtjärn and sample sites copyright Lantmäteriet, i2012/921.

2.2. Measurement of Water Chemistry Parameters and C

Manual sampling was conducted approximately once a month during open water periods from June 2011 to November 2012. The ice-covered period for Lake Gäddtjärn was approximately from December 2011 to mid-April 2012. Not all sites were sampled at every occasion due to logistical reasons. Water samples were refrigerated dark instantaneously upon sampling. DOC concentration was analyzed on water filtered directly after sampling through precombusted GF/F glass fiber filters (0.7 μm , Whatman), stored dark and cool, and analyzed within a week using a Total Carbon (TC) Analyzer (Sievers 900) equipped with a membrane-based conductivity detector. Particulate organic carbon (POC) was analyzed using a known volume of water filtered through a GF/F filter and combusting dried filters with an elemental analyzer (ECS 4010 Elemental Combustion System, CHNS-O) equipped with a thermal conductivity detector. Samples for DIC concentration were drawn into 60 mL polypropylene syringes, transferred to headspace-free glass vials without bubbling, and immediately stored dark and cool. DIC samples were analyzed within 2 days after sampling on a TC analyzer (Sievers 900). In rare cases when DIC samples had to be stored for up to 1 week after sampling, samples were acidified with 10% hydrochloric acid (HCl) solution directly upon sampling and stored in headspace-free vials.

The partial pressure of CO_2 ($p\text{CO}_2$) was analyzed using the headspace equilibration method [Sobek *et al.*, 2003], where samples were drawn into 60 mL polypropylene syringes and sealed with three-way stopcocks after removing any bubbles. A known ratio of sample water and ambient air was equilibrated in the syringe by shaking it vigorously for 1 min. The equilibrated air was extracted into another syringe and analyzed on a portable infrared gas analyzer (Infra Red Gas Analyser (IRGA), EGM-3, and 4 PP Systems). The volume ratio of water to ambient air was adjusted to get a $p\text{CO}_2$ reading within the range of the IRGA (usually 20/40 mL water/ambient air). The $p\text{CO}_2$ was calculated according to Weiss [1974] using the appropriate Henry's constant after correcting for temperature and atmospheric pressure as well as the amount of CO_2 added by the ambient air. The $p\text{CO}_2$ of ambient air was measured several times at each sampling occasion (mean 405 μatm , range 368–459, $n = 115$). Temperature, electrical conductivity (EC),

dissolved oxygen (DO), and pH were determined in situ by an HQ40d Portable Multiparameter Meter (HACH) and an 826 portable pH meter (Metrohm).

2.3. Continuous Measurement of Water Level and CO₂ With In Situ Sensors

Stream CO₂ concentration, water level, and temperature at GD inlet 1 and GD outlet of Lake Gäddejärn (Figure 1) were continuously monitored over a period of approximately one open water season, 31 August 2011 to 4 September 2012 (for the inlet; until 8 November 2012 for the outlet). CO₂ was measured according to *Johnson et al.* [2010] with an Nondispersive infrared sensor (GMT220, Vaisala) covered with a polytetrafluoroethylene membrane (International Polymer Engineering) that is impermeable to water but permeable to CO₂. A pressure transducer (MJK Expert 3400) and a temperature sensor (TOJO Skogsteknik Soil, TO3R) were also deployed. All sensors were connected to a data logger (Campbell Scientific CR1000) that recorded data hourly. The concentration of CO₂-C was calculated by using the appropriate Henry's constant corrected for temperature and atmospheric pressure [*Tang et al.*, 2003]. Hourly atmospheric pressure was obtained from the nearby meteorological station Kloten A (SMHI, 59.52°N, 15.15°E), and used for the sample sites of this study with correction for differences in altitude by the barometric formula [*Holton*, 2004].

Stream discharge (*Q*) was measured at each sampling occasion by the salt (NaCl) dilution method [*Day*, 1975, 1976]. Electrical conductivity (EC) was measured ~15 m downstream of the site of salt injection using an HQ40d Portable Multiparameter Meter (HACH). Based on the relationship of water level (obtained by the pressure transducer) and the measured discharge (power fit regression, GD inlet 1 $n = 7$, $R^2 = 0.92$, GD outlet $n = 10$, $R^2 = 0.88$) hourly discharge data were obtained from hourly logged water level.

2.4. Determination of Gas Transfer Coefficient in Streams

The gas transfer coefficient (k_{TC}) in streams was determined using a volatile gas tracer, propane (C₃H₈), previously used in similar type of studies [e.g., *Wallin et al.*, 2011; *Billett and Harvey*, 2013; *Genereux and Hemond*, 1990; *Öquist et al.*, 2009]. The gas tracer injections were performed during 2012 on three stream reaches (at four occasions for GD inlet 1 and GD inlet 2 and at two occasions for Wetland outlet stream, see Figure 1); in total, 10 tracer gas injections were made at various discharge conditions. Stream reaches were between 25 and 50 m in length and generally 0.5 to 1 m wide. The selected stream reaches covered sections with both pools and riffles to capture the morphological heterogeneity of the reaches and were representative for streams in the area. Prior to injecting the propane, *Q* and reach travel time (τ) were determined by the salt dilution method. EC was measured at each end of the reach using a duplicate setup of CS547A probes connected to a CR510 data logger (both probes and logger, Campbell Scientific). The τ was measured as the time difference between the EC peaks at the upper and lower reach ends. C₃H₈ (10 kg cylinder, Aga gas, Sweden) was injected to the stream through an air curtain 10–15 m upstream of the upper reach end at a constant rate approximately 10–15 min prior to sampling to achieve steady state within the stream reach. Bubble-free stream water samples (three replicates) were taken at upper and lower reach ends in 60 mL polypropylene syringes sealed with three-way stopcocks. The time between upper and lower reach sampling was set equivalent to τ , to sample the same water mass. A known and equal volume of headspace (ambient air) was introduced to all samples and equilibrated with the water mass by vigorous shaking for 1 min. The headspace was transferred to a ~25 mL glass vial, which was prefilled with saturated salt solution (NaCl) and sealed with 20 mm thick butyl rubber septa (Apodan Nordic). The vials were stored upside down until analysis to prevent any gas leakage through the rubber septa. C₃H₈ samples were analyzed on a gas chromatograph (7890A GC system, Agilent Technologies). In addition, *p*CO₂ was measured at upper and lower reach ends prior to each gas tracer injection according to the method described above.

2.5. Calculations of Lake and Stream CO₂ Evasion

Lake CO₂ evasion (CO_{2Ev}) to the atmosphere was calculated by the equation

$$CO_{2Ev} = k_{CO_2} \times (CO_{2w} - CO_{2sat}) \quad (1)$$

where CO_{2w} is the concentration of CO₂ in the surface water, CO_{2sat} is the concentration of CO₂ in the water if it would have been in equilibrium with the atmosphere (saturation concentration) [*Cole and Caraco*, 1998],

and k_{CO_2} is the gas transfer velocity (cm h^{-1}) estimated using temperature-dependent Schmidt numbers (Sc) for CO_2 (600 at 20°C) according to *Jähne et al.* [1987]:

$$k_{\text{CO}_2} = k_{600} \div \left(\frac{600}{Sc_{\text{CO}_2}} \right)^{-0.66} \quad (2)$$

where k_{600} is the gas transfer velocity (cm h^{-1}) normalized to a Sc for CO_2 at 20°C and -0.66 is a coefficient describing the characteristics of the water surface assuming lower wind speeds ($<3.5 \text{ m s}^{-1}$).

Sc_{CO_2} is the Sc prior to normalization of Sc to 600 and is dependent on water temperature:

$$Sc_{\text{CO}_2} = 1841 \times \exp^{-0.0549T} \quad (3)$$

where T is water temperature in $^\circ\text{C}$. The k_{600} is mainly controlled by wind speed in lakes and was calculated according to *Cole and Caraco* [1998]:

$$k_{600} = 2.07 + 0.215U_{10}^{1.7} \quad (4)$$

where U_{10} is the wind speed in m s^{-1} at 10 m above surface water. This model relates U_{10} to the mass loss of tracer gas that was added to a small, wind-sheltered lake and therefore integrates vertical and horizontal variability in wind speed and includes spatial variability in k [*Schilder et al.*, 2013]. It is therefore well suited to estimate whole-lake CO_2 evasion from the small and wind-sheltered lakes in our study. For this study we used wind speed data (U_{10} , daily averages) from the meteorological station Kloten A (mean value 1.75 m s^{-1} , range $1.0\text{--}3.8 \text{ m s}^{-1}$).

Stream CO_2 evasion was calculated according to equation (1) with the gas transfer velocity k_{CO_2} (cm h^{-1}) derived from the measurements of the gas transfer coefficient k_{TC} (min^{-1}) by the following steps:

The gas transfer coefficient for C_3H_8 ($k_{\text{C}_3\text{H}_8}$) was calculated according to the following equation by *Genereux and Hemond* [1990] and modified by *Wallin et al.* [2011]:

$$k_{\text{C}_3\text{H}_8} = \frac{1}{\tau} \times \ln \left(\frac{[\text{C}_3\text{H}_8]_U}{[\text{C}_3\text{H}_8]_L} \times \Delta A \right) \quad (5)$$

where $[\text{C}_3\text{H}_8]_U$ and $[\text{C}_3\text{H}_8]_L$ are the relative concentrations of C_3H_8 at the upper and lower reach ends and ΔA is the change in contributing catchment area between the upper and lower reach ends to compensate for any potential dilution by groundwater input along the stream reach (assuming same specific runoff). The $k_{\text{C}_3\text{H}_8}$ was converted to k_{TC} according to *Jones and Mulholland* [1998]:

$$k_{\text{TC}} = k_{\text{C}_3\text{H}_8} \times \left(\frac{d_{\text{C}_3\text{H}_8}}{d_{\text{CO}_2}} \right)^n \quad (6)$$

where n is set to 0.5 according to *Wallin et al.* [2011]; $d_{\text{C}_3\text{H}_8}$ and d_{CO_2} are the respective gas diffusion coefficients based on *Jähne et al.* [1987] and *Wise and Houghton* [1966]:

$$d_{\text{CO}_2} = 0.9477 \times \exp^{(0.0274T)} \quad (7)$$

$$d_{\text{C}_3\text{H}_8} = 1.092 \times \exp^{(0.0235T)} \quad (8)$$

k_{TC} was multiplied by the average stream depth z (cm) in order to determine the transfer velocity k_{CO_2} [*Wanninkhof et al.*, 1990]:

$$k_{\text{CO}_2} = z \times (k_{\text{TC}} \times 60) \quad (9)$$

All evasion rates were calculated in g C m^{-2} (water surface area) d^{-1} . To investigate temporal patterns of CO_2 evasion, CO_2 concentration from the continuous measurements of CO_2 at GD inlet 1 were combined with a modeled k_{CO_2} , based on a linear relationship between the measured k_{CO_2} and Q ($n = 10$, $R^2 = 0.83$; data not shown). To calculate CO_2 evasion from lakes and streams for the total open water-sampling period, we interpolated the evasion between the different sampling points when we measured $p\text{CO}_2$, to capture the seasonal variability, and summed over the entire open water period as $\text{g C m}^{-2} \text{yr}^{-1}$. Lake CO_2 evasion includes measuring points before, during, and after lake turnover. For streams we interpolated accordingly. Uncertainty estimates of CO_2 evasion were calculated as cumulative standard error of tracer injection replicates (streams), wind speed model-derived k_{CO_2} (lakes), and concentration measurements for CO_2 evasion.

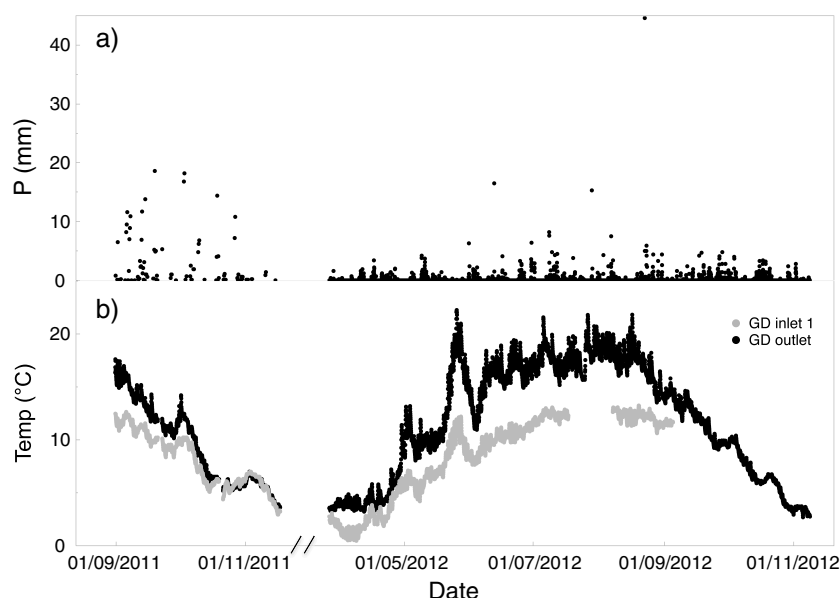


Figure 2. Time series of hourly (a) precipitation, P , for the nearby meteorological station Kloten A and (b) water temperature for GD inlet 1 and GD outlet.

2.6. Quantifying Fluvial Flux of C

To quantify the fluvial C flux, we used a regression model between Q and C load, where DOC and DIC concentrations were multiplied by Q measured at the same time point, to obtain a C load in g h^{-1} . We used this approach because there was no apparent relationship between DIC or DOC concentration and Q . The regression between C load and Q measured at the same time point ($n = 10$, $R^2 = 0.95$ and 0.96 for GD inlet 1 DOC and DIC; $R^2 = 0.96$ and 0.88 for outlet DOC and DIC; data not shown) was used to estimate fluxes from continuously measured hourly Q at the sites GD inlet 1 and GD outlet. The hourly loads were summed to obtain a total flux (in tons C) during the open water season, when Q was logged hourly. Uncertainty estimates of fluvial DIC and DOC flux were calculated as cumulative standard error of discharge measurements and concentration measurements of DOC and DIC. A flux for POC was not calculated because POC only accounted for 4% of the total organic carbon pool in the streams. Furthermore, during our sampling year, local authorities unexpectedly recommenced a previously ceased liming program. On 17 July Lake Gäddejärn was limed with 4.59 t CaCO_3 (corresponding to 0.55 t C) that was directly put onto the lake by airplane (information obtained from the Swedish county administrative board, www.kalkdatabased.lansstyrelsen.se) during summer stratification. Liming caused an increase in pH from ~ 6.1 to ~ 9.1 , followed by a drop to ~ 7.1 within 5 days, and a steady decrease back to the initial value within less than 2 months; a pH sensor deployed in the hypolimnion showed no response to the liming (H. E. Chmiel, manuscript in preparation). Together with the absence of inorganic carbon in lake sediment, these observations indicate that the lime completely dissolved in the epilimnion and was removed from the lake within ~ 2 months, which corresponds to the water retention time of the epilimnion (< 6 weeks). Due to the elevated pH after liming, the majority (60–85%) of the DIC was present in the form of dissolved carbonate and bicarbonate, and thus, the DIC added through lime was mainly flushed from the lake via the outlet, in particular since elevated $p\text{CO}_2$ in the lake outlet after liming indicates that also gaseous CO_2 was flushed from the lake via the outlet (Figure 3b). Hence, we subtracted the amount of C added by the liming from the measured DIC flux at GD outlet. While this procedure does not account for the small fraction of lime-derived C that left the lake via CO_2 evasion to the atmosphere, it should be noted that this does not affect the overall estimate of the total C loss from the lake.

Fluvial fluxes of C for the second, smaller inlet stream (GD inlet 2) were determined from manual measurements, as automated logging of water level was not installed at this site. For interpolation between measurements at GD inlet 2, we applied relationships between manual samplings and hourly fluxes of GD

Table 1. Water Chemistry Parameters and Concentrations of Different C Species for the Catchment Surface Waters^a

	<i>n</i>	Temperature (°C)	Conductivity (μS cm ⁻¹)	pH	DO (%)	DO (mg L ⁻¹)	pCO ₂ (μatm)	DOC (mg L ⁻¹)	DIC (mg L ⁻¹)	POC (mg L ⁻¹)	Q (L s ⁻¹)
Svintjärn	2	16.5 (11.6–21.3)	25.1 (16.6–33.6)	4.5 (4.4–4.5)	71.8 (56.5–87)	5.5 (3.8–7.3)	2772 (1524–4019)	19.5 (15.6–23.3)	1.8 (0.49–3.2)	1.9 (1.8–2.0)	NA
Svintjärn outlet stream	8	10.6 (2.8–13)	34.1 (25.3–54.6)	4.1 (3.8–5.4)	79.4 (72.3–95.5)	8.6 (5.2–11.9)	3644 (2576–6401)	27.1 (16.4–60.2)	2.4 (1.7–3.1)	1.5 (0.02–4.5)	ND
Kringeltjärn	6	17.4 (12.3–25.1)	17.7 (14.0–34)	4.9 (4.5–5.0)	83.5 (73–101)	8.0 (4.8–9.1)	1356 (689–3943)	9.1 (7.5–16.2)	0.76 (0.45–2.1)	1.2 (0.72–2.9)	NA
Prästjärn	6	16.7 (11–25.6)	20 (18.0–47.1)	4.7 (4.0–5.1)	72.5 (66–89.6)	6.8 (4.5–8.5)	2928 (1470–6458)	14.0 (11.8–27.3)	1.6 (0.71–3.4)	0.83 (0.47–1.6)	NA
Wetland outlet stream	11	11.5 (3.5–14.5)	22.5 (17.0–43.8)	4.7 (4.0–5.6)	81.7 (61.7–98.2)	8.8 (5.6–11.9)	3286 (1798–4994)	19.8 (13–25.1)	2.0 (1.3–3.7)	0.23 (0.03–0.73)	ND
GD inlet 1	14	11.3 (2.5–14)	24.1 (16.0–71.8)	4.4 (3.3–5.9)	93.7 (62–96.6)	10.0 (6.2–12.6)	1819 (1047–2650)	19.2 (12.5–24.2)	1.0 (0.74–1.5)	0.43 (0.003–1.1)	32.0 (8.5–79.2)
GD inlet 2	14	10.4 (2.1–13.1)	24.5 (16.0–43.3)	4.4 (3.9–5.6)	95.6 (65.3–99.1)	10.4 (6.6–12.9)	1353 (920–2815)	24.4 (13.5–33.8)	0.75 (0.67–1.9)	0.53 (0.19–1.7)	8.84 (2.5–52.8)
Gäddtjärn	8	11.5 (2.9–18.5)	24.2 (19.0–27.0)	6.0 (5.2–6.5)	88.4 (80.9–98.0)	9.31 (8.43–11.2)	1669 (1423–2332)	17.1 (12.1–20.1)	1.5 (0.99–2.1)	ND	NA
GD outlet	14	15.8 (2.8–24.2)	27.2 (19.3–37.0)	6.1 (4.5–7.1)	85.7 (64.9–98.6)	8.58 (5.79–11.4)	1773 (1212–2471)	15.4 (10.1–19.6)	1.8 (1.06–3.1)	0.83 (0.37–1.78)	54.5 (13.8–194)

^aValues are reported as median (min–max). NA = not applicable; ND = no data.

inlet 1 as follows: Average fluxes for site GD inlet 1 in g h⁻¹ were determined for the time period *T* between two manual sampling occasions, e.g., 31 August 2011 to 20 September 2011, and compared to the flux obtained solely from the measured time point MT, e.g., 31 August 2011 in g h⁻¹. The difference was used to calculate a representative flux for that time period for GD inlet 2 with fluxes in g h⁻¹ calculated from the manual sampling occasions according to following:

$$\text{Flux } C_{\text{GD inlet 2}} = \left(\frac{\overline{T_{\text{GD inlet 1}}}}{MT_{\text{GD inlet 1}}} \right) \times MT_{\text{GD inlet 2}} \times d \quad (10)$$

where *d* is number of days for *T* and flux *C* is in mass unit tones.

2.7. Estimating Surface Water and Catchment Areas

With the use of the Hydrological Extension in ESRI® ArcGIS™ version 10.0 software, data from Översiktskartan and a 2 m digital elevation model (DEM) (copyright Lantmäteriet, i2012/921) lake areas, the total catchment area of Lake Gäddtjärn, upstream drainage areas, and the length of stream networks were determined. The 2 m DEM provided a sufficient resolution to accurately capture the small-order streams present in the catchment of Lake Gäddtjärn, as verified in the field. To calculate total stream area (m²), we used 140 measurements of stream width across the stream network of the Gäddtjärn catchment to divide the stream reaches where tracer injections were not performed into two categories; stream reaches draining lakes and stream reaches draining wetlands, which generally were less wide than the streams draining lakes. As the spatial variability of stream width detected in the 140 measurements was not different from the overall variability of width in the studied stream reaches, we used the average width of the GD inlet 1 reach (covering four different discharge regimes as well as within-reach spatial variability) to calculate stream area

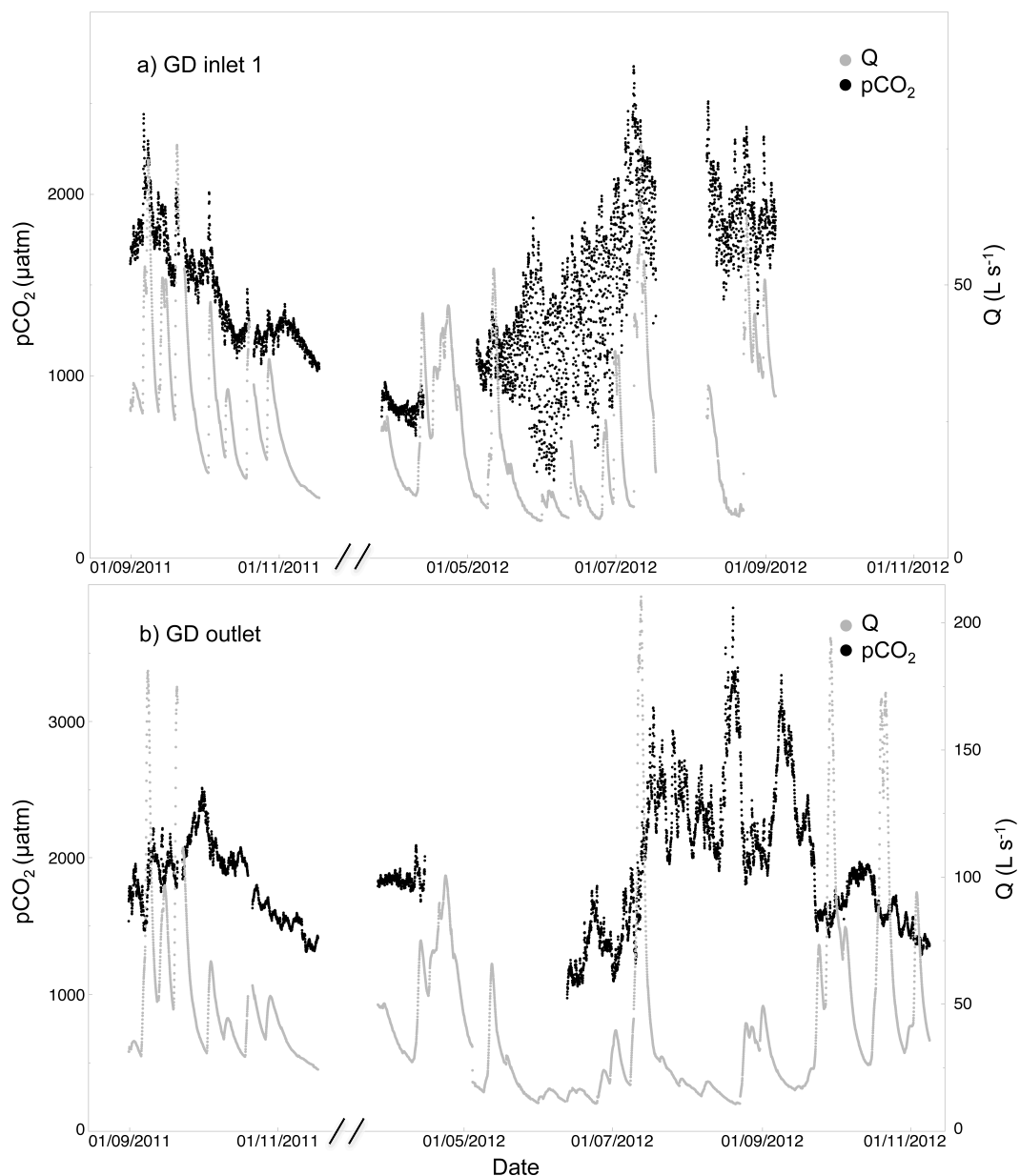


Figure 3. Time series of hourly discharge and $p\text{CO}_2$ for (a) GD inlet 1 and (b) GD outlet.

for the streams draining lakes; for the streams draining wetlands, we conservatively used half of width from the streams draining lakes to calculate total stream areas.

The analysis of the 2 m DEM resulted in a total catchment area for Lake Gäddtjärn of 226 ha, and drainage areas of 137 ha for GD inlet 1 and 45 ha for GD inlet 2. The total stream area was calculated to 0.23 ha corresponding to about 0.1% of the total catchment area.

3. Results

3.1. Hydrology and General Chemistry

Discharge varied between 6.8 and 75.7 L s^{-1} at GD inlet 1, between 10.7 and 210 L s^{-1} at GD outlet (3a and 3b), and between 2.5 and 52.8 L s^{-1} at GD inlet 2. In general, discharge at GD inlet 1 and GD outlet followed similar temporal dynamics, responding quickly to high-precipitation events (Figure 2a) followed by low-discharge conditions during dry and warm periods.

Table 2. Physical Properties, k_{CO_2} , $p\text{CO}_2$, and Median CO_2 Evasion for Lake Gäddejtjärn, Headwater Lakes, and Headwater Streams^a

System	Length (m)	Mean Width ^b (m)	Mean Depth (m)	Mean Slope (%)	Surface Water Area (km ²)	% Coverage of Catchment	k_{CO_2} (cm h ⁻¹)	$p\text{CO}_2$ (μatm)	CO_2 Evasion (g C m ⁻² d ⁻¹)
Gäddejtjärn	NA	NA	3.8	ND	67.5	3.0	2.1 (1.5–3.6)	1809 (1423–2332)	0.4 (0.3–0.8)
Headwater lakes	NA	NA	ND	ND	19.2	0.8	2.3 (1.9–2.8)	2673 (689–6458)	0.6 (0.08–1.5)
Headwater streams	5790	0.43 (0.05–1.2)	0.31 (0.12–0.60)	4.7 (2.0–8.5)	2.3	0.1	66.9 (35.2–301)	2266 (920–6401)	11.8 (4.9–64.3)

^aValues are reported as mean (min-max).

^bBased on 1 day measurements of spatial variability over 140 points in different stream reaches in the catchment in 2014. NA = not applicable; ND = no data.

Median water temperature for all sites over the study period varied from 10.4 to 17.4°C with lowest temperatures measured in the streams (Table 1) corresponding well to the variability presented in Figure 2b. Conductivity was low and relatively homogenous across sites and varied between 17.7 and 34.1 μS cm⁻¹. Median pH was low in the headwaters (4.1–4.9) with the lowest value measured as 3.3 in GD inlet 1. The median pH for Lake Gäddejtjärn and the GD outlet were higher, 6.0 and 6.1, respectively. All sites were oxygen rich with median DO varying from 8.0 to 10.4 mg L⁻¹, except Lake Prästtjärn and Svintjärn that contained less oxygen (median DO of 6.8 and 5.5 mg L⁻¹, respectively).

3.2. Spatial and Temporal Variability in C Concentrations

There was pronounced spatial variability in DOC concentrations in the catchment with median values ranging from 9.1 to 27.1 mg L⁻¹ between the sites. The highest concentrations were displayed in streams with close connection to wetland areas, such as Svintjärn outlet stream and Wetland outlet stream (Table 1). A similar pattern in spatial variability was shown for inorganic carbon where median values for $p\text{CO}_2$ ranged from 1353 to 3644 μatm and for DIC from 0.76 to 1.9 mg L⁻¹. The highest $p\text{CO}_2$ values and DIC concentrations were measured in Lake Prästtjärn and Svintjärn outlet stream. Within the same stream network, we observed a consistent $p\text{CO}_2$ decrease by about half from the upstream site Wetland outlet stream to the downstream site GD inlet 1 over a distance of ~150 m.

High temporal variability in DOC was observed at the Svintjärn outlet stream where the concentration varied from 16.4 to 60.2 mg L⁻¹ between the different sampling occasions. The temporal variability in $p\text{CO}_2$ was most evidently shown in Lake Prästtjärn and Svintjärn outlet stream, varying between the sampling occasions from 689 to 3943 and from 2576 to 6401 μatm, respectively. To further illustrate temporal variability, Figures 3a and 3b show different $p\text{CO}_2$ patterns in the streams before and after Lake Gäddejtjärn as measured by the continuous logger data at GD inlet 1 and GD outlet. At GD inlet 1, $p\text{CO}_2$ displayed a daily signal during low flows (~May to July 2012) and followed the pattern of discharge fluctuations. Seasonal variation was more evident at GD outlet from approximately July to November 2012.

3.3. CO_2 Evasion From Lakes and Streams

The median CO_2 evasion from Lake Gäddejtjärn was 0.4 g C m⁻² d⁻¹ (range 0.3–0.8) and was comparable to evasion from the headwater lakes with an estimated median evasion of 0.6 g C m⁻² d⁻¹ (range 0.08–1.5) (Table 2, all fluxes standardized to water surface area). These evasion estimates were based on median k_{CO_2} values of 2.1 cm h⁻¹ for Lake Gäddejtjärn (range, 1.5–3.6) and 2.3 cm h⁻¹ for the headwater

Table 3. Measured Length, Mean Width, Depth, Slope, and Q of the Stream Reaches During Tracer Injections^a

Stream Reach	Length (m)	Width (cm)	Depth (cm)	Slope (%)	Q (L s ⁻¹)
GD inlet 1	47	67 (32–118)	24 (11–39)	5.9	33.1 (10.8–79.2)
GD inlet 2	25	59 (24–126)	40 (14–72)	5.1	12.3 (4.0–28.2)
Wetland outlet stream	44	65 (12–121)	28 (10–69)	4.0	15.5 (9.8–21.3)

^aMinimum and maximum measured values are given in parenthesis.

Table 4. Fluvial Fluxes of DOC and DIC, Expressed as Total Mass for the Open Water-Sampling Period^a

Site	DOC (t)	DIC (t)
GD inlet 1	9.6 (8.0–12)	0.49 (0.42–0.58)
GD inlet 2	3.4 (2.8–4.1)	0.005 (0.0043–0.0059)
Total, inlet streams	13.0 (10.8–15.6)	0.50 (0.43–0.59)
Outlet	13.9 (11.5–16.7)	0.60 (0.43–0.74)

^aValues in parenthesis represent the minimum and maximum estimate obtained from cumulative standard error of discharge and concentration measurements for DOC and DIC.

lakes (range, 1.9–2.8), as derived from wind speed-modeled data. CO₂ evasion from the headwater streams was substantially higher than evasion from the lakes. Stream evasion was estimated at a median of 11.8 g C m⁻² d⁻¹ (range, 4.9–64.3) based on a median k_{CO_2} of 66.9 cm h⁻¹ (range, 35.2–301). Both the k values and CO₂ evasion in the headwater systems (streams and lakes) showed a high variability between sites, and over time. Table 3 shows the mean and variability of physical characteristics of the tracer injection measurements for

the stream reaches. By interpolating between sampling dates (see section 2) for the total open water-sampling period CO₂ evasion from Lake Gäddtjärn was 2.7 g C m⁻² of catchment, and 8.3 and 1.4 g C m⁻² of upstream catchment area for headwater streams and headwater lakes, respectively.

Headwater streams, the largest source of CO₂ evasion from the Lake Gäddtjärn catchment, showed a distinct seasonal pattern in evasion rate (Figure 4). The magnitude and variability in CO₂ evasion were largest in late summer and autumn and considerably lower during late autumn, spring, and early summer. The periods with low evasion rates were coinciding with periods of low discharge (Figure 2b).

3.4. Catchment-Scale C Fluxes

Lake Gäddtjärn received a similar amount of DOC through its inlets (13.0 t) than was exported through the outlet (13.9 t) (Table 4). Also, fluxes of DIC to the lake via inlet streams were similar (0.5 t) compared to DIC export via the lake outlet (0.6 t). CO₂ evasion from the streams surpassed the fluvial flux of DIC at the inlets by over 1 order of magnitude (Figure 5, DIC 0.3 g C m⁻² for GD inlet 1 and 0.01 g C m⁻² for GD inlet 2). The fluvial flux of DOC was greater than CO₂ evasion from both GD inlet 1 and GD inlet 2 (DOC 6.2 g C m⁻² and CO₂ evasion 4.96 g C m⁻² for GD inlet 1 and DOC 6.7 g C m⁻² and CO₂ evasion and 2.86 g C m⁻² for GD inlet 2). Comparing C loss pathways in the entire catchment, including Lake Gäddtjärn, Figure 6 illustrates that CO₂ evasion from the

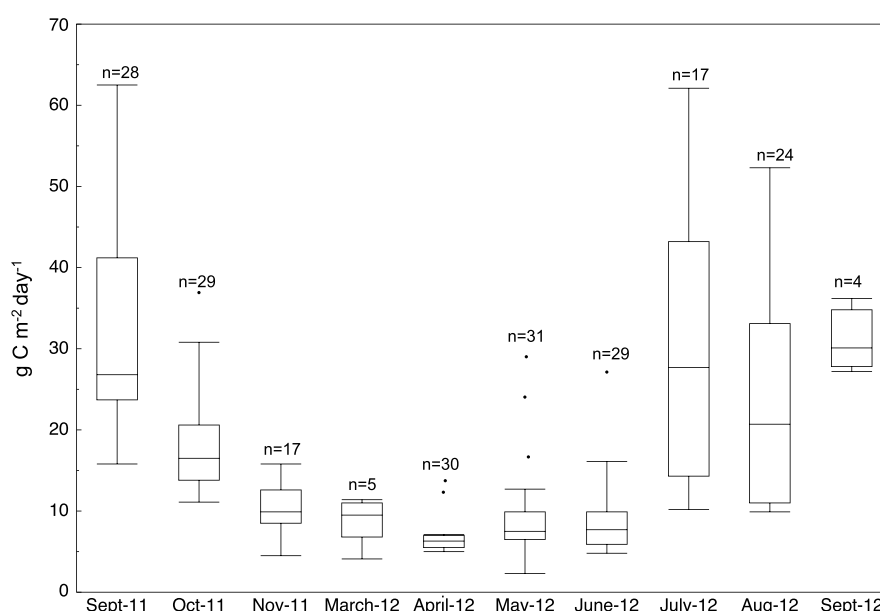


Figure 4. Temporal variability in CO₂ evasion for site GD inlet 1, estimated by the k_{CO_2} -Q relationship; n equals number of days per month estimated.

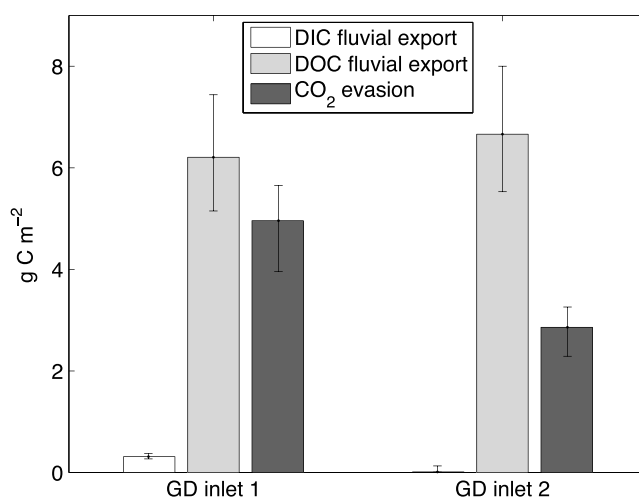


Figure 5. Distribution of C flux pathways of the different C species for the two inlet stream reaches during the open water-sampling period, standardized to upstream drainage area. Error bars represent the minimum and maximum estimate obtained from cumulative standard error of discharge and concentration measurements for DOC and DIC, and cumulative standard errors for tracer injection replicates and concentration measurements for CO₂ evasion.

from the lake and DOC and DIC export via the outlet) (Figure 6). At larger spatial scales, several studies have shown that stream evasion dominates total aquatic evasion and downstream export of carbon [Huotari *et al.*, 2013; Lundin *et al.*, 2013; Wallin *et al.*, 2013]. At the global scale, Raymond *et al.* [2013] estimated that streams and rivers evade 6 times more CO₂ to the atmosphere compared to lakes and reservoirs. Our study adds to current knowledge by using measurements of all relevant parameters in both streams and lakes within one catchment, and by accounting for pronounced spatial and temporal variability in C concentrations and fluxes, to show that stream CO₂ evasion dominates aquatic C loss even at a small spatial scale. The only comparable study

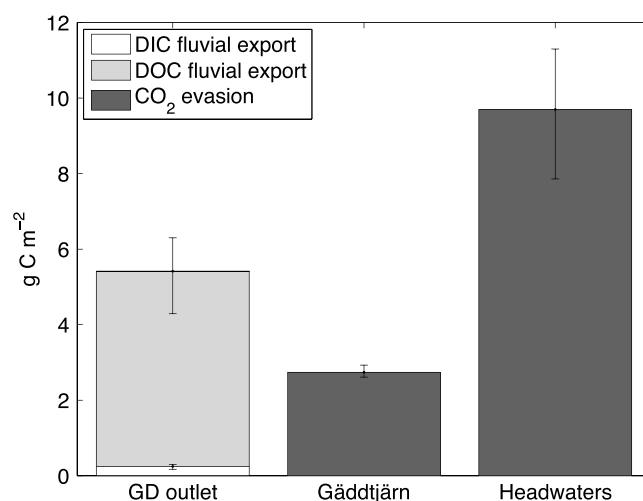


Figure 6. C losses from the Lake Gäddtjärn catchment for the open water-sampling period, standardized to catchment area for lake Gäddtjärn and upstream drainage area. Error bars represent the minimum and maximum estimate obtained from cumulative standard errors of discharge and concentration measurements for DOC and DIC, and cumulative standard errors of tracer injection replicates (streams), wind speed model-derived k_{CO_2} (lakes), and concentration measurements for CO₂ evasion.

headwater streams and lakes was the strongest C loss pathway. CO₂ evasion from headwaters (9.8 g C m⁻²) was 44% higher than the export of organic and inorganic C from GD outlet (total 5.4 g C m⁻²), and more than 3 times greater than CO₂ evasion from Lake Gäddtjärn (2.7 g C m⁻²).

4. Discussion

In this study, we show that CO₂ evasion from headwater systems and especially streams can be the dominant pathway of C loss within the aquatic continuum of a boreal lake catchment, even at small spatial scales. Streams covered only about 0.1% of the Lake Gäddtjärn catchment area yet were responsible for a C loss to the atmosphere greater than the sum of all C losses from Lake Gäddtjärn (including CO₂ evasion

from the lake and DOC and DIC export via the outlet) (Figure 6). At larger spatial scales, several studies have shown that stream evasion dominates total aquatic evasion and downstream export of carbon [Huotari *et al.*, 2013; Lundin *et al.*, 2013; Wallin *et al.*, 2013]. At the global scale, Raymond *et al.* [2013] estimated that streams and rivers evade 6 times more CO₂ to the atmosphere compared to lakes and reservoirs. Our study adds to current knowledge by using measurements of all relevant parameters in both streams and lakes within one catchment, and by accounting for pronounced spatial and temporal variability in C concentrations and fluxes, to show that stream CO₂ evasion dominates aquatic C loss even at a small spatial scale. The only comparable study was reported from a small boreal lake catchment in northern Finland, similar to our lake catchment but with a proportionally larger wetland coverage, and that study showed that downstream C export via streams was larger than stream C evasion [Juutinen *et al.*, 2013]. The lower importance of stream CO₂ evasion in the Finnish catchment may be related to higher share of wetlands, pointing toward the role wetland systems might have on C export. In our study catchment we have a wetland coverage (14%) closer to the national wetland coverage in Sweden (15%), thus making our study catchment more representative for Sweden. In addition, our study catchment has a different topography with steeper slopes than the Finnish catchment, resulting in higher k [Wallin *et al.*, 2011] and hence more CO₂ evasion from streams.

Our data suggest that the pronounced importance of headwater stream CO_2 evasion was largely the result of the high gas transfer velocities k_{CO_2} (median, 66.9 cm h^{-1}), which were much higher than the k_{CO_2} values for the lakes (median, 2.4 cm h^{-1}) (Table 2). In accordance with our study, the large variability in k for low-order streams has in a recent study also been revealed to be important for CO_2 evasion [Crawford *et al.*, 2014]. Our k_{CO_2} determinations for small streams further compare well with other studies performed in Sweden [Wallin *et al.*, 2011] and the UK [Billett and Harvey, 2013] for small streams by tracer injections, as well as the highest k_{CO_2} estimations by chambers determined in Alaska [Crawford *et al.*, 2013] and northern U.S. [Crawford *et al.*, 2014]. Studies that have determined k_{CO_2} for several stream orders [Butman and Raymond, 2011; Humborg *et al.*, 2010; Raymond *et al.*, 2012] have shown that k_{CO_2} is consistently higher for the lowest stream orders. Wallin *et al.* [2011] as well as Raymond *et al.* [2012] found a positive dependence of k_{CO_2} with slope of the stream reach. The majority of the streams within the Gäddejtjärn catchment have a relatively high slope (between 2.0 and 8.5%) creating high water turbulence conditions, thus explaining the relatively high-determined k_{CO_2} values and evasion rates. The importance of turbulence in water for gas transfer velocities is further illustrated by the fact that we found highest and most variable evasion rates during high discharge events, related to high k_{CO_2} (Figure 4). Apparently, the stock and production of DIC in forest and wetland soils is large enough to sustain high CO_2 evasion even at high discharge events, which also are periods of highest k , and thus, stream CO_2 evasion does not seem to be limited by the supply of new CO_2 , at least not at the time scales studied here. Apart from gas transfer velocity, surface area is a term that has high leverage on the calculated stream CO_2 evasion. Our estimate of total CO_2 evasion for the open water period was based on a detailed estimate of stream surface area and included seasonal variation. In addition, the measurements on stream widths compare well to previous estimates made in boreal headwater systems [Wallin *et al.*, 2011] with similar stream areas, lending further support to the robustness of the CO_2 evasion rates presented here.

The differences in $p\text{CO}_2$ between the streams and lakes are minor in comparison to differences in k , even though the streams displayed roughly twice as high $p\text{CO}_2$ compared to the lakes (Table 1). The high evasion rates, particularly in the main inlet stream GD inlet 1, imply that a majority of the stream DIC, either derived from groundwater supply or via in-stream mineralization of organic carbon, is evaded to the atmosphere before entering Lake Gäddejtjärn. The rapid loss in stream CO_2 was well illustrated along the main inlet stream to Lake Gäddejtjärn (GD inlet 1). The upstream site in the inlet stream (Wetland outlet stream) had almost twice as high $p\text{CO}_2$ as the downstream site (GD inlet 1; Table 1), showing substantial CO_2 loss from the stream within only ~150 m distance. A similar high and rapid CO_2 loss was observed in a boreal headwater stream in northern Sweden [Öquist *et al.*, 2009]. The high spatial difference in $p\text{CO}_2$ between the two sites GD inlet 1 and Wetland outlet stream illustrates the importance of wetland systems within the catchment that have been shown to pump large amounts of C to fluvial systems [Abril *et al.*, 2014; Wallin *et al.*, 2014]. However, in this study we show that this DIC is more likely to rapidly evade to the atmosphere, rather than being transported to downstream systems.

The k_{CO_2} values of lakes, estimated from wind speed in our study (median, 2.4 cm h^{-1}), correspond well to k_{CO_2} values determined across Lake Gäddejtjärn with floating chambers (median, 2.3 cm h^{-1} [Krenz, 2013]). Recently, Vachon *et al.* [2013] has shown that lake CO_2 evasion might be overestimated when calculated from wind speed- k_{CO_2} relationships and one-point $p\text{CO}_2$ measurements since both k_{CO_2} and $p\text{CO}_2$ can be spatially heterogeneous in lakes. Their study was, however, conducted on larger lakes (smallest 19 ha) than ours, where the variability in k_{CO_2} is higher due to higher wind speeds (5 m s^{-1}). Therefore, the uncertainty introduced by wind speed-dependent models most likely has the biggest impact on larger lakes and in upscaling to larger spatial scales. Furthermore, another uncertainty associated with estimating k_{CO_2} from wind speed measurements at 10 m height, is the potential effect of thermal stratification of the atmosphere on the wind profile [Arya, 1988]. By applying a theoretical correction (using the log wind law corrected for stability and varying the stability parameter z/L to 1 and -1 for a stable versus unstable thermal atmospheric stratification), our k_{600} would differ with -5 and $+2\%$, respectively, resulting in a corresponding change in CO_2 evasion. Hence, if data on atmospheric stability are available, future studies should incorporate stability corrections on k_{CO_2} when estimating gas evasion from lake systems. Other approaches for measuring lake greenhouse gas evasion, such as the eddy covariance method [Huotari *et al.*, 2011; Podgrajsek *et al.*, 2014] provide a better solution to overcome the shortcomings of wind speed-derived k_{CO_2} , although the method is

costly and challenging to adapt to smaller systems. For example, *Huotari et al.* [2011] revealed that CO_2 evasion rates from *Kortelainen et al.* [2006] based on wind speed-derived k_{CO_2} were $\sim 30\%$ higher than eddy covariance measurements for lakes of similar surface area as the lake in our study. This could imply an overestimated CO_2 evasion from the lakes in this study and that the importance of the CO_2 evasion from headwater streams for C loss in the catchment may be even greater. Several studies based on the eddy covariance method have suggested that waterside convection is driving k at low wind speed [*MacIntyre et al.*, 2010; *Podgrajsek et al.*, 2014; *Read et al.*, 2012]. Accordingly, lake k is not zero at zero wind speed (see equation (4), $k_{600} = 2.07 \text{ cm h}^{-1}$ at zero wind), and the effect of wind speed on k is relatively small at wind speed $< 3 \text{ m/s}$ (varying wind speed by $\pm 20\%$ around the mean daily wind speed of 1.75 m/s changes k_{600} by $\pm 7\%$) [*Cole and Caraco*, 1998]. In our wind-sheltered, small lakes, convection might be an important driver of k and thus CO_2 evasion.

Our results show pronounced spatial and temporal variability of DIC and $p\text{CO}_2$ (Table 1, variability between sites as well as site-specific ranges), which may reflect several processes. The processing of organic carbon into inorganic carbon can be the result of in-stream biological respiration, and DIC derived from soil respiration can be leaching into the stream via shallow groundwater flow, which is in turn closely linked to discharge [*Berggren et al.*, 2009; *Köhler et al.*, 2002]. The results in Figures 3a and 3b support this notion as $p\text{CO}_2$ at GD inlet 1 seems to increase with high discharge events, and during a period of low flow a stronger daily variation is present which may reflect in-stream processes. Our results further show that during a higher discharge event (e.g., Figure 3a GD inlet 1 starting mid-September 2011), the peak in discharge seems to lag somewhat after the peak in $p\text{CO}_2$ concentration. CO_2 supersaturated shallow groundwater and soil pore water has been shown to be the dominant source of C to fluvial networks in similar catchments as our study [*Grabs et al.*, 2012; *Öquist et al.*, 2009], and we suggest that groundwater input of C to fluvial systems is more dominant than in-stream processes in our streams. The importance of groundwater inputs to streams have further been shown for other studies conducted both in Sweden [*Humborg et al.*, 2010] as well as the U.S. [*Crawford et al.*, 2014; *Lottig et al.*, 2011]. The temporal pattern and stronger seasonal variation of $p\text{CO}_2$ at GD outlet on the other hand, more likely reflects the internal lake processes such as photochemical degradation, mixing events, and primary production and respiration fluctuating with temperature.

Furthermore, our results showed that more DOC and DIC are exported from Lake Gäddtjärn than can be explained by the import through the stream inlets (Table 4). In-lake production of DIC, from organic matter mineralization in both lake water and sediments, may provide an explanation for the higher DIC export from the lake than DIC import to the lake. Another potential C source to the lake may be direct groundwater input of shallow groundwater to the lake, since a geographic information system analysis showed that $\sim 20\%$ of the total catchment area of Lake Gäddtjärn drains directly into the lake, and not into one of the two inlet streams. The fact that the outlet discharge almost always was higher than the discharge of the two inlet streams (Figure 3b) also indicates additional water inputs to the lake via groundwater or precipitation on the lake. Presumably, this direct shallow groundwater or overland discharge into the lake may also represent an additional pathway of DOC input to the lake, which could explain the fact that DOC outflow was found to be higher than DOC inflow (Table 4). Comparing the total loss of C through DIC export for the open water period (0.24 g C m^{-2}) and Lake Gäddtjärn total CO_2 evasion for the same period (2.7 g C m^{-2}), other sources of DIC to the lake must be present, calling for the inclusion of shallow groundwater or overland C fluxes into future studies of lake C budgets. Our results indicate that stream DIC import is not a major driver of lake CO_2 evasion, while *McDonald et al.* [2013] modeled for lakes across the U.S. that stream DIC inputs into lakes were important for lake CO_2 evasion. Differences in study scale, approach, as well regional factors such as soil properties or carbon sources may explain different outcomes. Accordingly, our results compare better with similar catchment studies of total organic C export [*Agren et al.*, 2007; *Laudon et al.*, 2004] and combined DIC and DOC export studies [*Wallin et al.*, 2013] conducted in Sweden.

On a landscape scale, however, headwater streams remain the major sites of C loss through CO_2 evasion. Estimating the fluvial export for the total year, including open water-sampling period as well as estimated winter fluvial flux (manual measurement of Q , DOC, and DIC, data not shown) export of DOC and DIC from Lake Gäddtjärn were $6.5 \text{ g C m}^{-2} \text{ yr}^{-1}$ and $0.4 \text{ g C m}^{-2} \text{ yr}^{-1}$, respectively. Total aquatic CO_2 evasion estimated from the catchment for the open water-sampling period is 12.4 g C m^{-2} catchment (assuming no evasion during winter), $\sim 80\%$ coming solely from the headwaters. Our export rates are coherent with

rates previously reported for boreal Sweden by Wallin *et al.* [2013] ($3.9\text{--}9.3\text{ g C m}^{-2}\text{ yr}^{-1}$ for DOC and $0.3\text{--}1.4\text{ g C m}^{-2}\text{ yr}^{-1}$ for DIC).

Incorporating CO_2 evasion from inland waters into terrestrial net ecosystem exchange (NEE) models is rarely done (unless the footprint of eddy covariance towers includes aquatic systems) although inland waters are known to play an important role in total landscape ecosystem balances. Studies conducted over boreal forests for this region in Sweden have presented uptake rates of $52\text{ to }90\text{ g C m}^{-2}\text{ yr}^{-1}$ [Lagergren *et al.*, 2008; Lindroth *et al.*, 2008]. By assuming similar NEE rates for our catchment the proportion of C loss through aquatic systems corresponds to 25–43% of total NEE, of which 24–42% is attributed to CO_2 evasion, and 16–28% solely to headwater stream CO_2 evasion. Evidently, aquatic systems, in general, and headwater systems in particular are important components in the landscape C balance.

5. Conclusions

Studying the C losses from different types of inland waters at a small catchment scale revealed that headwater CO_2 evasion is the major pathway of aquatic C loss, with small headwater streams accounting for the greatest share of the aquatic C loss. The gas transfer velocity appeared to be the most important parameter of CO_2 evasion and thus catchment C loss yet is difficult to quantify. There is a wide variability of gas exchange rates for lake and streams reported in the literature, making global upscaling attempts challenging. For stream and river systems, we suggest that k_{CO_2} is the most critical parameter in determining evasion rates, where highly accurate k_{CO_2} measurements over temporal and spatial scales together with average $p\text{CO}_2$ may provide more robust estimates than are currently available.

To fully understand the sources and fate of C in catchments of small lakes, it is vital to investigate all components of production and pathways of import and export. As one of the few studies conducted on both small lakes and streams within the same catchment, investigating both inorganic and organic carbon fluxes, we provide strong evidence for the importance of smaller headwater systems on a landscape scale. However, the pronounced dynamics in aquatic C fluxes across space and time clearly, even at small spatial scales, illustrates the need for more comprehensive measurements of C processing and fluxes.

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