

Carbon dioxide and energy fluxes over a northern boreal lake

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We present a data set covering three months of carbon dioxide (CO₂) and energy fluxes measured by the eddy covariance method over a northern boreal lake that collects waters from a surrounding catchment dominated by upland forest and wetlands. The data period comprises more than half of the open-water period of 2013. The 30-min averages of CO₂ fluxes ranged from –0.02 to 0.05 mg m⁻² s⁻¹. The monthly CO₂ balances varied from 20 to 30 g m⁻² (emission) between July and September, and decreased in October. A small daytime uptake of CO₂, probably caused by the aquatic plants growing near the measurement mast, was observed from July to September. In September, we observed a temporary enhancement of CO₂ efflux, which was attributed to both high wind speed and rapid cooling of the water and subsequent water column overturn. This peak was accompanied by a period of high sensible heat flux (SHF) from the water to the atmosphere, which is known to enhance the mixing of the water. The seasonal CO₂ flux during the open-water period from the shallow part of the lake was estimated to be 120 g m⁻² yr⁻¹, which corresponds to a loss of approximately 25 g m⁻² yr⁻¹ from the terrestrial part of the catchment, assuming that the observed lake CO₂ emissions result from the decomposition of the imported carbon. At midday, the net energy received by the lake was used mostly to heat the water, and only a minor part of it was converted to SHF and latent heat flux (LHF), with more energy used for the latter. While the SHF showed a clear diurnal cycle with a peak early in the morning and no flux in the afternoon, the diurnal pattern of LHF was more even, with evaporation occurring throughout the day until the freezing of the lake. Our data from this northern lake highlight the importance of thermal water mixing in the air–lake CO₂ flux dynamics and imply that this flux constitutes a significant part of the annual catchment-scale carbon budget.

Introduction

Lakes are an important part of the carbon cycle, particularly within the boreal landscape (Jonsson

et al. 2007, Tranvik *et al.* 2009). A large proportion of the atmospheric carbon fixed into terrestrial ecosystems from the atmosphere in photosynthesis is transported via streams and ditches

downstream in the catchment, before finally entering a lake. In recent years, the aquatic flux of dissolved organic (DOC) and inorganic (DIC) carbon in the catchment-scale carbon balance has received increasing attention (Kling *et al.* 1991, Jonsson *et al.* 2007, Roulet *et al.* 2007, Nilsson *et al.* 2008, Worrall *et al.* 2009, Koehler *et al.* 2011, Gielen *et al.* 2011, Kindler *et al.* 2011). It is now recognized that the fluvial carbon flux may lead to a considerable reduction in the terrestrial net carbon uptake (Kortelainen *et al.* 2006, Kindler *et al.* 2011, Renou-Wilson *et al.* 2014). Therefore, it is crucial to understand the role that lakes play in the transformation and storage of the carbon that originates from the upper reaches of the catchment.

The significance of the aqueous flux varies with land use: in mineral soil catchments with intensive forestry, the aquatic transport has only a small (5%–15%) contribution to the total carbon balance (Jonsson *et al.* 2007, Gielen *et al.* 2011, Kindler *et al.* 2011, Huotari *et al.* 2011), whereas in catchments with a considerable proportion of peatlands this contribution is typically higher (Laudon *et al.* 2004, Rantakari *et al.* 2010, Huotari *et al.* 2013). In addition to stream transport, DOC can also be produced within the lake ecosystem by lake organisms (autochthonic production), although in the boreal landscape the contribution of the allochthonous carbon import from the surrounding terrestrial parts of the catchment is typically more important (Jansson *et al.* 2000). The aquatic carbon transport takes place mostly in the dissolved form; however, a small part of it may also occur in particulate or gaseous forms (Tranvik *et al.* 2009). The relative contribution of inorganic and organic carbon species in the fluvial import varies depending on the ecosystem type, peatland abundance, lithology, runoff and land use (e.g. Laudon *et al.* 2004, Rantakari *et al.* 2010, Tank *et al.* 2012, Camino-Serrano *et al.* 2014).

After entering the lake, the transported carbon is deposited as sediment on the lake bottom, processed in the lake or transported farther away from the lake (Tranvik *et al.* 2009, Kothawala *et al.* 2014). In the lake, decomposition of organic matter by microbes into carbon dioxide (CO₂) or methane (CH₄) strongly depends on the trophic status of the lake and, consequently, on the avail-

ability of oxygen within the lake bottom (Kortelainen *et al.* 2006, Juutinen *et al.* 2009). Greenhouse gas (GHG) fluxes between the water and the atmosphere were previously determined by measuring the gas concentration in the surface water and the concentration in equilibrium with the atmosphere and calculating the flux from this gradient (e.g. Kortelainen *et al.* 2006), by employing floating flux chambers (e.g. Huttunen *et al.* 2002, Eugster *et al.* 2003), and by using the micrometeorological eddy covariance (EC) method in short campaigns (Eugster *et al.* 2003) or in multi-year monitoring (Huotari *et al.* 2011). The magnitude of CO₂ fluxes over a lake is typically small in comparison with that observed in the terrestrial parts of the catchment (Vesala *et al.* 2012). Variations in CO₂ fluxes between different lakes have been explained by climate, lake size, management and peatland coverage in the terrestrial part of the catchment (Rantakari & Kortelainen 2005, Kortelainen *et al.* 2006).

Lakes affect the local climate by storing and releasing heat depending on the temperature difference between the air and water surface. The high heat capacity of waterbodies as compared with that of soil makes them large heat stores (Rouse *et al.* 2005). In addition, there are indications that the dynamics of the heat and carbon fluxes are coupled, e.g. through the impact of the lake–atmosphere sensible heat flux on the convective mixing of the water (Eugster *et al.* 2003). Mixing affects the lake–atmosphere gas exchange by modifying the concentration gradient within the water column, which is the main driver of GHG fluxes, and by enhancing the turbulence at the air–water interface.

Here, we present a new measurement site established at Pallasjärvi, a lake located in the north-boreal part of Finland, and report the lake–atmosphere exchange of CO₂ and energy measured with the EC technique from July to October 2013. The lake collects waters from a sub-catchment where carbon and nitrogen fluxes were intensively monitored during the last 10 years (Hatakka *et al.* 2003, Aurela *et al.* 2009, Lohila *et al.* 2010). A summary of the CO₂ balances and energy fluxes measured at these sites is presented by Aurela *et al.* (2015). The present paper was motivated by the need to understand the magnitude and variation of lake–atmosphere CO₂ and

energy fluxes at this northern site. More specifically, our aim was to (1) understand the diurnal and seasonal CO₂ and energy flux dynamics, (2) examine the relationships between the CO₂ and energy fluxes, and (3) estimate the CO₂ balance for one open-water period.

Material and methods

Site description

The study site was situated at Pallasjärvi (17.3 km²), a northern-boreal headwater lake with an almost pristine catchment area (105.2 km², including the lake itself). It is located in northern Finland ca. 180 km north of the Arctic Circle (Table 1). The mean and maximum depths of the lake are 9 and 36 m, respectively. The catchment area ranges in altitude from 267 m to 809 m a.s.l. Habitat types vary from coniferous forests and open and treed mires to treeless tundra on fells. The lake has four main inlet streams, two of which mainly originate in the adjacent Pallas fells, whereas the two other streams have a more peatland-dominated drainage area (Fig. 1).

Water chemistry monitoring

Regular monitoring of the water chemistry of Pallasjärvi started in 2004. The chemical monitoring of the lake was carried out by the Environmental Administration (Finnish Environment Institute and Centre for Economic Development, Transport and the Environment for Lapland). Water samples were taken at monthly intervals in March–September and in December from depths of 1, 5 and 15 m and from the bottom layer at a depth of 31–35 m. All samples were taken at the site of maximum water depth. Since 2005, continuous discharge monitoring, based on water level recording calibrated against discharge, has been performed in the outflow stream.

Chemical variables were determined using standardised methods in the laboratories of the Environmental Administration. Conductivity (K25) was measured conductometrically with a temperature-compensating cell. The pH value

was obtained electrometrically at 25 °C with a pH meter, and alkalinity was measured by Gran titration. Total phosphorus (P_{tot}) was analyzed by the molybdenum method after digestion with potassium persulphate (K₂S₂O₈). Total nitrogen (N_{tot}) was measured colorimetrically after oxidation with K₂S₂O₈ and reduction in a cadmium–copper (Cd–Cu) column, and the sum of nitrate (NO₃-N) and nitrite (NO₂-N), here referred to as NO₃-N, was measured by reduction of NO₃ to NO₂ in a Cu–Cd column followed by colorimetric determination of azo-colour. Ammonium (NH₄-N) was determined spectrophotometrically with hypochlorite and phenol. Total organic carbon (TOC) was oxidized to CO₂ by combustion and determined by infrared spectroscopy. Total inorganic carbon (TIC) was measured using infrared spectroscopy. Water colour (mg Pt l⁻¹) was measured by optical comparison with standard platinum cobalt disks. Dissolved oxygen content was measured using potentiometric titration. Chlorophyll *a* was measured with a spectrophotometer after filtering of the sample through a GF/C glass fibre filter and extraction using hot ethanol.

Land-use and habitat types

The proportions of different land-use and habitat types were estimated for both the lake catch-

Table 1. Morphometric, hydrographic and meteorological characteristics of Pallasjärvi and its catchment

Catchment area (km ²)	105.2
Lake area (km ²)	17.3
Lake volume (km ³)	0.155
Lake coverage (%)	17.1
Maximum water depth (m)	36
Mean water depth (m)	9
Mean annual (July/January)	
air temperature* in 2003–2013 (°C)	0.19 (11.3/–10.8)
Mean annual (July/January)	
air temperature* in 2013 (°C)	1.04 (12.7/–9.0)
Mean annual/June–September	
precipitation* in 2008–2013 (mm)	596/272
Mean annual/June–September	
precipitation* in 2013 (mm)	516/201
Mean discharge, 2005–2013/2013 (m ³ s ⁻¹)	1.55/1.38

* At Kenttäröva weather station, see Aurela et al. (2015).

ment area and a larger circular area with a radius of 25 km. We used the multi-source National Forest Inventory (MS-NFI) method (Tomppo *et al.* 2008, Tomppo *et al.* 2014) based on satellite image data for 2011 with a pixel size of 20 m × 20 m together with field sample plot data from NFI10 (2007–2008) and NFI11 (2009–2011) computationally updated to 31 July 2011. We also used digital maps to identify the main land-cover types and to separate the forested areas from the other land-cover categories, as well as peatlands from uplands. The data were classified in the following land-cover categories: built-up and agricultural areas, treeless fell-tops, coniferous, deciduous, mixed and treeless upland forests, treed and open mires, and open water areas. The surface areas were calculated with ArcGIS Spatial analyst tools.

Measurements of exchange fluxes and meteorology

In June 2012, a 2-m-tall steel mast was erected for flux measurements in shallow water at the tip of a small spit (68.005°N, 24.205°E) (Fig. 1) (flux site is hereafter called ‘Pallaslompoloniemi’). The spit is part of an esker gravel ridge that is partly located underwater. The ridge divides the area into a shallow and sheltered inlet where two streams from forest- and peatland-dominated terrain discharge (sector 180°–330° from the mast), and a deeper lake (350°–50° through north) (Fig. 1). The mean depth of the lake is about 1.5 and 5 m in the inlet and deeper parts, respectively. Only the flux data with wind from these sectors were included in further analysis, while the rest of the data represent forested terrain, and thus were not used here.

The flux measurement height was 2.5 m above the bottom of the lake. During the measurement period, the water level at the base of the measurement mast varied from 0.3 to 0.6 m. At the beginning of the measurements (from June 2012), the EC flux measurement system consisted of a fast-response LI-7200 CO₂/H₂O analyzer (LI-COR Inc.) and a sonic anemometer (uSonic-3 scientific, Metek GmbH). However, on 18 July 2013, the LI-7200 instrument was replaced by the LI-7000 (LI-COR, Inc.). The

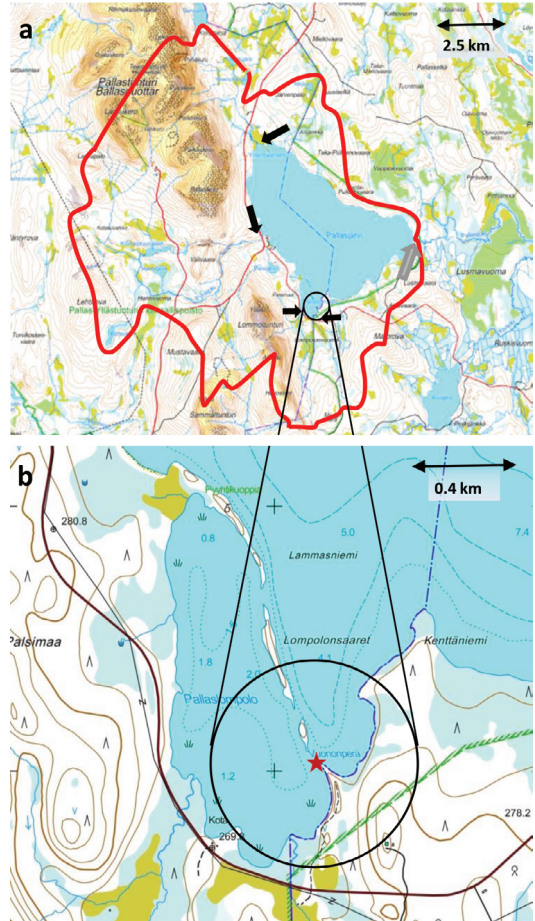


Fig. 1. Map of (a) Pallasjärvi and its catchment (top = north). The thick red line shows the border of the lake catchment. Black and grey arrows indicate the main inlet and outlet streams, respectively. (b) Measurement site Pallaslompoloniemi, located at the Pallaslompolo inlet into Pallasjärvi. The location of the flux mast is indicated with a red star, which is surrounded by a circle with a radius of 400 m to indicate the scale. The blue numbers in the map indicate the depth of the lake. The maps are from the National Land Survey of Finland Database 12/2014.

length of the sample tubing (Bev-A-Line IV, Thermoplastic processes, Stirling, NJ) (ID = 3.2 mm) was 7.5 m, and the flow rate was ca. 6 l min⁻¹. In the inlet there was a 3-cm piece of thicker tubing that prevented water from entering the tubing. In this paper, we show the data collected with the LI-7000 (from 18 July to 20 October 2013) due to technical problems encountered with the LI-7200. Meteorological and environmental variables which included air

temperature (recorded at 1.5 m above the water level), water temperature (Pt100; recorded ca. 0.2 m below the water surface level), water level height (Trafag 8438.66.2646), net radiation (NR lite, Kipp&Zonen), and incoming and reflected photosynthetic active radiation (PAR) (PQS1, Kipp&Zonen), were continuously measured on the lake shore, in the vicinity of the flux mast.

Processing of the flux data

The EC fluxes were calculated from 10-Hz raw data as 30-min averages, after performing a double rotation of the coordinate system (McMillen 1988) and taking into account the density fluctuations related to the water vapour flux (Webb *et al.* 1980). The signals from the sonic anemometer and gas analyzer were synchronized for each averaging period by cross correlation analysis, in which the maximum absolute correlation between the vertical wind speed and concentration is searched. In order to avoid systematic errors, a relatively narrow search window of 0.5–2.0 s was used for the lag time of concentration data, and a predefined value of 1.1 s was applied if the maximum was not found within this window (Laurila *et al.* 2012).

The results were corrected for systematic flux losses using the transfer function method of Moore (1986). The correction included the low-frequency losses due to the 30-min block averaging and, for CO₂ and H₂O fluxes, the imperfect high-frequency response of the measurement system. In order to characterize the high-frequency response, an empirical first-order transfer function was determined from the field data using the sensible heat flux as a reference. Based on spectral analysis, a time constant of 0.08 s was estimated. Idealized cospectral distributions (Kaimal & Finnigan 1994) were assumed when calculating the correction factors.

The 30-min data records were screened, in addition to the basic wind direction selection described above, according to the following criteria: (1) number of spikes in the vertical wind speed, and CO₂ and H₂O concentration data < 100, (2) the stationary test of Foken and Wichura (1996), and (3) variance of CO₂ concentration < 1 ppm².

A source area analysis was carried out with a micrometeorological footprint model to estimate how well the measured fluxes represent the lake area around the measurement mast. The relative source weight functions (flux footprints) were calculated for each 30-min averaging period using the footprint model of Kormann and Meixner (2001). The horizontal dimension of the lake was defined in different sectors, and the modelled cross-wind integrated footprint was accumulated over the radius of the sector corresponding to the observed wind direction. If this cumulative footprint was larger than 70%, the flux data of that period were considered sufficiently representative of the lake surface and accepted for further analysis.

Throughout this paper, we use the convention that a positive energy or CO₂ flux value indicates a flux from the lake to the atmosphere, while a negative value means a flux into the lake.

Results

Characteristics of the lake and catchment

The lake-to-catchment ratio of Pallasjärvi is 16.4%, and the proportion of waterbodies upstream of the lake is < 1%. The most common habitat types in the catchment area are boreal coniferous forests (53% of the land area), open and treed mires (15%) and treeless fells (14%), which is typical for the region and latitude (Table 2). The lake catchment area has a much higher percentage of treeless fells and waterbodies in comparison with the larger area (circle with a radius of 25 km) surrounding the lake, whereas open and treed mires are less common (Table 2). The forests consist mainly of Scots pine (*Pinus sylvestris*) (72% of the forest area) with some Norway spruce (*Picea abies*) (20%) and deciduous trees (mainly *Betula pubescens*, some *Populus tremula* and *Salix* spp.) (8%) (T. Penttilä pers. comm).

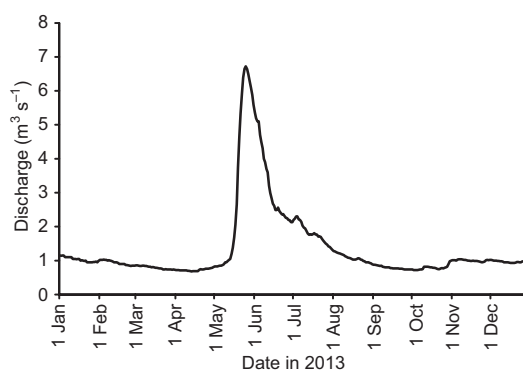
Pallasjärvi can be characterized as a clear-water (mean color 13 mg Pt l⁻¹, total organic carbon 2.7 mg l⁻¹), low-productive oligotrophic lake (mean total phosphorus concentration 5.5 µg l⁻¹, chlorophyll-*a* concentration 2.1 µg l⁻¹)

Table 2. Percentages of land-use types of the total land area within the catchment of Pallasjärvi and within the circle with a radius of 25 km around the study area.

Land-use type	Pallasjärvi catchment	Circle with 25 km radius
Treeless	1	2
Decid. + mixed forest	1	5
Coniferous forest	53	48
Treeless fell	14	2
Treed mire	9	23
Open mire	5	12
Built areas	1	1
Open water	17	6
Cultivated	0	0
Total	100	100

with low ionic strength and conductivity (mean K25 2.7 mS m^{-1}) (Table 3). During 2004–2013, anoxic conditions in the bottom layer of the lake were not detected.

In 2013, the ice melt of Pallasjärvi occurred during the last week of May. The discharge rate at the outlet of the lake showed a sharp flooding peak of nearly $7 \text{ m}^3 \text{ s}^{-1}$ at the end of May (Fig. 2). After that the outflow decreased towards the start of the winter; however, at the end of October there was a moderate increase again. The minimum flow rate in 2013 was measured just before

**Fig. 2.** Daily discharge rates in the outlet of Pallasjärvi in 2013.

the start of the spring flooding in mid-April. The lake froze over at the end of October 2013.

The maximum, daily, surface-water temperature during the study period ($16.9 \text{ }^\circ\text{C}$) was measured at the beginning of August (Fig. 3). The highest hourly (daily-average) air temperature was recorded on 25 July, when it reached $23.9 \text{ }^\circ\text{C}$ ($16.6 \text{ }^\circ\text{C}$). The temperature at which water is most dense (i.e. $4 \text{ }^\circ\text{C}$) was reached in the surface water layer in mid-October, after which it took three weeks for the temperature to reach $0.2 \text{ }^\circ\text{C}$, which was approximately the value at which the water temperature remained for the rest of the winter.

Table 3. Chemical characteristics of Pallasjärvi in the period 2004–2013. All the determinants, except dissolved oxygen and chlorophyll *a*, refer to samples taken from 1-m depth. Dissolved oxygen values refer to samples taken from the bottom layer at depths of 31–35 m. Water samples for chlorophyll-*a* measurements were taken as a composite sample (0–2 m). SD refers to standard deviation.

Variable	2004–2013			2013		
	mean	median	SD	mean	median	SD
Alkalinity (mmol l^{-1})	0.17	0.16	0.06	0.16	0.16	0.005
pH	7.06	7.08	0.16	6.95	7.07	0.37
TOC (mg l^{-1})	2.72	2.70	0.40	2.32	2.40	0.13
Colour (mg Pt l^{-1})	13	15	3.30	12	10	2.67
TIC (mg l^{-1})	1.99	2.00	0.25	1.48	1.50	0.28
P_{tot} ($\mu\text{g l}^{-1}$)	5.5	5.0	3.6	4.2	5.0	1.5
N_{tot} ($\mu\text{g l}^{-1}$)	132	130	28.5	124	130	11.3
$\text{NO}_3\text{-N}$ ($\mu\text{g l}^{-1}$)	7.44	1.50	13.4	7.27	1.00	8.45
$\text{NH}_4\text{-N}$ ($\mu\text{g l}^{-1}$)	5.26	2.50	5.5	3.57	2.50	2.83
K25 (mS m^{-1})	2.69	2.70	0.10	2.73	2.70	0.05
Secchi depth (m)	5.2	5.0	1.03	6	6	0
Dissolved oxygen (mg l^{-1})	8.6	9.0	2.50	7.7	8.5	2.78
Chlorophyll <i>a</i> ($\mu\text{g l}^{-1}$)	2.11	2.00	0.89	2.30	2.50	0.48

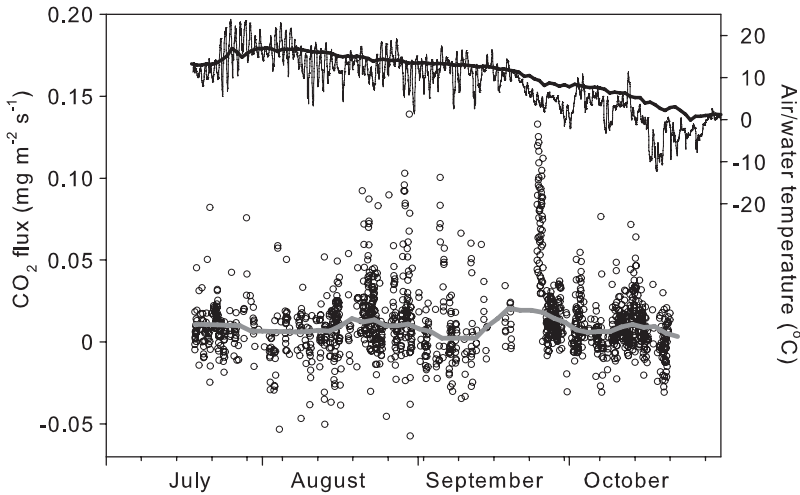


Fig. 3. CO₂ exchange fluxes between the lake and the atmosphere presented as half-hourly (circles) and running median (grey line, sampling proportion of 0.05) values, together with air (thin black line) and surface water (thick black line) temperatures from July to October 2013 at Pallasjärvi.

CO₂ fluxes

During the study period, the half-hourly CO₂ fluxes between Pallasjärvi and the atmosphere mainly fluctuated between -0.02 and 0.05 $\text{mg m}^{-2} \text{s}^{-1}$, a positive flux indicating emission from the lake. However, a single 2-day peak was observed at the end of September when the CO₂ flux reached approximately 0.15 $\text{mg m}^{-2} \text{s}^{-1}$ (Fig. 3). The mean (\pm SE) and median fluxes from mid-July to mid-October were 0.013 (± 0.0005) and 0.009 $\text{mg m}^{-2} \text{s}^{-1}$, respectively. We found no differences in the CO₂ fluxes measured in the two sectors, S to SW and N to NE, which represent the shallow and deeper parts of the lake, respectively (data not shown). Diurnal dynamics were, however, discernable in the CO₂ exchange, with near-zero fluxes observed during daytime and an emission of CO₂ from the water to the atmosphere during nighttime in July–September (Fig. 4). This pattern weakened in October.

We observed a peak CO₂ emission on 24–25 September 2013, simultaneously with both a rapid decrease in the surface water temperature and an increase in the sensible heat flux (Fig. 5). The peak efflux decreased on 26 September, when the water temperature stabilized for a while at about 4.5 °C. The prevailing wind direction at that time was from the north, i.e., from the deeper part of the lake. In addition, the wind speed was high. The relationship between the friction velocity (u_*) and CO₂ efflux (Fig. 6) suggests that the elevated wind speed was one

reason for the enhanced flux. During the September peak this relationship seemed to be even stronger (Fig. 6).

The monthly CO₂ balances (\pm SE) were estimated simply by multiplying the monthly medians of the 30-min fluxes by time. The resulting estimates of monthly CO₂ emissions increased from July (22.4 ± 2.8 g m^{-2}) to September (30.2 ± 4.4 g m^{-2}), and then decreased in October (13.2 ± 1.4 g m^{-2}). By summing the monthly CO₂ balances, we obtained a seasonal emission of about 100 g m^{-2} for the period between 18 July and 20 October 2013, which was very close to the date of the appearance of permanent ice cover. If we assume that (1) the data of the last two weeks of July are representative of the first two weeks with no data, and (2) the mean flux from the end of May (a typical time for ice melt) until the end of June is the same as the mean flux in July, then the annual CO₂ balance is approximately 120 g m^{-2} .

Energy fluxes

The hourly medians of latent heat flux (LHF) varied mainly between 60 and 100 W m^{-2} within the diurnal cycle in July and August, whereas the corresponding sensible heat flux (SHF) was smaller and ranged from 0 to 30 W m^{-2} (Fig. 7c). The highest SHFs, driven by the temperature difference between the water and the adjacent air layer, were observed in July–September in the

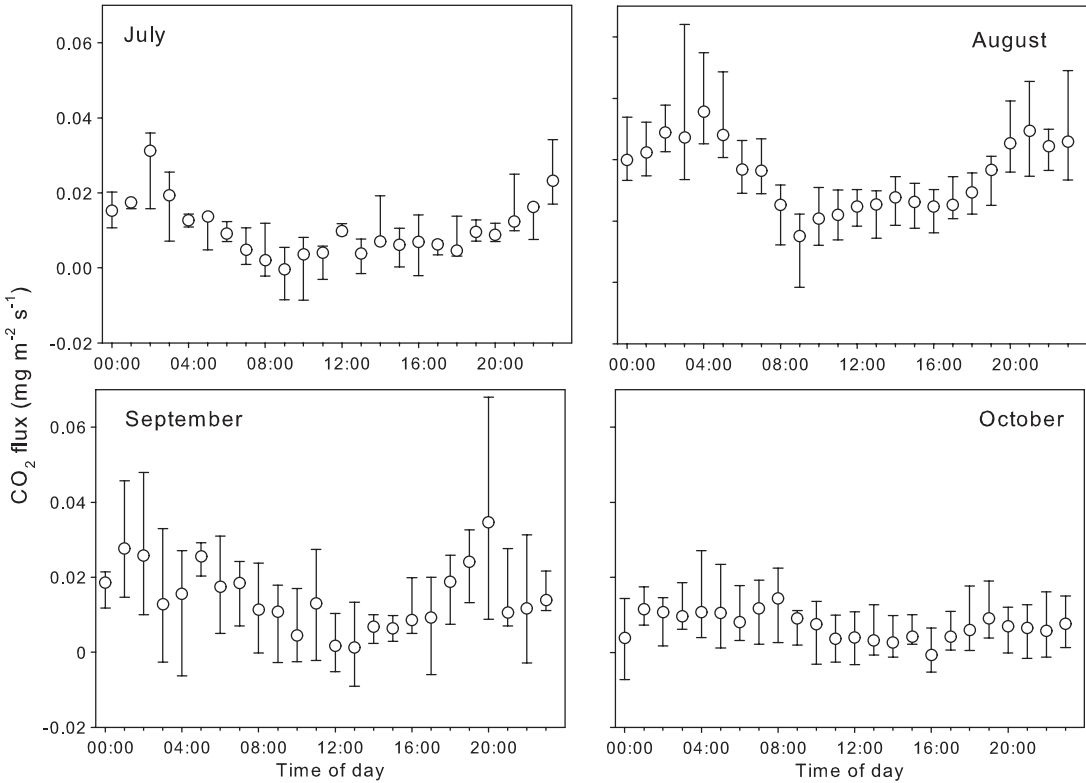


Fig. 4. Hourly median fluxes of CO₂ in July, August, September and October 2013. The error bars indicate the 25th and 75th percentiles. The numbers of accepted observations are 171 (July), 484 (August), 350 (September) and 472 (October).

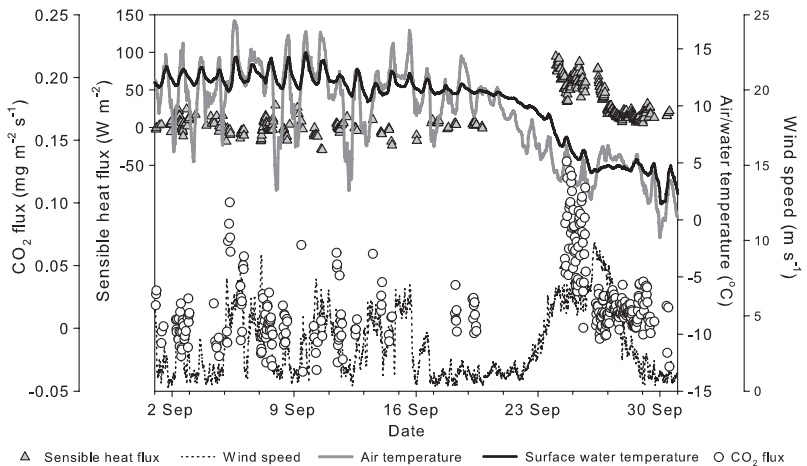


Fig. 5. CO₂ flux, wind speed, air and water temperatures and sensible heat fluxes in September 2013.

early morning at the time of the diurnal minima in air temperature. In October, when the air and water temperatures were already quite low, both LHF and SHF were attenuated; while SHF was close to 0, LHF averaged about 15 W m⁻².

A comparison of incoming net radiation (R_n), which peaked at around 450 W m⁻² in midday in July, with LHF and SHF indicated that only a small portion of the incoming energy was used for generating the turbulent fluxes (Fig. 7a–c);

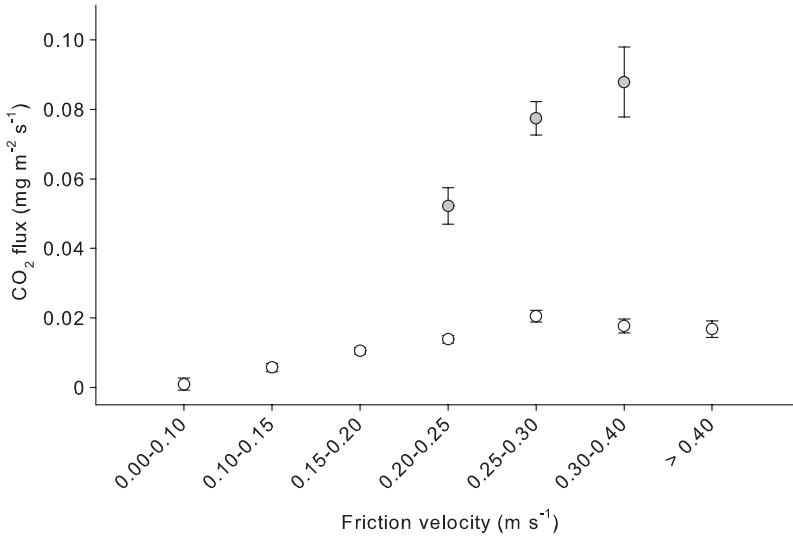


Fig. 6. Mean (\pm SE) CO₂ flux in different friction velocity (u_*) classes during the whole study period (circles) and the CO₂ peak in September 24–25 (filled circles).

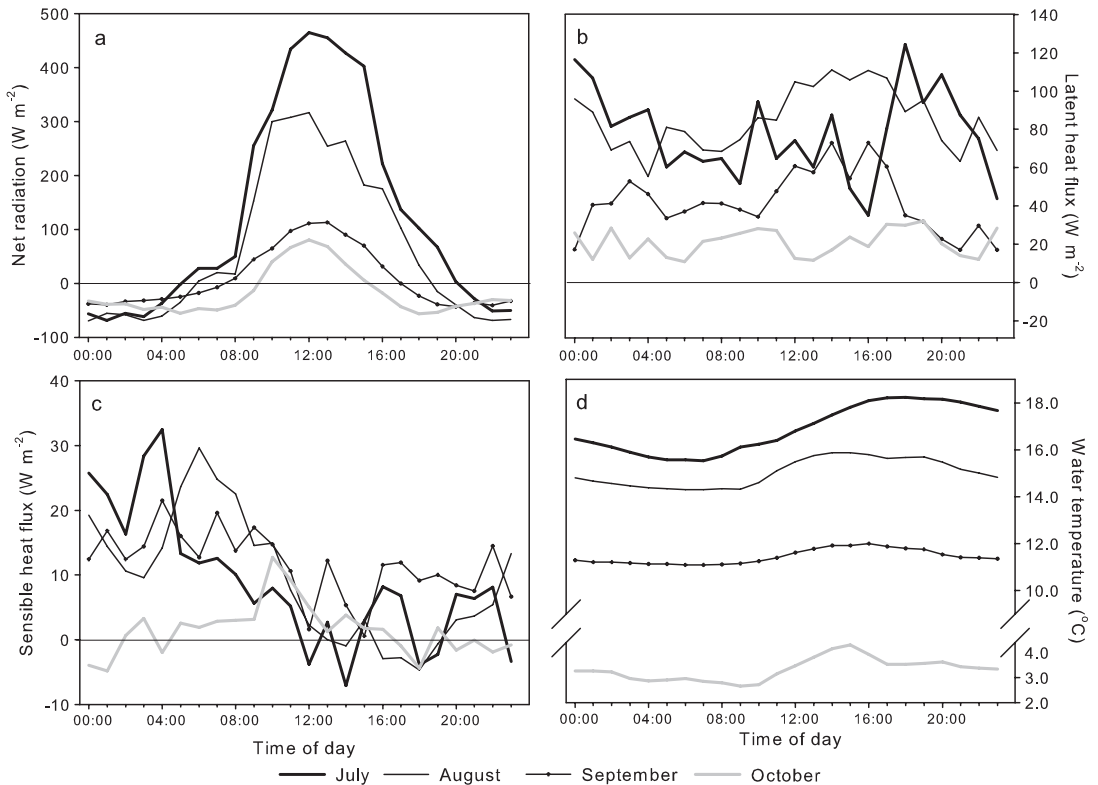


Fig. 7. Hourly medians of (a) net radiation (R_n), (b) latent heat flux (LHF), (c) sensible heat flux (SHF), and (d) water temperature for the period between 18 July and 20 October 2013. The numbers of accepted 30-min observations for R_n /LHF/SHF are 672/174/195 (July), 1488/607/609 (August), 1440/411/438 (September) and 942/528/532 (October).

most of the incoming energy was used for heating the water column during the daytime. This

can be deduced from the water temperature data that show a clear diurnal cycle in July and

August, with a difference between the daily minima and maxima that ranged between 1 and 2 °C (Fig. 7d).

Discussion

The lakes in northern Finland are typically less affected by human activities in comparison with other regions of the country. A colder climate with lower primary production and lower decomposition rates has also resulted in lower lake TOC concentrations in this region than in the country as a whole (Kortelainen 1993). Therefore dilute, clear-water and oligotrophic lakes, such as Pallasjärvi, are common within this remote region (Henriksen *et al.* 1997, Mannio *et al.* 2000).

The CO₂ fluxes observed over Pallasjärvi during the open-water period, with a median (\pm SE) of 210 \pm 14 mg C m⁻² d⁻¹ from July to October, were comparable or higher than those measured over an eutrophic Lake Soppensee in Switzerland, where the efflux averaged 289 \pm 153 mg C m⁻² d⁻¹, and an arctic Lake Toolik in Alaska, where the mean (\pm SE) flux was 114 \pm 33 mg C m⁻² d⁻¹ (Eugster *et al.* 2003). In both cases, the measurement period covered only a few days. In southern Finland, Vesala *et al.* (2006) measured an average flux of 210 mg C m⁻² d⁻¹ over a productive, brown-water lake (Valkea-Kotinen) during an open-water period. In another southern Finnish humic lake, Heiskanen *et al.* (2014) measured much higher fluxes, averaging 1200 mg C m⁻² d⁻¹ in August during seasonal stratification, while the fluxes were lower, averaging 580 mg C m⁻² d⁻¹, during the autumn turnover. In northern Sweden at a moderately humic, non-productive lake, Jansson *et al.* (2008) measured a median flux of 220 mg C m⁻² d⁻¹ from mid-June to mid-October. In a small woodland lake in Minnesota (USA), the CO₂ flux mainly varied from -180 to 210 mg C m⁻² d⁻¹ (Anderson *et al.* 1999). However, immediately after the lake thaw CO₂ was emitted at a rate of 2400–2800 mg C m⁻² d⁻¹, which is comparable to our peak efflux measured on 24–25 September, which averaged 1700 mg C m⁻² d⁻¹. Thus, our study and the few other studies conducted with the EC technique suggest that the mean daily CO₂ fluxes from lakes may

vary by an order of magnitude, even within the same climatic zone, although the variable length and timing of the measurement period hamper a rigorous comparison between these data.

The similarity of the CO₂ fluxes measured from the shallow and deep parts of Pallasjärvi can be attributed to the short measurement period, high variability in the 30-min data relative to the low absolute fluxes and the uneven distribution of wind directions during the measurements. We expected higher fluxes from the shallow, southern sector owing to the direct input of DIC and DOC from the stream water that originates from the forest- and wetland-dominated Pallaslompola catchment (Fig. 1; *see also fig. 1 in Aurela et al.* 2015). However, the northern deeper sector also has a shallow inlet (in the direction of 30°–50° from the mast) and, even though having no visible streams, it most likely receives water rich in inorganic and organic carbon from the surrounding forests and peatlands.

We assume that the diurnal pattern in the hourly median CO₂ fluxes from July to September (Fig. 4) can be attributed partly to the photosynthetic activity of the aquatic plants (*Carex acuta*, *Equisetum fluviatile*) growing on the shore of the shallow part of the lake. They were also found near the measurement mast, which was surrounded by these plants at a distance of about 5–10 m. This observation is in contrast to Vesala *et al.* (2006), who did not find a clear diurnal variation in CO₂ fluxes at Lake Valkea-Kotinen, but did observe a variation in the atmospheric CO₂ concentration. The lowered daytime concentration during the growing season was also strongly evident at our site (data not shown). Here, it must be noted that the footprint of the concentration is typically much larger than that of the turbulent flux (Vesala *et al.* 2008). By screening the flux data with the condition that at least 70% of the cumulative footprint had to be within the lake area, we tried to ensure that the CO₂ flux data were representative of the lake environment. However, it is likely that the advection from the surrounding terrains and CO₂ exchange by the border forests have some impact on the observed fluxes, probably manifested as increased nighttime fluxes.

The estimated monthly CO₂ emissions increased from July to September, after which

emissions decreased in October. A similar increase towards autumn was also reported by Huotari *et al.* (2011) based on a multi-year data set collected at Lake Valkea-Kotinen in southern Finland. However, since the number of accepted flux data from our site during the first three weeks of September was rather low (Fig. 4), the peak that was observed during the cooling event gained a relatively high weight in the calculation of monthly balances. Our estimate of approximately $120 \text{ g CO}_2 \text{ m}^{-2}$ for the annual balance is a much lower value than that ($250\text{--}355 \text{ g CO}_2 \text{ m}^{-2} \text{ yr}^{-1}$) reported by Huotari *et al.* (2011) for a more humic lake during the open-water period, which was 1.5–2 months longer than at our more northern site. Jonsson *et al.* (2008) estimated that the CO_2 emission from a northern Swedish lake during the 2-month period from the end of June to the end of August was about $50 \text{ g CO}_2 \text{ m}^{-2}$. This value is close to our estimate for July–August.

Obviously, a CO_2 emission per square metre of lake surface is largely meaningless in terms of the landscape-scale carbon balance, unless we are able to relate it to the size of its catchment of origin. To understand the role of the lake CO_2 fluxes in a regional carbon balance, we divided the total annual CO_2 emissions from the whole lake by the area of the terrestrial part of the Pallasjärvi catchment (including treeless fell slopes, forests and wetlands) (*see* Table 2). From this calculation we obtained a value of $24.6 \text{ g CO}_2 \text{ m}^{-2} \text{ yr}^{-1}$ (equal to $6.7 \text{ g C m}^{-2} \text{ yr}^{-1}$). If we assume that (1) all of CO_2 emitted from the lake is allochthonous, i.e. imported, (2) all of the imported carbon is emitted to the atmosphere and not stored, for example, in the bottom sediment or lake organisms, and (3) the mean emissions measured at our flux site are representative of the whole lake, then this should equal the amount of carbon leached annually from the land into the lake. Typically in lake environments, CO_2 efflux is the largest gaseous carbon flux (e.g. Riera *et al.* 1999, Repo *et al.* 2007), while DOC is more abundant than DIC in the stream waters of peatland-dominated catchments (e.g. Dawson *et al.* 2004); however, enhanced groundwater impact may make the DIC equally important (Giesler *et al.* 2014).

Our estimate for the leaching of carbon from soil to the lake is slightly lower than that of Jons-

son *et al.* (2007) for a boreal catchment dominated by coniferous forests ($8.6 \text{ g C m}^{-2} \text{ yr}^{-1}$), which represented 6% of the net ecosystem exchange (NEE) of the terrestrial vegetation ($-139 \text{ g C m}^{-2} \text{ yr}^{-1}$, negative value indicating net ecosystem uptake of CO_2). In our study, the proportion was markedly higher (129%), due to low CO_2 uptake by the surrounding terrains within our catchment ($-5.2 \text{ g C m}^{-2} \text{ yr}^{-1}$; Aurela *et al.* 2015). In comparison, Buffam *et al.* (2011) estimated that lake and riverine efflux corresponded to 3% of the forest NEE. On the other hand, Juutinen *et al.* (2013) reported that the proportion of all aquatic fluxes (including gaseous efflux, sedimentation and outflux of dissolved and particulate carbon) of the terrestrial net C intake was 13%. The main reason for the high percentage at our catchment is the small net carbon uptake of the forests in our catchment rather than the high emissions from the lake (cf. Aurela *et al.* 2015). It must be noted here that not all of the DIC that enters the lake through the stream discharge is of biological origin; the groundwater also contains DIC as a result of water–rock reactions (Striegl and Michmerhuizen 1998, Tank *et al.* 2012). In addition, due to the abundance of peatlands in the catchment of the two inlets, the DOC inflow into the footprint of our flux measurement is probably higher than into the whole lake on average. As our measurement mast was located within the littoral area and we extrapolated the results over the whole lake area, we probably overestimated the emission from the lake. It is also clear that part of the carbon leaves the lake via the outlet, but this flux was not taken into account in our calculations.

The energy balance of lakes has gained interest not least because lakes play an important role in the regional energy balance (Rouse *et al.* 2005). At our site, most of the available solar energy was used for heating the water column during the daytime, as was also observed in earlier studies (e.g. Venäläinen *et al.* 1999, Rouse *et al.* 2005, Nordbo *et al.* 2011). In the absence of water temperature profile data, we did not attempt to quantify the energy absorbed by the water column, although from previous studies it seems plausible to expect an energy storage of up to $150\text{--}200 \text{ W m}^{-2}$ in northern lakes in summer (Venäläinen *et al.* 1999). This could explain

most of the “missing” energy (Fig. 7). However, if the short-term fluxes are converted into monthly sums, cumulative R_n exceeds the sum of latent and sensible heat fluxes only in July, thereby indicating a net heating of the water; from August onwards, the sum of the turbulent fluxes exceeds the cumulative R_n , which suggests a cooling of the lake water. It is also noteworthy that in all studies, ours included, LHF does not drop to zero during the summer nights, as is typically the case for vegetated surfaces, where the stomatal closure of plants during the nighttime decouples the soil moisture from the atmosphere.

The dominance of LHF over SHF at our site is consistent with previous observations over boreal lakes (Nordbo *et al.* 2011, Heiskanen *et al.* 2014). The variation in 30-min LHF was large and no clear diurnal pattern was found in July. However, in August and September a diurnal maximum was discernable in the afternoon. This is expected, since LHF is driven by the vapor pressure deficit of the ambient air, which above a lake surface on a clear summer day typically has its minimum in the afternoon (Nordbo *et al.* 2011). The highest SHFs, driven by the temperature difference between the water and the adjacent air layer, were observed in July–September in the early morning at the time of the diurnal minima in air temperature. In October, when the lake froze, both LHF and SHF were already attenuated close to zero. In general, these observations are in line with earlier energy flux studies conducted on boreal lakes (e.g. Heikinheimo *et al.* 1999, Venäläinen *et al.* 1999, Vesala *et al.* 2006, Nordbo *et al.* 2011, Heiskanen *et al.* 2014). However, at our site LHF were slightly lower and the diurnal cycle was less evident.

Linking energy and CO₂ fluxes is necessary to more thoroughly understand the factors that drive the fluxes, particularly in northern ecosystems with a short growing season and limited available energy (Rouse 2000). The most obvious link is typically found between CO₂ exchange and snow/ice melt which is strongly coupled to the decrease in albedo and increase in R_n . In lakes, the ice cover prevents the atmosphere–lake exchange, which is restored after the ice-covered period. However, there are also other ways by which the constituent and energy fluxes may be coupled. For example, the peak

CO₂ efflux observed at our site in September, which occurred at the same time as the cooling of the lake water, is a likely example of such a link.

Although the September flux peak occurred at the same time as strong winds (Fig. 5), the magnitude of CO₂ efflux (Fig. 6) suggests that other important phenomena also contributed to the peak. It is probable that the high efflux resulted from the water column overturn that followed the cooling of the surface water, which started to sink due to increased density. At the same time, more CO₂-rich water from the bottom layers presumably rose closer to the surface, after which the excess CO₂ was released into the atmosphere. This explanation is supported by the temperature profile data taken from the deepest part of the lake, which show that a full overturn took place a few days later (data not shown). Our observation is similar to that of Eugster *et al.* (2003) for two different lakes in Switzerland and the Canadian Arctic: convective mixing in the water column markedly enhances the CO₂ emission rate. Similarly, Jonsson *et al.* (2008) observed high CO₂ emissions in September at the time of autumn circulation in the lake, and Huotari *et al.* (2011) reported bursts of CO₂ out of a boreal lake, associated with similar convection events. In general, both wind speed (mechanical turbulence) and buoyancy flux (thermal turbulence) are known to contribute to the gas transfer coefficient and thus to the air–water gas exchange (MacIntyre *et al.* 2010, Heiskanen *et al.* 2014). Bearing in mind that the gas flux can be expressed as a product of the concentration gradient and the transfer coefficient, we can conclude that the high CO₂ emissions observed in September resulted from both the enhanced air–water transfer rate and the increased concentration gradient between the water and the atmosphere.

Conclusions

Pallasjärvi, a northern oligotrophic lake, showed CO₂ emission rates which fall within the wide range of fluxes reported in other EC studies that were conducted on lakes in northern and middle latitudes. Although the fluxes were small compared with those typical for terrestrial ecosystems,

they must be accounted for in the landscape-scale carbon balance estimates, since they significantly decrease the land-surface carbon accumulation. We found no difference in the CO₂ fluxes among the different parts of the lake, which however may be partly due to the short measurement period. A CO₂ flux peak that coincided with the high wind speed and water column overturn was observed in September. The peak was not only explained by the strong winds, but also by the thermal mixing in the water column which raises deep waters with higher CO₂ concentration to the surface. The sensible and latent heat fluxes between the lake surface and the atmosphere were typically low, which showed that a large part of the received energy is used to heat the water. Diurnal variation was observed in sensible heat fluxes, which was driven by the difference in air and water temperatures. In latent heat fluxes the diurnal cycle was less clear, although the flux peaked in the afternoon in August and September. More extended measurements will reveal if the dynamics shown here will persist over the full open-water period and will determine the frequency of CO₂ flux peaks related to the cool-air-warm-water interactions. The exchange rates of CO₂ and CH₄ between the lake and the atmosphere in spring, both during and after the ice melt, will be of particular interest at this northern lake with a long ice-cover period. In future, it will be a challenge to integrate the lake CO₂ fluxes with all the different carbon flux components measured within the Pallaslompola catchment, including vertical (ecosystem-atmosphere) CO₂ and CH₄ fluxes and horizontal (lateral water-carried) DIC and DOC fluxes in order to compile a full landscape-level carbon balance. This would improve our understanding of the complex interactions between the different forms of carbon and their formation and transport mechanisms within a catchment under varying climatic conditions.

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