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Spatial and temporal variability of N_2O emissions in a subtropical forest catchment in China

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Abstract. Subtropical forests in southern China have received chronically large amounts of atmogenic nitrogen (N), causing N saturation. Recent studies suggest that a significant proportion of the N input is returned to the atmosphere, in part as nitrous oxide (N_2O). We measured N_2O emission fluxes by closed chamber technique throughout two years in a Masson pine-dominated headwater catchment with acrisols (pH ∼ 4) at Tieshanping (Chongqing, SW China) and assessed the spatial and temporal variability in two landscape elements typical for this region: a mesic forested hillslope (HS) and a hydrologically connected, terraced groundwater discharge zone (GDZ) in the valley bottom. High emission rates of up to 1800 µg N_2O-N m^{-2} h^{-1} were recorded on the HS shortly after rain storms during monsoonal summer, whereas emission fluxes during the dry winter season were generally low. Overall, N_2O emission was lower in GDZ than on HS, rendering the mesic HS the dominant source of N_2O in this landscape. Temporal variability of N_2O emissions on HS was largely explained by soil temperature (ST) and moisture, pointing at denitrification as a major process for N removal and N_2O production. The concentration of nitrate (NO_3^-) in pore water on HS was high even in the rainy season, apparently never limiting denitrification and N_2O production. The concentration of NO_3^- decreased along the terraced GDZ, indicating efficient N removal, but with moderate N_2O-N loss. The extrapolated annual N_2O fluxes from soils on HS (0.54 and 0.43 g N_2O-N m^{-2} yr^{-1} for a year with a wet and a dry summer, respectively) are among the highest N_2O fluxes reported from subtropical forests so far. Annual N_2O-N emissions amounted to 8–10 % of the annual atmogenic N deposition, suggesting that forests on acid soils in southern China are an important, hitherto overlooked component of the anthropogenic N_2O budget.

1 Introduction

The global atmospheric concentration of nitrous oxide (N_2O), an important greenhouse gas and decomposer of stratospheric ozone, has increased from a pre-industrial level of 270 ppbv to 322 ppbv in 2008 (WMO, 2009). The global source strength is estimated to be 17.7 Tg N yr^{-1}, with agriculture contributing 2.8 (1.7–4.8) Tg N yr^{-1} and soils under natural vegetation 6.6 (3.3–9.0) Tg N yr^{-1} (IPCC, 2007; Hirsch et al., 2006). Both estimates based on bottom-up approaches (Stehfest and Bouwman, 2006) and on observations of the N_2O atmospheric column (Kort et al., 2011; D'Amelio et al., 2009; Hirsch et al., 2006) suggest that 50–64 % of the atmospheric N_2O derive from the (sub)tropical zone 0° to 30° N. Much of the N_2O attributable to this zone is emitted from forest soils (Melillo et al., 2001; Werner et al., 2007a; Rowlings et al., 2012). So far, N_2O flux data from subtropical forests are scarce making this biome an underinvestigated component of the global N_2O budget.

Large parts of southern China are situated in the humid subtropics and the dominant forest types are evergreen broadleaf and coniferous forest, many of which are found as patches in densely populated areas with intensive agriculture. Due to strong increases in the emission of nitrogen oxides (NO_x) and ammonia (NH_x), caused by combustion of fossil

fuels and massive fertilizer use, respectively (Xiong et al., 2008; Liu et al., 2011b), these forests receive high amounts of reactive nitrogen (N_r) by atmogenic deposition, mostly as ammonium (NH_4^+). A recent study of five forested watersheds in southern China found various degrees of N saturation (Chen and Mulder, 2007a), i.e. rates of atmogenic N_r input exceeding uptake by vegetation (Aber et al., 2003) and causing elevated concentrations of nitrate (NO_3^-) in the root zone. Larssen et al. (2011) reported moderate NO_3^- export with stream water in these watersheds, despite low forest productivity, suggesting major unaccounted N sinks in these ecosystems.

One possible fate of excess N may be gaseous emission as nitric oxide (NO), N_2O or dinitrogen (N_2) produced during nitrification and denitrification. Known factors controlling N_2O emissions are availability of inorganic N and degradable organic carbon (C), soil temperature, soil moisture and soil pH (Parton et al., 1996; Flessa et al., 1995; Smith et al., 2003; Weier et al., 1993; Simek and Cooper, 2002). The monsoonal climate in southern China provides favorable conditions for both nitrification and denitrification as much of the annual N input, dominated by NH_4^+, occurs during rain storms in summer when soils are warm and N turnover rates high. Moreover, forests in southern China are often found on acidic soils. Acidity has been reported to support high N_2O/N_2 product ratios in denitrification (Liu et al., 2010b; Bergaust et al., 2010) which could result in high N_2O emissions. Acidity also inhibits autotrophic nitrification by reducing the availability of ammonia (NH_3), the actual substrate for membrane-bound ammonia mono-oxygenase in ammonia oxidizing bacteria (De Boer and Kowalchuk, 2001); on the other hand, low soil pH may enhance the apparent N_2O yield of nitrification, possibly involving chemodenitrification of nitrite (NO_2^-) from either nitrification or dissimilatory NO_2^- reduction (Mørkved et al., 2007). Therefore the net effect of acidity on the production of nitrification-related N_2O is still uncertain.

Few studies have addressed N_2O emissions in Chinese subtropical forests. Tang et al. (2006) studied the temporal variation of N_2O fluxes in a pine, a mixed and an evergreen broadleaf forest on acid soils in the Dinghushan catchment, southern China throughout one year and found highest emissions during the wet-hot season. Based on weekly to biweekly measurements, they estimated an annual emission of 3.2 kg N_2O-N ha^{-1}, which is above the reported average N_2O emission of 1.2–1.4 kg N ha^{-1} yr^{-1} for tropical forests (Stehfest and Bouwman, 2006; Werner et al., 2007b). Fang et al. (2009) reported 2.0 to 2.4 kg N_2O-N loss ha^{-1} yr^{-1} based on monthly measurements along a hillslope in an evergreen broadleaf forest in the same catchment. Lin et al. (2010, 2012) estimated smaller average annual N_2O emissions of 0.13 and 0.71 kg N ha^{-1} for pine forests in Hubei province, probably due to the higher pH of these forest soils.

Estimates of annual N_2O emissions in forested catchments are fraught by large temporal and spatial variability commonly reported for N_2O fluxes. Forests in subtropical China are typically found on sloping terrain and mountain ridges, resulting in hydrological gradients known to affect N_2O emissions. For example, Fang et al. (2009) found that N_2O emissions increased with soil moisture from hilltop to the bottom of a hillslope. Besides soil moisture, other soil factors, such as pH, texture, vegetation type and productivity may vary along hydrological flow paths, resulting in distinct spatial patterns of N_2O emissions (Osaka et al., 2006; Lark et al., 2004). Ultimately, soil factors related to topography may shape distinct nitrifier and denitrifier communities with respect to N_2O turnover (Philippot et al., 2009; Dörsch et al., 2012; Wessen et al., 2011; Banerjee and Siciliano, 2012). Thus, understanding of spatial and temporal variability of N_2O emission fluxes and their drivers on a landscape level are indispensable for improving regional estimates of N_2O emissions.

The objective of the present study was to estimate N_2O fluxes in an N-saturated forested catchment in Southwest China and to explore their main drivers in space and time. More specifically, we investigated the seasonal distribution of emission fluxes (dry versus wet season), the role of storm flow conditions on peak N_2O emissions and the spatial distribution of N_2O emissions along a hydrological flow path on a hillslope (HS) and a groundwater discharge zone (GDZ). We hypothesized that the emission strength is higher during the wet season than the dry season and that a substantial part of the annual N_2O emission occurs during transient N_2O emission peaks triggered by rain episodes. In addition, based on previous observations that the water-saturated GDZ has large N retention (Larssen et al., 2011), we hypothesized the GDZ to act as a zone of increased denitrification and hence high N_2O emission relative to the well-drained HS. Emissions of N_2O were measured together with soil parameters from summer 2009 to autumn 2010, along two transects in each of the two landscape elements, hillslope (HS) and groundwater discharge zone (GDZ). To improve flux estimates in time, a regression model is proposed interpolating temporally discrete N_2O measurements on the HS based on continuously recorded soil moisture and temperature.

2 Materials and methods

2.1 Site description

The Tieshanping (TSP) catchment is a 16.2 ha headwater catchment, 450 m a.s.l., located on a forested ridge about 25 km Northeast of Chongqing city, SW China (29°38′ N 104°41′ E, Fig. 1a). Details on climate, vegetation and soil characteristics as well as atmogenic N deposition can be found elsewhere (Chen and Mulder, 2007a). The area has a typical subtropical monsoonal climate with a mean

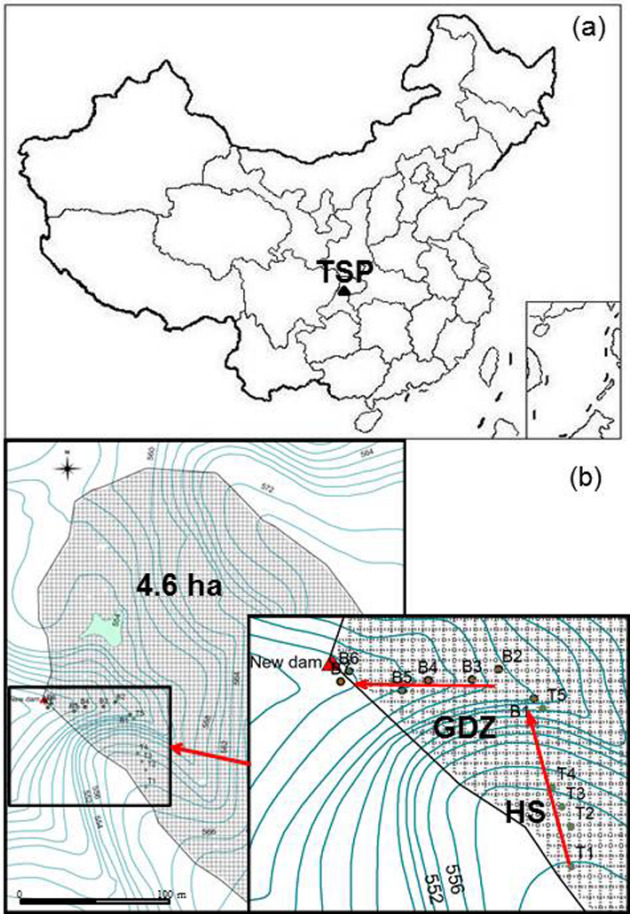

Fig. 1. (a) Location of the Tieshanping (TSP) forest catchment, Chongqing, China and **(b)** plot layout in the TSP catchment. Plots T1 to T5 constitute transect *T* on the hillslope (HS); plots B1 to B6 constitute transect *B* in the groundwater discharge zone (GDZ). Increasing plot numbers denote decreasing elevation along the transects. Plots T5 and B1 are at the intersection of HS and GDZ.

annual precipitation of 1028 mm and a mean annual temperature of 18.2° C (three year average, 2001–2003). Seventy-five percent (75 %) of the rainfall occurs during summer (April to September). Mean annual inorganic N deposition (from 2001–2003) was $4\,g\,m^{-2}$, 61 % of which occurred in the form of NH_4^+-N (Chen and Mulder, 2007b). Data for more recent years show increasing N deposition rates to $> 5\,g\,m^{-2}\,yr^{-1}$ and confirm the relative importance of NH_4^+ (Lei Duan, personal communication, 2011). The vegetation is a coniferous-broadleaf mixed forest dominated by Masson pine (*Pinus massoniana*) with a well-developed understory of evergreen shrubs (Chen and Mulder, 2007a). Land use is a naturally regenerated, non-managed secondary forest after the original forest was cut during 1958–1962.

For the present study, a representative 4.6 ha sub-catchment was selected, consisting of a NE facing hillslope (HS) and a hydrologically connected SE–NW oriented, terraced valley bottom acting as a groundwater discharge zone (GDZ) (Fig. 1b). The HS is dominated by acidic, loamy yellow mountain soils (Haplic Acrisols; WRB, 2006), developed from sandstone, the predominant bedrock in the area. The soils have only a thin organic layer (O-horizon; 0–2 cm) (Table 1), likely reflecting high turnover rates of soil organic matter (SOM) in the warm and wet climate (Raich and Schlesinger, 1992; Zhou et al., 2008). Soils in the GDZ have developed from colluvium (Cambisols) and were terraced during the 1960s for vegetable cultivation but abandoned shortly after. The soils in the GDZ have a low hydraulic conductivity (Sørbotten, 2011), lack horizontal differentiation (no distinct O-horizon) and have high groundwater levels. The vegetation in the GDZ consists of sparse shrubs and grasses while trees taller than 2 m are absent. Clay mineralogy on HS and in GDZ is dominated by kaolinite. A recent study (Sørbotten, 2011) showed that considerable interflow occurs along HS during rainfall episodes. Excess water moves laterally down slope in surface horizons overlaying the argic B-horizon, which has a low hydrological conductivity, thus draining the O/A- and AB-horizons on the HS relatively quickly. Discharge of interflow into GDZ results in episodically high groundwater levels and periodic waterlogging during summer (Sørbotten, 2011). Major soil characteristics for both locations are summarized in Table 1.

2.2 N_2O flux measurements

N_2O emissions were measured manually by closed chamber technique along two transects in the sub-catchment. Transect (*T*) was established perpendicular to the contour lines on the HS stretching from the hilltop close to the watershed divide to the bottom of the hill close to the GDZ. Transect (*B*) was established in the GDZ following an elevation gradient of six hydrologically connected terraces (Fig. 1b) towards a flume, which was used to measure runoff from the sub-catchment. Six plots for N_2O emission measurements ($n = 3$) and sampling of soil and pore water were established along the *T* transect on HS (T1 to T5 and B1) and five plots along the *B* transect in GDZ (B2 to B6; Fig. 1b) with plot B1 representing the transition between HS and GDZ. Two sampling strategies were applied to explore N_2O fluxes. A first measurement campaign during summer 2009 (July to September) with frequent flux measurements explored N_2O emission dynamics in response to sudden increases in soil moisture after rainfall episodes. A more regular sampling scheme with bi-weekly measurements was applied throughout the dry-cool season (November 2009 to March 2010) and during the subsequent summer (from April to August 2010), interrupted by more intensive measurement campaigns after individual rain storms (> 20 mm). For logistic reasons, only a selection of plots (T1, T3, T5, B1, B2, B4 and B6) was monitored during the dry season. In total 38 triplicate flux measurements were conducted on plots T1, T3, T5, B1, B2, B4 and B6 and 20 on plots T2, T4, B3 and B5.

Table 1. Characteristics of top soils at individual plots on the hillslope (HS) and in the groundwater discharge zone (GDZ).

Plot		Horizon/ layer	pH (H_2O)	TOC	TN	C/N	BD[a]	NH_4^+ ex	NO_3^- ex	NH_4^+ sw	NO_3^- sw	IDR[b]	Ex situ potential of N_2O loss[b]
				mg g^{-1}			g cm^{-3}	μg N g^{-1} dw soil		mg N L^{-1}		nmol N_2O-N g^{-1} dw soil h^{-1}	nmol NO_3^- g^{-1} dw soil
HS	T1	O	3.96	226.1	8.7	26.1	–	49.6 (20.2)	37.9 (14.7)	–	–	8.4	523
		A	3.78	10.7	0.5	20.4	0.78 (0.05)	21.5 (8.6)	59.6 (29.2)	0.11 (0.05)	15.7 (3.1)	12.2	552
	T2	O	3.74	354.5	15.7	22.6	–	81.9 (11.7)	23.7 (39.4)	–	–	16.7	2336
		A	3.84	15.0	0.8	19.4	0.72 (0.03)	34.4 (11.9)	12.7 (19.2)	0.75 (0.50)	14.9 (4.6)	13.3	1056
	T3	O	4.07	220.2	9.9	22.2	–	85.9 (19.6)	35.7 (25.1)	–	–	40.4	1924
		A	4.13	12.1	0.6	19.1	0.73 (0.15)	59.6 (30.1)	20.1 (11.2)	0.15 (0.04)	13.3 (2.8)	31.4	1624
	T4	O	3.74	326.0	13.4	24.3	–	89.2 (16.6)	29.1 (14.2)	–	–	6.4	460
		A	3.76	14.3	0.7	20.6	0.47 (0.03)	32.1 (11.6)	13.0 (4.7)	0.07 (0.02)	13.5 (4.1)	16.3	1556
	T5	O	3.67	218.1	10.1	21.6	–	55.3 (19.4)	33.6 (43.6)	–	–	26.5	1706
		A	3.75	6.5	0.4	17.2	0.76 (0.16)	25.8 (6.7)	17.4 (6.5)	0.07 (0.02)	13.4 (1.7)	14.7	1394
	B1	O	3.99	234.9	11.1	21.1	–	83.0 (16.6)	25.6 (2.9)	–	–	40.8	1236
		A	4.02	6.05	0.3	17.6	0.42 (0.07)	23.3 (5.8)	6.6 (2.3)	0.17 (0.07)	13.8 (1.5)	11.3	1014
GDZ[a]	B2	0–20 cm	4.46	14.5	0.9	15.6	1.57 (0.09)	3.6 (1.3)	6.3 (1.1)	1.25 (1.31)	6.2 (0.5)	1.8	318
	B3		4.31	12.8	0.7	18.6	1.67 (0.13)	2.8 (0.8)	2.6 (1.5)	0.01 (0.01)	1.5 (0.7)	12.3	494
	B4		4.38	23.1	1.6	14.5	0.69 (0.02)	7.9 (1.8)	1.6 (0.5)	0.02 (0.01)	5.4 (0.5)	21.0	53
	B5		4.42	11.8	0.8	14.4	0.63 (0.08)	2.9 (1.0)	2.4 (0.5)	0.02 (0.01)	0.7 (0.4)	20.1	201
	B6		4.80	40.5	2.8	14.7	1.01 (0.13)	4.3 (0.4)	0.6 (0.5)	1.12 (0.16)	0.2 (0.0)	28.0	357

TOC: total organic carbon; TN: total nitrogen; C/N: ratio of TOC and TN; BD: bulk density; NH_4^+ ex and NO_3^- ex: average 2M KCl-extractable NH_4^+ and NO_3^-; NH_4^+ sw and NO_3^- sw: dissolved NH_4^+ and NO_3^- in soil water and groundwater in summer 2010 (11 May–26 September 2010); IDR: instantaneous denitrification rate in anoxic soil slurries with ample NO_3^-; ex situ potential of N_2O loss: nmol NO_3^- g^{-1} dw soil denitrified before maximum N_2O reduction to N_2 was observed in anoxic soil slurries with ample supply of NO_3^-. Data in parentheses are the standard error.

[a] Data from Sørbotten (2011) obtained using 100 cm^3 steel cylinders in three replicates for A-horizon for each sampling plot.

[b] Data from Zhu et al. (2013a).

For flux sampling, closed, vented, zinc-coated iron chambers (30 cm diameter and 30 cm in height) (Hutchinson and Mosier, 1981) were carefully pushed approx. 2 cm into the soil. Given the loamy texture and the limited number of roots in the mostly moist surface soil, this deployment method was preferred over installing permanent chamber bases which might impair surface run off and interflow during rain storms. The chamber deployment was quality assured by checking for deviations of initial (1 min after deployment) chamber N_2O concentration from ambient ($\sim$ 0.34 ppmv) and using CO_2 accumulation over time (not shown) to check for leakage, both of which were regular in most of the cases. Gas samples were taken from a sampling port on top of the chambers at 1, 5, 15 and 30 min after deployment by a 20-mL plastic syringe. Samples were transferred immediately to pre-evacuated 12-mL vials, crimp-sealed with butyl septa (Chromacol, UK), resulting in an overpressure to avoid contamination due to pressure changes during shipment. Chamber temperature was recorded by inserting a handheld digital thermometer into the sampling port at beginning and end of chamber deployment. Gas samples were shipped to the Norwegian University of Life Sciences and analyzed within half a year after sampling. The vials kept overpressure during shipment and storage. The samples were analyzed by ECD-gas chromatography (Model 7890A, Agilent, Santa Clara, CA, US). Carbon dioxide (CO_2) and N_2O were separated on a 20-m wide bore (0.53 mm diameter) Poraplot *Q* column run at 38 °C after passing a packed Heysept column used for backflushing water. Helium (He 5.0) was used as carrier gas. The ECD was run at 375 °C with 17 mL min^{-1} Ar/CH_4 (90/10 vol %) as makeup gas. N_2O emission rates (μg N m^{-2} h^{-1}) were calculated based on the rate of change in N_2O concentration in the chamber, the internal volume of the chamber, the covered soil surface area and the average chamber temperature during deployment. For determining the rate of change in N_2O concentration in the chamber, which is the slope of a linear or a second-order polynomial fit of the concentration data against time, the R-square of both regressions were compared. Cumulative N_2O emissions for observation periods in summer 2009 (27 days, 12 July to 8 August) and summer 2010 (106 days, 10 May to 24 August) were calculated assuming a constant flux rate between sampling dates. For the calculation of cumulative N_2O emissions, only the periods with sampling frequency high enough to obtain the cumulative N_2O emission rates of the whole summer based on measurements were used.

2.3 Soil sampling and analysis

Composite soil samples were taken on 15 October 2009 from the eleven plots used for N_2O flux measurements, and analyzed for pH, total organic carbon (TOC) and total nitrogen (TN). Samples at plots T1 to T5 and B1 were taken from the O- (0–2 cm), A- (2–8 cm), AB- (8–20 cm) and B- (20–40 cm) horizons. In GDZ (plots B2 to B6), where an organic horizon was absent, the homogeneous 0–20 cm and 20–40 layers were sampled. Denitrification characteristics of the soils, investigated in a laboratory study, are reported elsewhere

(Zhu et al., 2013a). Briefly, the instantaneous denitrification rate (expressed in nmol N_2O-N g^{-1} dw soil h^{-1}) in soils from the different plots was estimated from the N_2O accumulation rate in acetylene-treated, anoxic soil slurries at 20°C in the presence of ample NO_3^--N (2 mM). The soils' ex situ potential for N_2O loss (in nmol NO_3^--N g^{-1} dw soil) was estimated as the amount of NO_3^--N respired to N_2O before N_2O reductase activity was fully expressed and rapid N_2O reduction to N_2 occurred. Since instantaneous denitrification rates were low in the deeper horizons, only data from the O- and A-horizons of HS and the 0–20 cm layer of GDZ are presented here.

Soil pH (H_2O) was measured in a suspension of 10 g dry weight (dw) soil in 50 mL deionized water with an Orion SA720 electrode pH-meter connected to an Orion ROSS Ultra pH Electrode. The contents of TOC and TN in the soil were measured in the fine earth fraction (< 2 mm), obtained after drying at 105 °C and sieving, using an element analyser (Vario EL III, Elementar Analysensysteme GmbH, Germany) at the Research Center of Eco-environmental Science, Chinese Academy of Science (RCEES, CAS). In addition, soil samples from the horizons and layers of all plots were collected three times (15 October 2009, 22 January 2010 and 29 April 2010) to measure KCl-extractable nitrate ($NO_{3\,ex}^-$) and ammonium ($NH_{4\,ex}^+$). After being sampled, soils were stored in a freezer and extracted within one month, except the soil taken on 22 January 2010, which was extracted together with the samples taken on 29 April 2010. Soil samples were extracted by shaking 10 g dw equivalents of fresh soil with 50 mL of 2M KCl for 1 h. The concentrations of NO_3^- and NH_4^+ in the filtered extract were determined using Flow Injection Analysis (San^{++}, Skalar, the Netherlands) at RCEES, CAS.

Ceramic suction cup lysimeters (P80; Staatliche Porzellanmanufaktur, Berlin) and Macrorhizon soil moisture sampler (Rhizosphere Research Products, The Netherlands) were installed at each plot on HS and GDZ, respectively. On the HS, lysimeters were installed in triplicate in the A- (4–5 cm), the top of the AB- (10 cm), the bottom of the AB- (20 cm) and in the B-horizon (40 cm). In the GDZ, one Macrorhizon was installed at each of three depths (30, 60 and 100 cm). Soil water from HS was sampled from 11 November 2009, while groundwater in the GDZ was sampled from 5 August 2009; both soil water and groundwater were sampled at weekly intervals until spring 2010 and on each N_2O flux measurement date in summer 2010. Soil water from triplicate lysimeters per soil depth on HS was pooled prior to analysis. Concentrations of NO_3^- and NH_4^+ in lysimeter water were measured in each sample during wet-hot seasons and in pooled samples (four weeks) during the dry-cool season, using ion chromatography (DX-120 for NH_4^+ and DX-500 for NO_3^-, DIONEX, USA) at the Chongqing Academy of Environmental Sciences and Monitoring, China.

Soil bulk density (BD) of surface horizons/layers was measured from intact soil cores (100cc steel cylinders) sampled in triplicate at two depths (3.5–7.2 cm and 10–13.7 cm), corresponding to the A- and AB-horizons, respectively (Sørbotten, 2011).

Soil temperature (ST) and volumetric moisture (VM, $cm^3\,cm^{-3}$) at 10-cm depth were recorded every 10 min by TDR probes (Hydra Probe II) installed at plots T3 and B1 on 11 October 2009 and stored by a data logger (Campbell CR200). Data were obtained until 4 May 2011, with a malfunctioning period from 25 December 2010 to 8 February 2011. Soil water-filled pore space (WFPS) was calculated using bulk density (BD) at a depth of 10.0–13.7 cm, assuming a soil particle density (PD) of 2.65 g cm^{-3} (Linn and Doran, 1984) as

$$\text{WFPS}\ (\%) = \frac{\text{VM}}{1 - \frac{\text{BD}}{\text{PD}}} \times 100. \quad (1)$$

Groundwater level was monitored in PVC tubes, with a 30-cm nylon filter at the lower end, and installed at each plot along the B-transect in GDZ (B1 to B6). The PVC tubes were covered with a perforated lid to prevent pressure differences with the atmosphere. Groundwater level was measured daily from July 2009 to the beginning of August 2009 during a period of heavy rainstorms and on dates of N_2O measurements thereafter.

Meteorological data (air temperature (AT), precipitation (precip) and relative humidity (RH)) were obtained every 5 min from a weather station (WeatherHawk 232, USA), at the roof of the TSP Forest Bureau situated 1 km south of the catchment. Vapor pressure deficit (VPD, kPa) was calculated from equation 2 (ASCE, 2005) as

$$\text{VPD} = \left(\frac{\text{RH}}{100} - 1\right) \cdot 0.6108 \cdot \exp\left(17.27 \cdot \text{AT}/(\text{AT} + 237.3)\right) \quad (2)$$

2.4 Statistical analysis

Cumulative N_2O emissions for each of the eleven plots for summer 2010 (106 days) were analyzed together with soil parameters in the top soil (depth-weighted for O and A horizons on HS and from 0–20 cm in GDZ; Table 1) by Principal Component Analysis (PCA). Bulk density, cumulative N_2O flux, $NH_{4\,sw}^+$ and $NO_{3\,sw}^-$ were normalized using minus reciprocal, natural logarithmic, natural logarithmic and sinusoidal transformation, respectively.

To explore the seasonal drivers of the N_2O emission flux, log-transformed average N_2O emission rates on HS (average of 4 plots, T1, T3, T5 and B1, which had full N_2O flux datasets including the dry season) were subjected to stepwise multiple linear regression with ST, WFPS (average of daily ST and WFPS obtained from TDR probes at both T3 and B1 from 12 October 2009 onwards), as well as pore water $NO_{3\,sw}^-$ and $NH_{4\,sw}^+$ concentration in these four plots as independent variables. The resulting empirical relation was

used to interpolate the N_2O flux on HS between measurement dates on the basis of continuously measured ST and WFPS and to estimate the annual N_2O emission of two complete annual periods: one from 5 May 2009 to 4 May 2010 and the other one from 5 May 2010 to 4 May 2011.

Missing values for ST and WFPS (from 1 May to 11 October 2009 and from 25 December 2010 to 8 February 2011) were estimated on a daily basis, using a statistical model based on multiple linear regression, using available meteorological data (average air temperature, rainfall and average VPD). General linear model (GLM) with the mixed-effects (sampling dates and location) was used for the comparison of different plots, the two landscape elements and different seasons.

All statistical analyses were conducted with Minitab 16.1.1 (Minitab Inc.).

3 Results

3.1 Weather, soil moisture and soil temperature

Rainfall distribution during the wet season (April to October) differed between the two years (Fig. 2). Summer 2009 (1054 mm) was wetter than summer 2010 (850 mm) and precipitation occurred mostly as intensive rainstorms (up to 385 mm during 41 h between 3 and 5 August 2009). In contrast, precipitation was more evenly distributed in summer 2010. Only 117 mm rain fell in the dry-cool season from November 2009 to March 2010. Soil WFPS at 10-cm depth on the mid slope (T3, mean WFPS 55.8 %) was significantly smaller than at the foot slope of HS at the transition to GDZ (B1, mean WFPS 70.0 %) (Fig. 3a), most likely reflecting convergence of interflow from HS at B1. In winter, WFPS was relatively stable with values around 56.7 % for T3 and 68.5 % for B1. In contrast, large variations in WFPS were observed in spring and summer, driven by precipitation, and subsequent drainage and evapotranspiration. In plot T3 (HS), but not in plot B1 (transition to GDZ), WFPS decreased significantly in late summer 2010 (July through September), indicating drier conditions on the upper part of HS as described earlier for this site (Mulder et al., 2005). ST at 10-cm depth varied seasonally between 10 °C and 25°C with no significant difference between T3 and B1, although the drier T3 plot tended to have somewhat higher soil temperatures during the wet season (Fig. 3b).

The groundwater level in GDZ varied between +15 cm (above surface) and < −100 cm (below surface), with highest values at B5 and lowest at B1 (Fig. 4). The heavy rainstorm between 3 and 5 August 2009 resulted in a rapid increase of the groundwater level. The groundwater level reached its lowest values from late summer onwards, but increased in early spring, due to increased precipitation in March and April 2010. In general, the groundwater level was deepest at B1 and increased to near-surface in the order

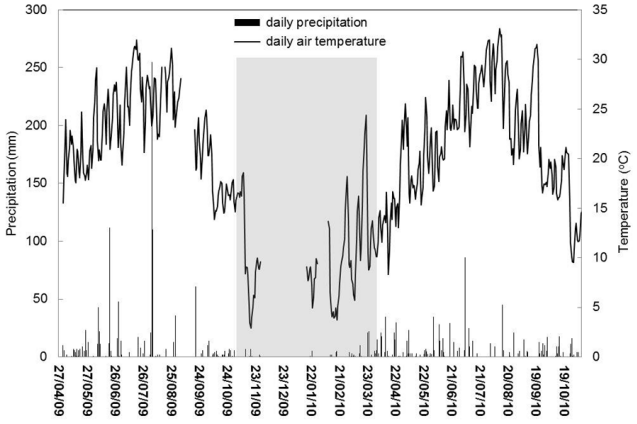

Fig. 2. Daily average air temperature and daily precipitation at TSP. The temperature data from 5 September to 18 September 2009, 29 November 2009 to 16 January 2010 and 29 January to 8 February 2010 were missing. Shaded area indicates the dry-cool season.

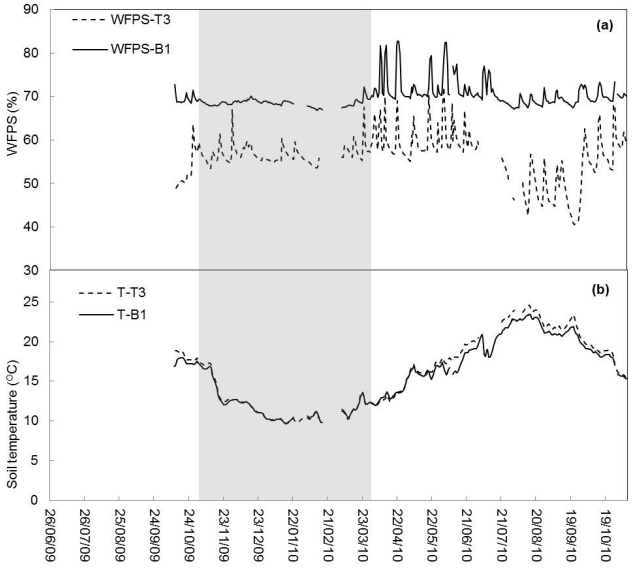

Fig. 3. (a) Soil water-filled pore space (WFPS) and **(b)** soil temperature (ST) at 10-cm depth at plots T3 and B1. Shaded area indicates the dry-cool season.

B1 < B2 = B3 = B4 < B5 ≤ B6 and was most stable close to the outlet (plots B5 and B6). The groundwater level was only occasionally detectable (> −100-cm depth) at plots B1, B2 and B3 (Fig. 4).

3.2 N_2O fluxes

Emission fluxes of N_2O showed a pronounced seasonal pattern with highest values during summer (wet-hot season) and significantly lower emission rates during winter (dry-cool season) ($p = 0.000$, Fig. 5). During summer, N_2O emission rates varied 1–2 orders of magnitude, with highest values

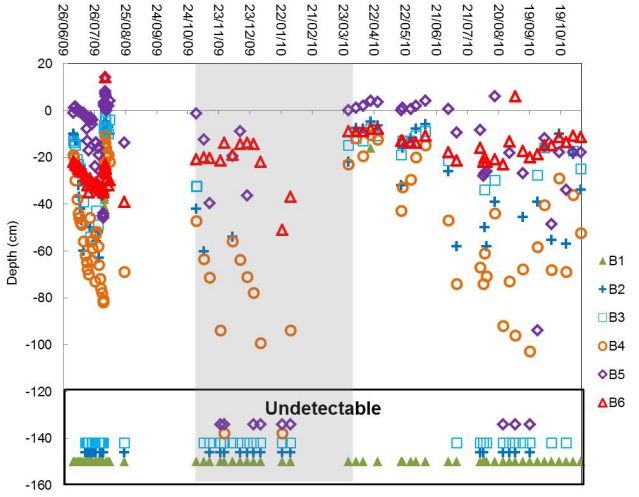

Fig. 4. Groundwater level (GWL) in GDZ. The lower panel indicates plots and dates at which the groundwater level was below the bottom of the monitoring well. Shaded area indicates the dry-cool season.

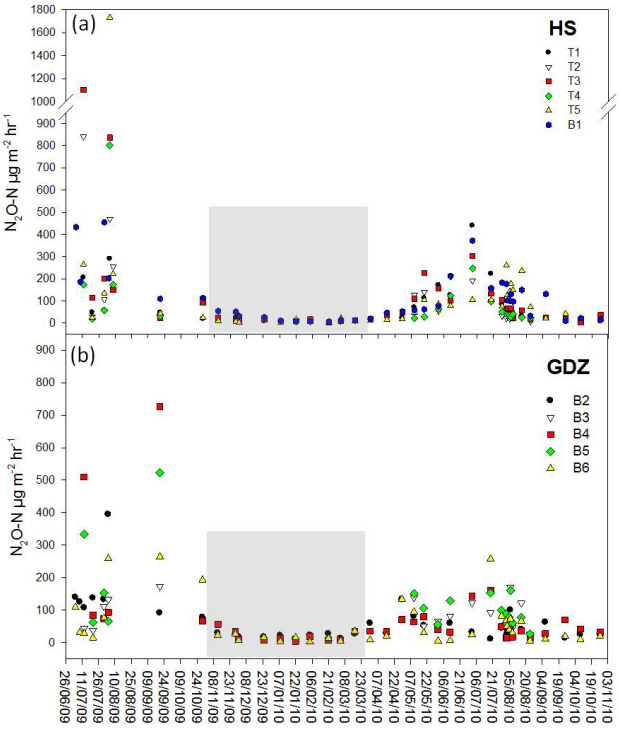

Fig. 5. (a) Mean N_2O flux ($n = 3$) at the plots on the hillslope (HS) and **(b)** in the groundwater discharge zone (GDZ). Standard errors are not shown to maintain readability of the figure. Shaded area indicates the dry-cool season.

immediately following precipitation events (compare Figs. 2 and 5) when WFPS was high (Fig. 3a). In summer 2010, with its more even distribution of rainfall, peak N_2O emissions were smaller (up to 450 µg m^{-2} h^{-1}) than in summer 2009 (up to 1730 µg N m^{-2} h^{-1}) (Fig. 5). The analysis with a general linear model (GLM) indicated that N_2O emission rates were significantly higher on HS than in GDZ ($p = 0.011$). The daily average N_2O emission on HS was higher in summer 2009 (6.4 mg N m^{-2} d^{-1}) than in summer 2010 (3.2 mg N m^{-2} d^{-1}; Fig. 6). Furthermore, the average N_2O emission was smaller in GDZ than on HS, emitting on average 3.2 mg N_2O-N m^{-2} d^{-1} and 1.8 mg N_2O-N m^{-2} d^{-1} in 2009 and 2010, respectively. The spatial variability within the plots was high particularly on HS (Fig. 6) and no significant differences in N_2O emissions were found between plots within the same landscape element. The spatial pattern of the cumulative N_2O emissions for the plots within one transect was not consistent for the two summer periods (Fig. 6).

3.3 Spatial variability in soil factors controlling N_2O emissions

Surface soil had a greater bulk density in GDZ than on HS (Table 1) even though there were exceptions (plots B4 and B5). The pH of the surface soil was significantly higher in GDZ than on HS (Table 1). For TOC, TN and C/N the opposite was the case, reflecting the lack of an organic surface layer in GDZ. Values for the concentration of NO_3^- in soil extracts ($NO_{3\,ex}^-$) and the concentration of NO_3^- in soil water ($NO_{3\,sw}^-$) were significantly higher on HS than in GDZ. In the O-horizon, TOC and TN were higher at T2 and T4 than at other plots on HS. In the GDZ, Plot B6 was highest with respect to these variables. The $NO_{3\,sw}^-$ was high in winter and early spring and decreased sharply at the start of the growing season in early April, probably due to biological uptake of N and dilution by ample rainfall at the end of March 2010 (Fig. 7). During winter, $NO_{3\,sw}^-$ showed consistent differences among plots on HS with highest values at plot T1 and lowest at plot T2, whereas concentrations were lower and more uniform throughout summer, fluctuating with precipitation events. There were consistent differences in $NO_{3\,sw}^-$ in GDZ, with concentration levels decreasing in the order B2 = B4 > B3 > B5 ≥ B6. At B3, B5 and B6, $NO_{3\,sw}^-$ concentrations were below 5 mg N L^{-1} during most of the year with the greatest values at the end of the dry season (February and March 2010). The concentration of NH_4^+ in soil water ($NH_{4\,sw}^+$) was small, with values amounting to less than 10 % of total inorganic N in solution, except for plots B2 and B6, and there was no seasonal pattern in any of the plots (data not shown). No correlation was found between $NH_{4\,sw}^+$ and $NO_{3\,sw}^-$ in soil water.

We used PCA to assess the relationship between soil parameters and the cumulative N_2O flux in summer 2010 (Fig. 6) at each plot (Fig. 8). The PCA divided the plots into two groups (T1 to T5 and B1 as one group (HS); the others as the second group (GDZ)). The first two components of PCA explained 63.9 % of the total variation, with most

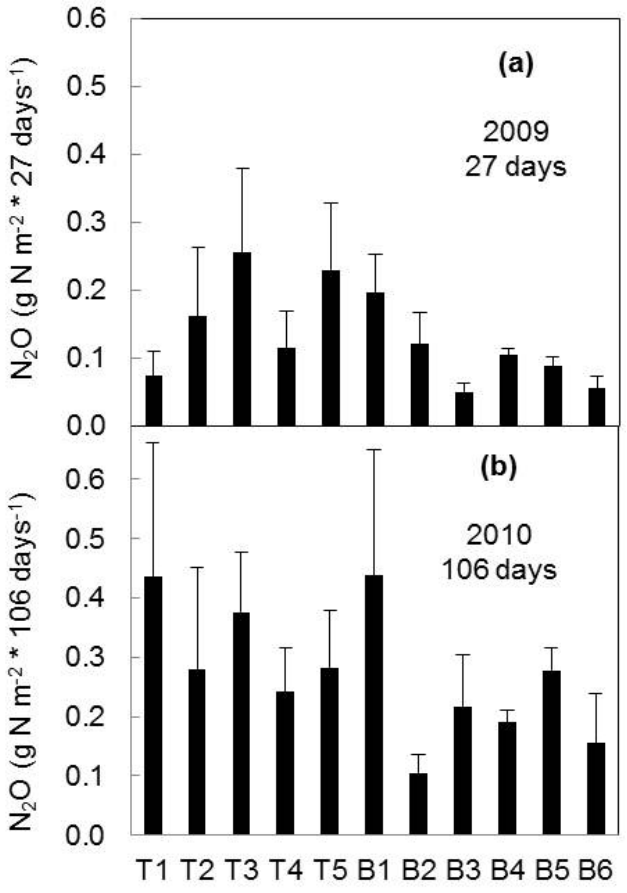

Fig. 6. **(a)** Cumulative N_2O fluxes ($g\,N_2O\text{-}N\,m^{-2}$) during summer 2009, from 12 July to 8 August, 27 days) and (**(b)** 2010, from 10 May to 24 August, 106 days) at individual plots ($n = 3$). Error bars represent standard error.

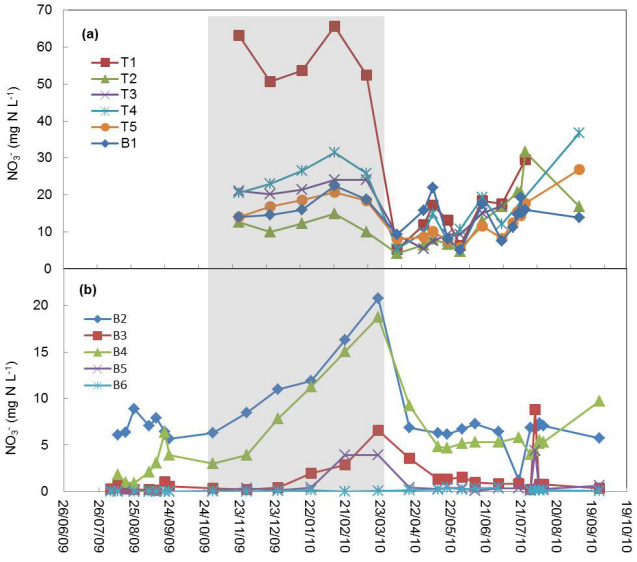

Fig. 7. **(a)** NO_3^- concentration ($mg\,N\,L^{-1}$) in soil pore water at 5-cm depth at each plot of HS and **(b)** of groundwater at 30-cm depth in GDZ. NO_3^- values from some of the HS plots during summer 2009 are missing because samplers were damaged by worms. Shaded area indicates the dry-cool season.

of the physicochemical parameters (pH, bulk density, TOC, TN, $NO_{3\,ex}^-$, $NO_{3\,sw}^-$ and $NH_{4\,ex}^+$) and the ex situ potential for N_2O loss contributing to the first component (50.3 %). The second component (13.6 %) included $NH_{4\,sw}^+$. The instantaneous denitrification rate contributed to both components, but had relatively small effect on the N_2O flux compared to other parameters. Total organic carbon (TOC) and most of the variables associated with the availability of N (TN, $NO_{3\,ex}^-$, $NO_{3\,sw}^-$ and $NH_{4\,ex}^+$) were positively correlated with soil N_2O flux, whereas soil pH and bulk density (BD) were negatively correlated with N_2O flux. Also, the instantaneous denitrification rate and the ex situ potential for N_2O loss were positively correlated with N_2O flux, whereas $NH_{4\,sw}^+$ concentration in soil water did not correlate with N_2O flux.

3.4 Temporal variability of seasonal drivers for N_2O emission

In summer, peak N_2O emission was associated with rainfall resulting in high WFPS values. Emission of N_2O on HS was highest at WFPS values between 63 % and 72 % (averages at 10 cm soil depth for T3 and B1, respectively) and appeared somewhat lower at values larger than 75 % (Figs. 3a and 5). Figure 5 illustrates the large interannual variability of N_2O flux between a wet year with intensive rainfall episodes (2009) and a relatively dry year (2010) with only one episode with precipitation $> 50\,mm\,day^{-1}$ (82 mm in 9 h on 5 July 2010). The N_2O flux on HS was weakly, but negatively correlated with $NO_{3\,sw}^-$ ($p = 0.064$), whereas no correlation was found with $NH_{4\,sw}^+$ ($p = 0.659$). At individual sites in GDZ, the N_2O flux was weakly, but negatively correlated with the groundwater level, i.e. higher N_2O flux occurred when the groundwater level was low. Low ST together with low WFPS seemed to be the main constraints for N_2O emission activity in winter (Figs. 3 and 5). Stepwise multiple linear regression indicated that ST and WFPS were the only significant variables explaining temporal variability of N_2O emissions (Eq. 3) on HS; $NO_{3\,sw}^-$ and $NH_{4\,sw}^+$ concentrations did not add significantly and were excluded by the regression. The residuals of the simulated values were normally distributed (not shown), suggesting unbiased fitting of the equation (standard errors of the coefficients in parenthesis):

$$\ln\,(N_2O_{HS}) = -4.80(\pm 1.24) + 0.088(\pm 0.018)\ \mathrm{WFPS} + 0.178(\pm 0.024)\ \mathrm{ST}. \quad (3)$$

Adjusted coefficient of determination:

$(R - Sq\,(\mathrm{adj})) = 68.9\,\%$

Since no data for ST and WFPS were available before 11 October 2009 and from 25 December 2010 to

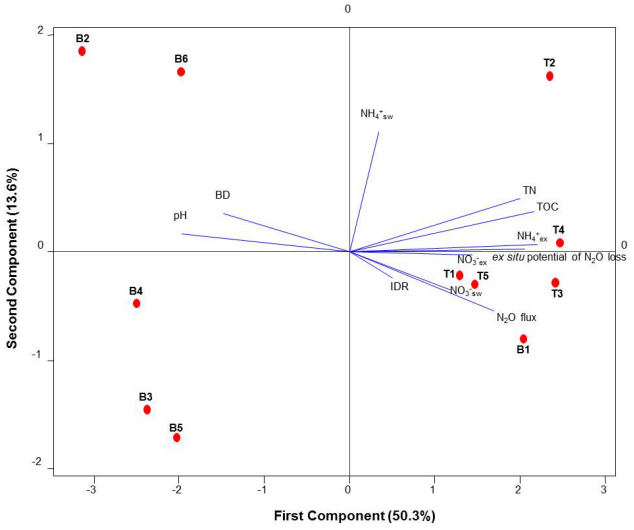

Fig. 8. Principal component analysis of cumulative N_2O fluxes and soil parameters (Table 1) at TSP. Before analysis, the N_2O flux and several of the parameters were transformed, whereas others (TOC: total organic carbon; TN: total nitrogen; C/N: the ratio of TOC and TN; IDR: instantaneous denitrification rate; ex situ potential of N_2O loss (see text for explanation); NH_4^+ ex and NO_3^- ex: 2M KCl -extractable NH_4^+ and NO_3^-) were not. The transformations were as follows: N_2O flux: ln(cumulative N_2O flux from each plot in summer 2010; Fig. 6); bulk density (BD): -1/(bulk density); NH_4^+ sw: ln(dissolved NH_4^+) in soil water; NO_3^- sw: sin(dissolved NO_3^- in soil water). Red points indicate the individual plots.

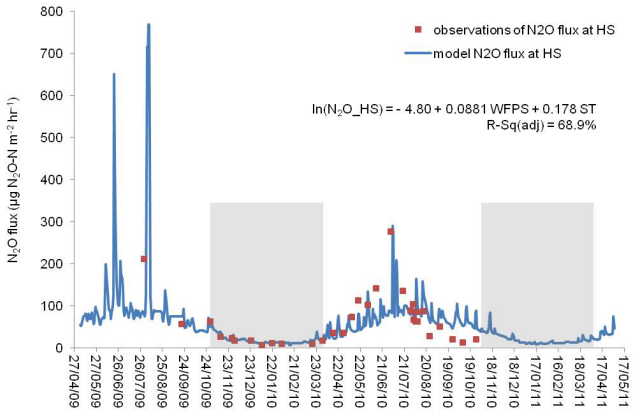

Fig. 9. Observed and simulated average N_2O flux at hillslope (HS). The shaded area denotes the dry-cool season.

8 February 2011, we used multiple linear regression to derive equations that link ST and WFPS to meteorological data. The resulting regression models, presented in Eqs. (4) and (5) were then used to estimate the missing values (standard errors of the coefficients in parenthesis):

$$\mathrm{ST} = 7.84(\pm 0.41) + 0.579(\pm 0.029)\ \mathrm{AT} + 1.20(\pm 0.187)\ \mathrm{VPD}. \quad (4)$$

Adjusted coefficient of determination:
$(R - Sq(\mathrm{adj})) = 67.4\,\%$.

$$\mathrm{WFPS} = 63.2(\pm 0.27) + 0.209(\pm 0.014)\mathrm{precip}_5 + 1.84(\pm 0.17)\mathrm{VPD}_5 \quad (5)$$

Adjusted coefficient of determination:
$(R - Sq(\mathrm{adj})) = 58.8\,\%$.

Where precip_5 (mm) is the day-weighted 5-day average precipitation prior to the date for which the computation was done (calculated as $1.0 \cdot \mathrm{precip}_{\mathrm{day}(0)} + 0.8 \cdot \mathrm{precip}_{\mathrm{day}(-1)} + 0.6 \cdot \mathrm{precip}_{\mathrm{day}(-2)} + 0.4 \cdot \mathrm{precip}_{\mathrm{day}(-3)} + 0.2 \cdot \mathrm{precip}_{\mathrm{day}(-4)}$) and VPD_5 (kPa) the day-weighted 5-day average vapor pressure deficit calculated analogously. The selected weights fitted the observations best. The residuals of the model simulations were normally distributed (not shown). Equation (3) was used to estimate N_2O emissions using measured and simulated values for ST (Eq. 4) and WFPS (Eq. 5). Measured and simulated values are shown in Fig. 9.

3.5 Annual N_2O flux

Annual N_2O emissions on the HS were calculated at a daily time step from equation 3 for two 1-yr periods (5 May 2009 to 4 May 2010 and 5 May 2010 to 4 May 2011) (Fig. 9). The first annual period included a relatively wet summer, whereas the second period had a relatively dry summer. Annual N_2O fluxes for HS were 0.54 and 0.43 g N m^{-2} for the two years, respectively, about 24 % higher in the first year than in the second. Higher annual N_2O emission in the first year may be attributed to the contribution of the frequent rain episodes in the wet summer of 2009. The N_2O emission occurring during the first 5 days after the largest rainfall episode (4 to 5 August 2009) contributed 15.8 % to the annual N_2O emission. Emissions during the dry-cool season in the first year (November 2009 to March 2010) contributed 12.0 % to the annual flux of N_2O. Also the emissions during the dry-cool season in the second year (November 2010 to March 2011), solely based on model estimates, was found to be low, contributing 16.9 % to the annual flux of N_2O.

4 Discussion

N_2O emission rates recorded after the heavy rainstorm in 2009 (up to 1730 µg N_2O-N m^{-2} h^{-1}; Fig. 5) are the highest found for forest ecosystems so far and exceed rates reported for (sub)tropical forests which typically vary between 11 to 600 µg N_2O-N m^{-2} h^{-1} (Liu et al., 2011a; Zhang et al., 2008; Silver et al., 2005; Ishizuka et al., 2005; Koehler et al., 2009; Rowlings et al., 2012; Kiese and Butterbach-Bahl, 2002).

Temporal variability (Fig. 5) exceeded spatial variability in both landscape elements (HS and GDZ). However, small-scale (within-plot) variability of the cumulative N_2O flux over the active season was high as seen from the high

standard errors in Fig. 6 (particularly for HS). It is likely that this small scale variability was blurring existing trends in flux strength between plots along the HS. Large small-scale spatial variability of N_2O emissions is often reported for forest soils (Robertson and Klemedtsson, 1996; Bowden et al., 1992; Werner et al., 2007b) and is commonly attributed to small-scale variation in mineral N availability, litter quality and soil moisture. We arranged the plots along transects perpendicular to the contour lines in the two landscape units because gradients in WFPS on HS and in groundwater level in GDZ might affect N_2O emissions by controlling aeration, substrate level and ultimately microbial activity (Garten et al., 1994). However, we did not find any consistent relationship between cumulative N_2O emission and plot position in either of the two landscape elements. Concentrations of $NO_{3\,sw}^-$ on HS were similar at all plots during the wet-hot season (Fig. 7a) and did not correlate with N_2O emissions. The availability of NO_3^- was higher than in other studies (e.g., Koehler et al., 2009; Zhang et al., 2008) and appeared to be non-limiting for N_2O production on the HS, probably explaining the lack of a directional gradient in cumulative N_2O emissions as reported by others (Nishina et al., 2009; Fang et al., 2009). Surprisingly, the clear difference in WFPS between the mid slope position T3 and the foot slope position B1 (Fig. 3a) in spring and early summer did not result in significantly different cumulative N_2O emissions (Fig. 6). This was probably due to the fact that soil moisture was measured at 10-cm depth (in the upper AB-horizon), rather than in the zone of maximum denitrification activity, located in the uppermost soil layers (O- and A-horizons; Zhu et al., 2013a). Soil moisture data records for 5-cm depth were incomplete due to sensor problems, but comparing existing data and bulk densities at 5-cm depth suggests that WFPS was more similar in the uppermost soil layer (data not shown). The pronounced decrease in WFPS at T3 (but not B1), later in summer, markedly reduced N_2O emissions as compared with B1 (insert Fig. 5).

In GDZ, decreasing NO_3^- concentrations along the hydrological flow path (Fig. 7b) suggested strong N retention. Both NO_3^- and the dissolved N_2O (data not shown) concentrations in the groundwater were found to be very low at the outlet of GDZ (B5 and B6), suggesting more complete N_2O reduction along the flow path as NO_3^- became depleted. There was a weak trend of decreasing cumulative N_2O emissions with increasing average groundwater level along the hydrological flow path in the GDZ (with the exception of B3) during summer 2009, when the groundwater level was very dynamic (Fig. 6a). No such pattern was seen in the hydrologically more stable summer of 2010 (Fig. 6b). In general, groundwater level fluctuations are likely to interact with NO_3^- in controlling N_2O emissions, as the rise of groundwater level can transport NO_3^- into C-richer upper soil layers and create anoxia favorable for denitrification. At the same time, the residence time of N_2O in the soil profile increases, due to decreasing diffusion, thus increasing the chance of N_2O being reduced to N_2. Conversely, a drop in groundwater level may relieve gaseous diffusion constraints leading to release of N_2O from the soil. In summer 2010 (Fig. 5), for instance, N_2O emission rates were lowest at B6 until groundwater level dropped about 10 cm in the middle of July (Fig. 4), resulting in a single N_2O emission peak while returning to low levels quickly thereafter. Similar N_2O dynamics have been observed in rice paddies (e.g. Qin et al., 2010).

We found clear differences in N_2O source strength and environmental controls when comparing HS and GDZ; average cumulative N_2O emission in GDZ was roughly half of this at HS in both summers (Figs. 5 and 6), which was surprising, as we expected GDZ to behave as a riparian zone with elevated denitrification activity. Riparian zones are commonly considered as "hot spots" for N_2O emissions, because they receive dissolved N (DON, NO_3^-) from surrounding slopes while providing optimal conditions for denitrification (DOC, anoxia) (Groffman et al., 2000; Hefting et al., 2003). However, unlike common riparian zones which typically develop soils rich in organic matter, the GDZ studied here is formed by formerly cultivated terraces on colluvial deposits. Soils are dense and have low hydrological conductivities, resulting in low productivity and low TOC contents (Table 1). Laboratory incubations revealed that the GDZ soils had instantaneous denitrification rates similar to HS soils, despite their markedly lower TOC contents (Zhu et al., 2013a). This indicated a similar N removal potential in GDZ as compared with HS. However in their study, Zhu et al. (2013a) also found that denitrifying communities on HS and in GDZ differed functionally with respect to their product stoichiometry, with the GDZ communities being more efficient in reducing NO_3^- all the way to N_2. Thus, the lower N_2O/N_2 ratio, and consequently the reduced N_2O emission rate, in GDZ may be attributed to a number of site-specific factors, directly or indirectly affecting denitrifier functioning and stoichiometry: (1) pore water at GDZ had lower NO_3^- concentrations (Figs. 7b and 8), which may decrease the N_2O/N_2 ratio by forcing denitrifiers to utilize N_2O as an additional electron acceptor; (2) diffusion of N_2O in soil is slow in dense GDZ soils, resulting in higher dissolved N_2O concentrations in pore water (data not shown), which supports denitrifier communities to express N_2O reductase; and (3) soil pH at GDZ was 0.5–0.6 units higher than at HS, which directly (Bergaust et al., 2010) or indirectly (Liu et al., 2010a; Simek and Cooper, 2002; Dörsch et al., 2012) lowers N_2O/N_2 ratios.

Even though this study was not designed to discriminate sources of N_2O emission, several observations indicate that N_2O emission rates during the active periods were more sensitive to denitrification controls than nitrification controls: (1) on the HS, high N_2O emission occurred when soil WFPS was high; (2) in the GDZ, dissolved N_2O decreased with NO_3^- concentration in groundwater; (3) the difference in magnitude in N_2O flux on HS and in GDZ was congruent to differences in denitrification stoichiometry observed in soils from

the two landscape elements in laboratory incubations (Zhu et al., 2013a); (4) N_2O concentrations measured in soil air were negatively correlated with pO_2 ($R^2 = 0.40$, $p = 0.000$, data not shown). Low nitrification rates have been reported for acidic forest soils in SW China (Cai and Zhao, 2009). However, given the prevalence of NH_4^+ in deposited N and its rapid disappearance in the soil solution, NH_4^+ oxidizing processes could be a potentially important source for N_2O emission. We conducted an in situ $^{15}NO_3^-$ labeling experiment in summer 2010 at two positions along the HS reported elsewhere (Zhu et al., 2013b) and found that 71–100 % of the emitted N_2O derived from the labeled soil NO_3^- pool. It seemed that even though the fate of the deposited NH_4^+ remains unclear in this catchment, nitrification is not the main source for the high N_2O emission observed.

We found large temporal variability in N_2O fluxes. Multiple linear regression identified ST and WFPS as the main drivers, explaining 68.9 % of the temporal variability of N_2O emission fluxes on HS. Such a high coefficient of determination with only two variables is remarkable as most studies conducted in forest ecosystems report much lower degree of explanation by ancillary variables (Morishita et al., 2011; Gu et al., 2011). As shown in Fig. 9, the resulting regression model captured reasonably well both the seasonal distribution of emission fluxes and the transient emission peaks after rainfall events on HS except for the first measurement point. Peak N_2O emissions (Fig. 5) were triggered by spikes in WFPS (Fig. 3a) occurring after big rain episodes, resulting in a highly skewed distribution of N_2O fluxes as found in many other studies (Khalil et al., 2007; Venterea et al., 2009). Low ST, together with relatively low WFPS throughout winter resulted in low N_2O fluxes. This implies that N_2O emission fluxes can be estimated fairly well from discontinuous flux measurements when continuous data on ST and WFPS are available. However, it is important to sample fluxes closely around major rain fall events in summer as they contribute disproportionately to overall emissions. Overly long intervals between samplings may result in over- or underestimation of the cumulative flux (Parkin, 2008).

Annual N_2O emissions for the TSP catchment were estimated for the HS only, since 96.4 % of the area consists of hillslopes, rendering GDZ emissions a minor component of the overall N_2O budget. Annual fluxes showed considerable variation between the wet year (May 2009 to April 2010) with several intensive rain episodes ($0.54\,g\,N_2O\text{-}N\,m^{-2}\,yr^{-1}$) and the dryer year (May 2010 to April 2011) with more evenly distributed precipitation ($0.43\,g\,N_2O\text{-}N\,m^{-2}\,yr^{-1}$). The average annual N_2O emission flux in TSP ($0.48\,g\,N\,m^{-2}\,yr^{-1}$) is at the high end of annual N_2O emission fluxes ($0.005–0.47\,g\,N\,m^{-2}\,yr^{-1}$) reported for unfertilized forestland in China (Cai, 2012). The fluxes are even comparable to the high end of reported N_2O fluxes in tropical rainforests where temperatures are high throughout the year (Dalal and Allen, 2008) and to a tropical lowland forest in Panama which was experimentally N-enriched for 9–10 years (Koehler et al., 2009). Expressed as a fraction of the annual atmogenic N-deposition, N_2O-N emission amounted to between 8 and 10 % of the incoming inorganic N. This "emission factor" was much higher than the Tier 1 IPCC default factor (1 %) used for accounting both direct N_2O emissions arising from mineral N fertilizer application to managed soils and indirect N_2O emissions arising from the redeposited N to uncultivated soils (IPCC, 2006). The emission factor derived from our study is at the high end of the reported N_2O emission factors for deciduous forests and coniferous forests, which were on average 6.5 % and 2.4 %, respectively (van der Gon and Bleeker, 2005). This pinpoints the prominent role of nitrogen-saturated, subtropical forests on acid soils as a secondary source of anthropogenic N_2O, which should be taken into account for regional N_2O budgets.

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