Investigating unproductive water losses from irrigated agricultural crops in the humid tropics through analyses of stable isotopes of water

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Abstract. Reliable information on water flow dynamics and water losses via irrigation on irrigated agricultural fields is important to improve water management strategies. We investigated the effect of season (wet season, dry season), irrigation management (flooded, non-flooded), and crop diversification (wet rice, dry rice, and maize) on soil water flow dynamics and water losses via evaporation during plant growth. Soil water was extracted and analysed for the stable isotopes of water (δ^2H and $\delta^{18}O$). The fraction of evaporation losses were determined using the Craig–Gordon equation. For dry rice and maize, water in shallow soil layers (0 to 0.2 m) was more isotopically enriched than in deeper soil layers (below 0.2 m). This effect was less pronounced for wet rice, but still evident for the average values at both soil depths and seasons. Soil water losses due to evaporation decreased from 40 % at the beginning to 25 % towards the end of the dry season. The soil in maize fields showed stronger evaporation enrichment than in rice during that time. A greater water loss was encountered during the wet season, with 80 % at the beginning of the season to 60 % at its end. The isotopic enrichment of ponding surface water due to evaporation was reflected in shallow soils of wet rice. It decreased towards the end of both growing seasons during the wet and the dry season. We finally discuss the most relevant soil water flow mechanisms, which we identified in our study to be that of matrix flow, preferential flow through desiccation cracks, and evaporation. Isotope data supported the fact that unproductive water losses via evaporation can be reduced by introducing dry seasonal crops to the crop rotation system.

1 Introduction

Soil water studies are essential for a better understanding of the role soils play in the hydrological cycle, to estimate the water budget and water availability for plants, groundwater recharge, other organisms as well as solute transport. Stable isotopes of

water (δ^2H and δ^{18}O) as natural tracers have become a powerful tool for such studies (Kendall and Caldwell, 1999) They are particularly helpful to better understand the evaporation dynamics in soil water (Braud et al., 2009; Kool et al., 2014; Rothfuss et al., 2015) because the composition and distribution of stable isotopes of water in a soil profile provide insight into evaporation fractionation and water flux processes (Wenninger et al., 2010).

5   The determination of soil evaporation and the fraction of evaporation in relation to total evapotranspiration have been widely studied using several methods for different crops. For example, Liu et al. (2002) studied evapotranspiration from winter wheat and maize, using weighing lysimeters. Zhou et al. (2016) partitioned evaporation and transpiration fluxes for corn, soya bean, grassland, and forests using flux tower measurements. Kool et al. (2014) applied different methods such as chamber measurements, micro-lysimeter, and soil heat pulse to estimate the evaporation. They used stable isotopes of water to separate

10 evaporation from transpiration. Soil isotopic profiles can be subdivided into two parts (Barnes and Allison, 1984), first, the upper part in shallow soil, in which water moves by vapour diffusion and which is affected by evaporation, and second, the lower part in deep soil, where direct flows take place and which is barely affected by diffusion. However, the isotopic composition of soil water is not only affected directly by evaporation, mixing of new and old water (Gazis and Feng, 2004) and altering input signals (Barnes and Turner, 1998), for example through infiltration of rainwater, but also indirectly by other

15 processes such as transpiration (Barnes and Allison, 1988), water transport (Kutilek and Nielsen, 1994; Melayah et al., 1996), and hydrodynamic dispersion (Wang et al., 2017). Transpiration can selectively remove soil water of a specific recharge period and isotopic composition. During dry periods, the isotopic enrichment of shallow soil water is generally driven by evaporation (Gangi et al., 2015; Liu et al., 2015) and is affected by equilibrium and kinetic fractionation (Gat, 1996; Gonfiantini, 1986). Due to this complexity, many experiments on the effects of evaporation on soil water using isotope methods are often restricted

20 to the laboratory-scale or short-term field studies or one particular location (Beyer et al., 2016; Gaj et al., 2016; Oerter and Bowen, 2017; Rothfuss et al., 2015; Sprenger et al., 2017; Twining et al., 2006; Volkmann et al., 2016).

   Studying water fluxes in rice-based cropping systems is essential, because rice (Oryza sativa L.) is the dominant staple food for nearly half of the world's population, yet water resources are limited. More than 80 % of the global rice production area is located in Asia (Kudo et al., 2014). Rice is one of the highest water-consuming grain crops (Janssen and Lennartz, 2007;

25 Mekonnen and Hoekstra, 2011), consuming approximately 30 % of all freshwater resources worldwide (Maclean et al., 2002). Since rice is susceptible to water shortages (Bouman and Tuong, 2001), 80 % of rice in Asia is cultivated under conventional flooded conditions (Towprayoon et al., 2005); it is also called wet rice, anaerobic rice, or lowland rice. Water scarcity is a severe environmental problem, especially in the irrigation of agricultural land (Pfister et al., 2011). Therefore, water-saving strategies need to be developed to secure rice production (Belder et al., 2004). Introducing non-flooded crops during the dry

30 season (e.g., maize or non-flooded rice, also called dry rice, aerobic rice or upland rice) is an interesting alternative and has been increasingly applied in food and fodder production in Southeast Asia (FAO, 2016; Timsina et al., 2010). To establish efficient water-saving management based on crop rotation and season, a functional understanding of hydrological processes of these new rice-based cropping systems is required (Daly et al., 2004; Heinz et al., 2013; Zwart and Bastiaanssen, 2004).

Understanding water flow dynamics and unproductive water losses from irrigated soils is still incomplete, particularly for rice-based cropping systems. Unproductive water losses are those that unlike transpiration do not lead directly to biomass production, and include, for example, leaching, evaporation from the soil, or ponding water (Bouman, 2007). Studies on the effects of evaporation, its seasonal variability, as well as the impact of various crop rotations, are still missing. None of the studies conducted so far have quantified the fraction of evaporation losses in rice-based cropping systems, taking into account the effect of crop species and various growing stages. The objectives of this study are, therefore (I) to investigate soil water isotopic profiles to study the impact of crop species (wet rice, dry rice, and maize) and growing stages on evaporation during the wet and dry season; (II) to understand flow mechanisms of soil water in the soil matrix of agricultural soils; and (III) to quantify the fraction of evaporation losses from agricultural fields based on stable isotopes of water.

2 Material and Methods

2.1 Site description and experimental design

The field trial was established at the experimental station of the International Rice Research Institute (IRRI), in Los Baños, Laguna, Philippines (14°11'N, 121°15'E, 21 m a.s.l.). The experiment was conducted during WS 2015 and DS 2016. The average total precipitation during this period was 1,700±50 mm during the wet season (WS, June to November), and 300±25 mm during the dry season (DS, December to May). The mean seasonal temperature and relative humidity were 28.5±0.9°C and 83±6 % during the WS 2015, as well as 27.6±1.8°C and 74±11 % during the DS 2016, respectively. Climate data were obtained from the climate unit at IRRI. Both seasons represented typical weather conditions in the region. The soil type in the study area is classified as a Hydragric Anthrosol (He et al., 2015) with clay-dominated soil texture (Table 1). The clay fraction mainly consists of vermiculite and smectite as three-layer clays and kaolinite as a two-layer clay. Three-layer vermiculite is primarily responsible for the swelling and shrinking of the soil matrix (Tertre et al., 2018).

The experimental design (Fig. 1) consisted of nine fields (3 wet rice–wet rice, 3 wet rice–dry rice, 3 wet rice–maize) with an average field size of about 540 m², each split into three plots with different treatments. Of these plots, only those with straw application and the control plots (without straw) were used for our experiment. Straw was not applied as a typical mulch layer to reduce evaporation but was partly worked into the soil to reduce crack formation during dry soil conditions and resulting in preferential flow losses. During the WS, all nine fields were cropped with wet rice (cultivar NSIC Rc222). During the DS, three fields each were cultivated with wet rice, dry rice (cultivar NSIC Rc192), and maize (Pioneer P3482YR). Wet rice fields were maintained at water-flooded conditions, except for the first and last two weeks between transplanting and harvest (Fig. 2). Dry rice and maize fields were only irrigated when weather conditions suggested a water shortage (i.e., 5–10 times during the growing season for maize fields). Field workers from the IRRI were responsible for watering dry crops in times of soil water shortage. The decision of watering was not set by specific thresholds or indicators, but by expert knowledge. The total irrigation amount for wet rice fields was 470±50 mm during the WS, and 1,270±300 mm, 517±50 mm, and 212±50 mm for wet rice, dry rice, and maize during the DS, respectively. Note that irrigation water was taken from an open reservoir, located

next to the fields. The reservoir is regularly filled with groundwater that is characterized by a uniform seasonal composition with an isotopically depleted characteristic (Mahindawansha et al., 2018a). Transplanting and harvesting dates for rice were July 21st and October 30th during the WS. During the DS, the transplanting date was January 8th, and harvesting dates were April 10th for wet rice and April 17th for dry rice, and January 6th and May 11th for maize in 2016, respectively (Fig. 2).

2.2 Soil and root sampling

Samples were collected during the three main growing stages (GS) described by Counce et al. (2000), i.e., at the vegetative stage (GS1, from germination to panicle initiation), the reproductive stage (GS2, from panicle initiation to flowering), and the ripening stage (GS3, from flowering to maturity). The growing stages were used as reference points during the growing season (Fig. 2). Growing stages for rice and maize were assumed to be similar to maintain consistency of sampling conditions. Samples were taken on one day during each growing stage: 26, 55, 85 days after transplanting during the WS, and 40, 60, 90 days after transplanting during the DS, respectively. Soil cores were taken using a manual soil corer (length=0.6 m, diameter=0.05 m). Each core was divided into 9 depth intervals from the surface to 0.6 m (0, 0.05, 0.1, 0.15, 0.2, 0.2–0.3, 0.3–0.4, 0.4–0.5, 0.5–0.6 m). Altogether, 972 samples were taken (9 fields x 2 treatments x 2 seasons x 3 growing stages x 9 soil depths). A plastic ring (diameter=0.5 m) was used to drain the water around the sampler prior to coring in wet rice fields. Samples were stored in sealed aluminium bags (CB400–420BRZ, 80 mm x 110 mm, Weber packaging, Güglingen, Germany) and immediately placed in an ice-filled Styrofoam box for transfer to the laboratory where they were kept frozen.

Soil water was extracted from soil aliquots (10–15 g of the sample) via cryogenic vacuum extraction (Orlowski et al., 2013) at the Institute for Landscape Ecology and Resources Management (Justus Liebig University Giessen, Germany) for four hours at 200°C under a pressure of 0.3 Pa. The gravimetric soil water content along the soil profiles was determined based on the soil weight loss following cryogenic water extraction. Soil water content determined this way deviates from the classical oven drying method and results in slightly lower values. In the case of oven drying, samples are taken via stainless steel cores. These soil cores still have intact pore systems that contain pore water. However, as soil samples taken for cryogenic extraction are disturbed soil samples, they do not include all the pore water. We use the gravimetric soil water content from cryogenic extraction not as an absolute value, but rather as a relative value to identify differences along the soil profile. Groundwater and ponded surface water of flooded rice were collected once a week from each plot at existing sampling stations (Heinz et al., 2013). Low-cost car wiper pumps (Art. Nr. 103 158, TOPRAN, Bremen, Germany) were used to pump water with a pumping rate of 40 L h⁻¹ to the sampling container which was installed in the centre of the nine fields (Fig. 1). The installation length below ground was 2.0 m. The groundwater table varied depending on the season. During the WS the groundwater table was at 0.5-0.6 m below ground and at 0.6-1.7 m during the DS. Rainwater and irrigation water were sampled event-based. For detailed information on the experimental design and sample collection as well as field preparation such as puddling and ploughing, see Mahindawansha et al. (2018b, 2018a).

2.3 Isotopic measurements

The oxygen and hydrogen isotopic compositions of the water samples (extracted soil water and liquid samples) were measured via off-axis integrated cavity output spectroscopy (OA–ICOS, DLT–100–Liquid Water Isotope Analyzer, Los Gatos Research Inc., Mountain View, CA, USA) and reported in per mil [‰]. The analytical precision for $\delta^{18}O$ and δ^2H was 0.2 ‰ and 0.6 ‰, respectively. All water sources (isotopic data) were checked for spectral interferences using the Spectral Contamination Identifier (LWIA-SCI) post-processing software (Los Gatos Research Inc.). According to this test, none of the soil water samples were contaminated. The global meteoric water line (GMWL) was defined following Rozanski et al. (1993) ($\delta^2H = 8.2\delta^{18}O + 11.3$). The Local Meteoric Water Line (LMWL) was calculated with $\delta^2H = 7.52\delta^{18}O + 5.86$, using stable isotope compositions of local precipitation collected from 2000 until 2015 (GNIP–IAEA, 2016). Line conditioned excess (lc-excess) was calculated for soil water samples as suggested by Landwehr and Coplen (2006), with lc-excess $= \delta^2H - a\delta^{18}O - b$, where a and b refer to the slope and intercept of the LMWL, respectively. The lc-excess is an indicator for evaporation, with lower values reflecting larger evaporative losses. We used the lc-excess to infer the seasonal dynamics of evaporation fractionation (Sprenger et al., 2017).

2.4 Calculation fraction of evaporation

The joint effect of equilibrium and kinetic isotopic fractionation during the phase transition from liquid water to vapour can be estimated using the Craig–Gordon model (Craig and Gordon, 1965). Sprenger et al. (2017) have recently used Equation 1 to estimate evaporation from the topsoil (0–0.1 m). Equation 1 is based on the Craig–Gordon model and formulations introduced by Gonfiantini (1986) to estimate the fraction of evaporation loss (F_E) from the soil water based on an isotope mass balance approach as follows:

$$F_E = 1 - \left[\frac{(\delta_S - \delta^*)}{(\delta_P - \delta^*)}\right]^m \tag{1}$$

where δ_S is defined as the isotopic signal of the soil [‰], δ_P is the original isotopic signal of soil water [‰], δ^* is the limiting isotopic enrichment factor [‰], and m is the temporal enrichment slope [–]. In our study, the original isotopic signal δ_P is the signal of the water input via precipitation or irrigation. During the WS, δ_P was estimated as the weighted average of the isotopic signals from the frequently occurring large precipitation events (larger than 10 mm). For the DS, we used the weighted mean of the irrigation water as the input signal. We assumed steady state conditions, as the samples were taken between 10–12 a.m. and thus at a time when steady-state conditions in rice fields can be assumed (Wei et al., 2015). Variables δ^* and m were calculated following Equations 2 and 3, respectively, as described in Benettin et al. (2018) and Gibson et al. (2016):

$$\delta^* = \frac{(RH\delta_A + \varepsilon_k + \varepsilon^+/\alpha^+)}{(RH - 10^{-3}(\varepsilon_k + \varepsilon^+/\alpha^+))} \tag{2}$$

$$m = \frac{(RH - 10^{-3}(\varepsilon_k + \varepsilon^+/\alpha^+))}{(1 - RH + 10^{-3}\varepsilon_k)} \tag{3}$$

where δ_A is the isotopic composition of atmospheric vapour [‰] (assuming that the isotopic composition of atmospheric vapour is in equilibrium with precipitation; see also Eq. 4 in the supplementary materials), RH is the relative humidity, ε_k is the kinetic fractionation factor [‰], and α^+ [–] and ε^+ [‰] are equilibrium fractionation factors. The temperature-dependent parameter α^+ was calculated for δ^2H and $\delta^{18}O$ separately (Benettin et al., 2018). Furthermore, ε_k was calculated according to Benettin et al. (2018), presuming diffusive transport in soil pore spaces (Barnes and Allison, 1983). The equilibrium isotopic separation between liquid and vapour was computed as $\varepsilon^+ = (\alpha^+ - 1)10^3$ [‰] (Benettin et al., 2018). As part of the calculation of ε_k, the aerodynamic diffusion parameter n [–] has to be set. It reaches 1 when the soil is dried to residual moisture levels (Mathieu and Bariac, 1996), presenting turbulent conditions. We anticipated $n=0.5$ for wet rice fields with saturated soils (Good et al., 2014), $n=0.7$ for dry rice, and $n=0.9$ for maize.

2.5 Statistical analysis

We tested for significant statistical differences ($p \leq 0.05$) of stable isotopes of water (δ^2H and $\delta^{18}O$) during seasons, growing stages, and treatments between all water sources. Normal distribution was tested by the Shapiro Wilk test and homogeneity of variances by the Fligner Killeen test (Python 2.7.10.0). Because of the non-normal distribution of data, we further carried out a non-parametric rank-based test considering no ties. The isotopic values of the two treatments with straw and without straw as a control plot were combined for each crop for further analysis, as there were no significant differences for stable isotopes of water between the two treatments ($p>0.05$).

3 Results

3.1 Soil water isotopic distribution

The original isotopic signal of the incoming water changed depending on the season, especially during the WS. As a result of frequent precipitation events with different amounts introducing strong variations in the isotopic composition (δ^2H from –55.20 to –10.89 ‰ and $\delta^{18}O$ from –7.91 to –2.54 ‰), the isotopic signal of the incoming water varied significantly (Fig. 2). Both δ^2H and $\delta^{18}O$ values of ponded surface water and groundwater were higher at the beginning of each season and decreased towards the end (Fig. 3). During both seasons, ponded surface water and groundwater under wet rice showed a relatively similar range of isotopic composition with no statistically significant differences (Table 2). A distinct difference in the composition of GW was only observed under maize in the DS. Stable isotope compositions of irrigation water were not significantly different between seasons. Rainwater was isotopically similar to groundwater and ponded surface water during the WS, unlike during the DS, where it was significantly different.

Figure 3 displays the δ^2H and $\delta^{18}O$ together with the water content and lc-excess values in soil water as a function of soil depth during GS1, GS2, and GS3 of wet rice during the WS, along with wet rice, dry rice, and maize during the DS with the standard deviation of the replicates. The range of isotopic composition of rainwater and irrigation water defines the water input to the system for each season (average values are presented in Table 2). The isotopic composition of soil water from fields with different crops during the DS were statistically different (more enriched) from the wet rice during the WS. Within the DS itself, there was a tendency for more depleted conditions in the upper soil horizons of wet rice compared to maize and dry rice. We did not find such a distinct difference for the soil layers below 0.2 m. Results for GS2 and GS3 of maize and wet rice were statistically different during the DS, and maize and dry rice were statistically different except for the GS3 of dry rice. The isotopic signals of the soil profiles down to a depth of ~0.2 m were highly variable, becoming more stable further below. Therefore, soil water isotopic values can be divided into two categories: shallow soil water from 0 to 0.2 m, and deep soil water from 0.2 to 0.6 m. In the wet rice soil, the isotopic values increased until the depth of 0.05 m and then decreased again until about 0.2 m (Fig. 3a, b, e, f). Interestingly, in wet rice soils, the depth of the highest isotope enrichment, which is just below the soil surface, decreased deeper in the soil during the growing period from GS1 to GS3 in both seasons. In contrast, the shape of the isotopic profiles of dry rice and maize follow a different pattern compared to wet rice, with higher δ^2H and $\delta^{18}O$ values at the soil surface and an exponential decrease down to around 0.2 m soil depth (Fig. 3i, j, m, n). The isotopic composition of shallow soil in dry rice fields decreased from GS1 towards GS3, where the values were stable in maize fields during all phases of plant growth. The isotopic values in deep soil were nearly stable in all the profiles regardless of the crop during both seasons.

Maize was characterized by dry soil conditions at the surface and shallow depths compared to both rice varieties (Fig. 3c, g, k, o). The highest water content was found for wet rice at the surface soil (17.7±1.2 %), and it was nearly constant below a depth of 0.2 m (12.0±1.3 %) during both seasons. The water content in dry rice soils was rather evenly distributed along the soil profile except for the soil surface. Soils under maize were getting drier as plants were growing, while such clear patterns were not observed for the rice crops.

We found an exponential increase in the lc-excess along the soil profile, particularly for maize, but also, though less apparent, for dry rice soils (Fig. 3l, p). For wet rice during the DS, the exponential pattern was even less obvious, but shallow soil layers still depicted lower lc-excess between -10 to -5 ‰ than deeper soil layers with values of -5 to 0 ‰. In contrast, lc-excess values of shallow soils in wet rice fields of the WS (Fig. 3d) generally decreased with depths of up to 20 cm and then levelled out at around -7 to -9 ‰. These patterns indicate a higher evaporation signal in shallow soils for the DS crops compared to the WS crop. The highest evaporative fractionation was found near the surface in maize fields with significantly lower lc-excess values during the last growing stage GS3. For maize, the lc-excess values decreased in most soil layers from GS1 to GS3, which was the opposite for dry and wet rice during the DS. No distinct, clear pattern could be found along the growing stages for wet rice during the WS. The lc-excess for rainwater for both seasons were positive and ranged from 0 to 9 ‰ during the DS and from -2 ‰ to 4 ‰ during the WS. The positive lc-excess for rainfall values during the DS indicate the influence of changes in moisture sources of the rainfall. Irrigation water has a nearly similar mean of -5 ‰ for the WS and -4 ‰ for the DS.

The δ^2H and δ^{18}O values of soil water plot on a line below the LMWL due to the evaporation effect (Fig. 4). The slope of the regression line and coefficient of determination (R^2) were higher in the DS (avg. slope=5.1, R^2=0.92) than during the WS (avg. slope=3.5, R^2=0.54). Soil water δ^2H and δ^{18}O compositions were higher (enriched) in shallow soils and deviated more strongly from the LMWL than soil water from deep soils. We observed lower slopes and more clustered data points in wet rice soil during the WS, indicating lower soil evaporation compared to the DS. During the WS, several shallow soil isotopic values plotted close to the LMWL, and some deep soil values deviated more from the LMWL (Fig. 4a–c). During the DS, slopes of the regression lines were lower for wet rice (slope=5.2, R^2=0.88) than for dry rice (slope=6.0, R^2=0.94) and maize (slope=5.5, R^2=0.91) (Fig. 4d–l). Due to less frequent and shorter precipitation events during the DS, the isotopic signal of the incoming water was dominated by irrigation water, with nearly constant isotopic composition during the growing period. Small precipitation events were subjected to higher evaporative loss and resulted in enriched isotopic composition during this time (Table 2).

3.2 Fraction of evaporation loss from soil water

The estimated fraction of evaporation F_E at each soil depth was derived by means of an evaporative enrichment of heavier isotopes in the soil water. Fig. 5 shows F_E estimated based on both isotopes for the growing seasons WS and DS, the growing stages GS1-GS3, the different crops wet rice, dry rice, and maize as well as for different soil depths.

A clear trend of F_E with soil depth can be depicted at all growing stages during the DS for both crops, maize, and dry rice, reaching below 0.2 for dry rice and even below 0.1 for maize in deep soils (Fig. 5g-l). During the DS, soils in dry rice fields showed high soil F_E at shallow depths at the beginning of the first two growing stages GS1 (0.54±0.1) and GS2 (0.50±0.1) which decreased to F_E of around 0.27±0.1 at GS3. Further, we observed lower F_E average values (0.2±0.1) in deep soils between 0.25 and 0.6 m in these fields. For maize, F_E remained stable at 0.3±0.1 in shallow soils throughout the season and decreased with depth for both isotopes (to around 0.07±0.05) (Fig. 5j, k, l). The F_E in shallow soils of wet rice in the DS ranged from 0.42±0.08 to 0.20±0.08 (similar for both isotopes) and remained nearly stable in deep soils at 0.13±0.1 (Fig. 5d, e, f). Overall, we did not find a similar decreasing trend with depth, as reported for dry rice and maize. Instead, particularly during GS2 and GS3, the highest fractions of F_E were found at moderate soil depths of 0.1 to 0.2 m. Results regarding the estimation of F_E based on δ^2H and δ^{18}O are fairly similar for all dry season crops. For wet rice in the WS, F_E of both isotopes differed significantly by 0.1 to 0.2 for the topsoil and even 0.5 for deep soils (Fig. 5a, b, c). During the WS, F_E in shallow soil decreased from around 0.72±0.1 (GS1) to 0.47±0.06 (GS3) for δ^2H and from 0.87±0.07 (GS1) to 0.76±0.07 (GS3) for δ^{18}O. The general trend with slightly higher F_E at moderate soil depths and again decreasing F_E further down, which we observed for wet rice during the dry season, was also confirmed for wet rice during the wet season. The soil water in wet rice fields during the WS carried a larger signal of high evaporation losses down along the soil profile. The estimated F_E from ponding surface water (data not shown in Fig. 5) was found to be larger during the WS than during the DS with no significant difference between δ^2H and δ^{18}O. The F_E of ponded water during the WS did not fluctuate with time, and remained close to 0.92±0.07, while

during the DS values decreased from GS1 (0.67±0.03) to GS3 (0.24±0.01). Here, F_E of ponded surface water indicates a high evaporation loss during the WS. The evaporation signal is carried to deeper layers by subsequent infiltration and percolation.

4 Discussion

General mechanisms in soil water movement

5   In the absence of lateral water transfer and assuming negligible fractionation from root water uptake, the isotopic profiles in soil water reflect a balance between mixing from infiltration, percolation, and fractionation from soil evaporation (Hsieh et al., 1998, Barnes and Allison, 1984). Depending on the evaporation effect on soil water isotopic composition and water transport processes, we found a change of the isotopic composition at around 0.2 m below the surface at our study site. This phenomenon has developed predominantly due to the existence of the dense, less permeable plough pan, which separates the puddled

10   shallow soil and non-puddled subsoil in paddy fields. The plough pan is a result of repeated ploughing and puddling over many years due to cultivation (Chen and Liu, 2002). The isotopic profiles we observed are a response to three major mechanisms that drive soil water movement at our sites, i.e., 1) matrix flow, 2) preferential flow, and 3) evaporation. These three mechanisms will be discussed in the following sections.

4.1 Matrix flow

15   In the unsaturated zone in dry rice and maize fields, the diffusive vapour transport process is dominant (Bittelli et al., 2008). Kinetic fractionation leads to the accumulation of heavy water molecules (formed by ^2H and ^{18}O) at the water-air interface, which are subsequently transported downwards and then mix with the soil matrix (Horita et al., 2008). Downward water movement at steady-state or slowly changing conditions results in an exponential evaporation profile with depth during the drying stage that is comparable to those found in soils beneath dry rice and maize (Fig. 3i, j, m, n) (Zimmermann et al., 1966;

20   Barnes and Allison, 1988; Rothfuss et al., 2015).

Under flooded conditions of wet rice, water slowly percolates from the ponding, open water body. The upper soil layer is affected by isotopically enriched water via a gravity-driven, piston-like matrix flow. The enrichment of heavier isotopes of soil water increases with depth (until it reaches the most enriched culmination point) (Fig. 3a, b, e, f). We assume that this is a result of the successive displacement of pre-existing mobile soil water by infiltrating water. During ponding, infiltration

25   modifies the soil water isotopic composition in the uppermost part of the profile, and re-evaporation of infiltrated water occurs. Still, soil water in fine pores represents a quasi-stationary storage, exchanging water and isotopes with the mobile phase (Gazis and Feng, 2004). As a result, the isotopic composition indicates a depletion until reaching a stable value below approximately 0.2 m. Baram et al. (2013) found a similar isotopic pattern in clay soil in Israel, which they explained by gravity-driven, piston-like matrix flow under continuous ponded infiltration.

30   In soils with fine pores, capillary rise could have further affected the observed isotopic patterns, depending also on depth to groundwater. It has been shown that capillary rise of shallow groundwater can influence soil moisture and its isotopic

composition in the upper meter of clayey soil (Baram et al., 2013; Clark and Fritz, 1997). An upward matrix flow through capillary rise has probably occurred in our system as well, given the fine texture of the soils (Table 1). However, the effect seems to be negligible, as the GW signatures we measured were more enriched than the soil water found at greater soil depths (Fig. 3). This is probably due to the fact that GW head levels were often substantially below the deepest soil layer we sample (Mahindawansha et al., 2018a).

The observed isotopic signals in the shallow soils could also indirectly be explained by a transpiration bias. Transpiration decreases the soil moisture, but preserves the isotopic composition (Baram et al., 2013). With decreasing soil moisture, incoming water has a relatively stronger imprint on the soil's isotopic composition. Further, hydraulic redistribution of water in the vadose zone is an important process of passive transport of soil water along a hydraulic gradient through the root system (Richards and Caldwell, 1987). Therefore, hydraulic redistribution can influence the pore water stable isotopic composition and reshape the soil water isotopic profile. Sprenger et al. (2016) discussed the significance of hydraulic redistribution in the soil hydrological cycle. While the review of (Walter, 2010) only indicated a limited impact of hydraulic redistribution on the isotopic composition of soil water, the selective removal of water combined with redistribution can be relevant. Still, isotopic measurements alone are not sufficient to estimate redistribution volumes (Emerman and Dawson, 1996) and therefore the potential impact of hydraulic redistribution requires a combination of physical transport modelling and isotopic composition and should be the focus of further studies.

4.2 Preferential flow through desiccation cracks

Desiccation cracks in maize fields (below 0.2 m) reached deeper (∼0.2 m) and were narrower (∼0.02 m) than those developed in dry rice fields (own observation). We assume that this feature controls the development of soil water isotopic composition in the uppermost soil compartment. This isotopic signal is then carried to greater soil depth via leaching. At greater depth, mixing of soil water with water transported through cracks may occur. The concept of preferential transport and subsequent mixing transfers the isotopic signal of evaporation to a greater depth. Open desiccation cracks can also allow for diffusive water transport to crack surfaces at greater depth. Baram et al. (2012) observed that naturally formed desiccation crack systems can create preferential flow paths that reach more than a meter deep. In our maize fields, we observed that the isotopic composition of groundwater is strongly influenced by irrigation water, suggesting the existence of fast flow conduits (Mahindawansha et al., 2018a). He et al. (2017) have also observed leaching losses of water and nutrients in a lysimeter experiment, which they attributed to crack flow mechanisms at the same study site. Preferential flow through desiccation cracks is therefore likely to be a dominant flow pathway in rice-based cropping systems, especially for crops grown in the dry season that are planted to replace water-demanding wet rice. In line with this concept, we have recorded a gradual isotopic depletion towards deep soils of dry rice and maize fields. We attribute this to subsurface mixing of isotopically enriched soil water with the depleted irrigation water from the nearby reservoir that percolated into the deep vadose zone via preferential flow paths (Baram et al., 2012; Nativ et al., 1995).

4.3 Evaporation effect

Evaporation and the lc-excess

Systematic isotopic depletion and increasingly negative values of lc-excess profiles indicated declining evaporation from GS1 to GS3 in rice (Fig. 3), particularly in shallow soils. In both dry and wet rice, the isotopic profiles showed a clear shift from more to less enriched values across the growth stages, especially in shallow soils and regardless of the season. However, the most isotopically enriched water in wet rice was transferred down to even greater depths in conjunction with plant growth (Fig. 3a, b, e, f). In maize fields, the evaporation fraction gradually increased towards the end of the season when irrigation ceased (Fig. 2 and 3), resulting in dry soil conditions and a soil water deficit. Therefore, we conclude that there is an influence of the crop type and growth stage on evaporation fractionation in the soil water, matching previous reports that the plant cover reduces kinetic fractionation processes in the soils (Burger and Seiler, 1992; Dubbert et al., 2013).

Evaporation and the LMWL

Comparisons of regression lines of soil water samples to the GMWL in the dual isotope space ($\delta^{18}O$, δ^2H) helped to identify the environmental conditions during soil evaporation with regard to season and crop (Fig. 4). The slope of the $\delta^{18}O$–δ^2H relationship decreases because of kinetic fractionation (Dansgaard, 1964). This deviation can be used to estimate evaporation losses (Clark and Fritz, 1997). The steeper slopes of the dry soils (maize, dry rice) can be explained by an increase in the effective thickness of the vapour transport layer (Barnes and Allison, 1988) compared to the soils of wet rice. For soils under wet rice, a steeper gradient near the surface was found, similar to observations regarding saturated soils made by Allison (1982). During the WS, the $\delta^{18}O$–δ^2H relation of deep soil water under wet rice fell even further below the LMWL from GS2 to GS3 (Fig. 4b, c). In contrast, shallow soils plotted closer to the LMWL indicated lower evaporation rates. Furthermore, deep soil water showed isotopic similarity to the irrigation water. Following these observations, we assume that the deep soil isotopic profiles result from a mixing of soil water with irrigation water, the latter likely stemming from the previous DS (memory of the old isotopic signal) that moved downward via matrix flow. Due to the low rates of percolation of 1 to 5 mm d^{-1} in clay paddy soils (Bouman and Tuong, 2001), deep soil profiles with multiple soil layers may reveal a record of antecedent evaporation conditions or preferential flow shortcuts between the layers.

Apart from this, all soil profiles presented enriched values and distinct evaporation processes during the WS (Fig. 4a–c). Lower slopes of evaporation lines in wet soil compared to dry soil point to greater kinetic effects (Cooper et al., 1991). Slopes of evaporation lines <3.5 were reported to indicate diffusion processes (Allison et al., 1983). We, therefore, assume that diffusion processes in the subsurface were relevant for shaping soil isotopic profiles in the WS, especially at GS1 and GS3 (Fig. 4a, c). During GS2, mixing processes between infiltrating water dominated and limited diffusion processes due to continuous intense precipitation events during that time. In line with this, a higher correlation between plant water (xylem) and rainwater during this time compared to the other growing stages was reported by Mahindawansha et al. (2018b). Overall, an enrichment of soil water isotopic composition during the WS and a depletion during the DS is comparable to observations made by Hsieh et al.

(1998) in an arid to humid transect in Hawaii. Similar differences between depleted winter and enriched summer isotopic profiles in combination with mixing processes were also reported by Baram et al. (2013) and DePaolo et al. (2004).

Unproductive water losses via evaporation

Kinetic fractionation and its imprint on soil water isotopic profiles in the shallow soil is relatively small in tropical climates given generally high relative humidity (Gonfiantini, 1986). Nevertheless, our observations point to kinetic fractionation down to a depth of ~0.2 m, shallower than the average depth in temperate regions (~0.3 m) (Gazis and Feng, 2004; Sutanto et al., 2012), the Mediterranean (~0.5 m) (Oshun et al., 2016; Simonin et al., 2014), or in arid climates (~3 m) (Allison and Hughes, 1983; Singleton et al., 2004). Shallow soils exhibit a decreasing trend of F_E during both WS and DS from the beginning of the growing season towards its end (Fig. 5), most likely driven by an increase in the leaf area of the aboveground vegetation. Rothfuss et al. (2010) made comparable observations in a lab-based experiment on soil columns and reported changes in F_E over time. They found values starting with 100 % at bare soil conditions, and dropping to 5 % at full development of the deep-rooting perennial grass grown in the columns. In our study, the fraction of soil evaporation was estimated as 40 % at the beginning of the DS and decreased to 25 % towards the end, while it dropped from 80 to 60 % during the WS. Values of about 30 % (maize) and 50 % (dry rice) evaporation losses were reported for Asia at the same sites in the Philippines based on eddy covariance measurements by Alberto et al. (2014). Similar values were confirmed by Bouman et al. (2005) based on a review of tropical upland and lowland rice varieties under irrigated aerobic conditions. Simpson et al. (1992) report 40 % for flooded rice fields in a semiarid region of south eastern Australia and Maruyama and Kuwagata (2010) about 60% for paddy fields in southwestern Japan. During the WS however, F_E was higher in the shallow soil compared to the DS. This finding might be related to the high temperatures leading to higher water pressure deficits between soil water and the atmosphere. The substantially large difference of F_E during the WS between δ^2H- and δ^{18}O-based assessments, can be related to the different hydrogen compounds. The values we obtained refer to the fraction of water loss from the soil matrix and small/intermediate pores. With isotope methods, we only estimated unproductive evaporation losses from the soil. However, Wei et al. (2018) showed that an isotopic approach can also lead to higher estimates of the fractions compared to model results for rice and maize in Tsukuba, Japan. Overall, we conclude that the isotope method provides comparable results for previous studies.

4.4 Differences in fractionation of δ²H and δ¹⁸O

Apart from the highly depleted isotopic signal for δ^2H observed in deep soil under wet rice fields during the WS (Fig. 3b), there was a systematic deviation of about 20 % between δ^2H and δ^{18}O fractionation in shallow soil and 40 % in deep soil. (Fig. 5a–c). This difference may have resulted from the formation of specific hydrogen compounds under continuous inundation conditions. Flooding affects soils chemically, physically, and biologically, resulting in a reduction of redox potential (Fageria et al., 2011; Zhang et al., 2015). Due to the anaerobic conditions that develop in submerged soil, hydrogen compounds such as CH_4, H_2S, H_2, and NH_4^+ can be produced via microbial anaerobic respiration (Fageria et al., 2011; Gerardi, 2003). The formation of these hydrogen compounds leads to isotopic exchange and bias in δ^2H, as observed by Baram et al. (2013) in clay

soils below ponded wastewater conditions. CH_4 emissions in wet rice fields on our study site were higher during the WS compared to the DS (Weller et al., 2016), and this may have caused lower slopes in the dual isotope plots as observed (Fig. 4a–c).

Furthermore, the equilibrium constant for isotopic partitioning of liquid water with vapour (1,000 $\ln\alpha$) is a function of the temperature (here we present the values at 27°C) and the sign of the value (positive), e.g., $H_2O_{(l)} \leftrightarrow H_2O_{(g)}$ for $\delta^{18}O$ +9.2 (Freidman and O'Neil, 1977; Majoube, 1971) and +74.3 for δ^2H (Majoube, 1971). Water vapour δ^2H further isotopically fractionates with $CH_{4(g)}$ (1,000 $\ln\alpha$=+23.4, see Bottinga, (1969)), $H_2S_{(g)}$ (1,000 $\ln\alpha$=+851.0 as in Galley et al., ((1972); Clark and Fritz, (1997)), as well as liquid water with $CH_{4(g)}$ with 1,000 $\ln\alpha$=+242.1 (Horibe and Craig, 1995), leading to higher δ^2H (enriched) in both phases. Moreover, liquid water and water vapour further manifest an equilibrium with $H_{2(g)}$ with higher equilibrium fractionation (Bottinga, 1969; Rolston et al., 1976). As a result, the assumption of δ^2H enrichment is further reinforced. The difference between δ^2H and $\delta^{18}O$ has been found to be more pronounced at a greater depth, suggesting the formation of hydrogen compounds in deeper soil (Fig. 5 a–c). Besides, exchange rates and fractionation with kaolinite and smectite (Gilg and Sheppard, 1996) are faster and more pronounced for δ^2H. The assumption for this dissimilarity between δ^2H and $\delta^{18}O$ can be quantified by a sensitivity analysis, giving a relative depletion by 5±2 ‰ of δ^2H. Because of the above processes, bias can result in the calculation of F_E during the WS. Due to the high standard deviation of the isotopic composition in extreme precipitation events during the WS, prediction of the original water source at a time was also more uncertain. The F_E values are sensitive to the isotopic composition of atmospheric vapour and original water input. Nevertheless, only seasonal averages were assigned in the calculation. This difference was not prominent in wet rice fields under anaerobic conditions during the DS, where oxidizing conditions occurred in time gaps between irrigation events; it was also not observed in dry rice and maize fields.

In addition, vacuum-extracted soil water also contains bound water plus adsorbed water, making isotopic composition lower (Gaj et al., 2017; Velde, 1992), separate from additional systematic errors resulting from the extraction method (Orlowski et al., 2016). High water-holding capacity (Brouwer et al., 2001; Hazelton and Murphy, 2016) and the shrinking and swelling behavior (Baram et al., 2013; Dasog et al., 1988) of clayey soil add complexity to the analysis. Determination of α_k can also result in estimation errors (i.e., a maximum uncertainty of the partitioning) of 1 to 29 %, depending on the value of α_k and the day of the partitioning (Rothfuss et al., 2010).

5 Conclusions

We identified three main processes, which are responsible for variations in the soil water isotopic profile: soil evaporation, slow soil water movement via matrix flow, and the refilling of deep soil water through preferential flows via desiccation cracks. Apart from this, we also quantified the relative fraction of soil water returning to the atmosphere as direct evaporation as unproductive soil water losses and related its pattern to crop types and seasons. However, independent tools to confirm the findings of complex soil water isotope studies on evaporation would be needed. There was a clear isotopic separation between

shallow and deep soil, with higher enrichment in shallow soil at around 0.2 m below the surface. Deep soil in wet rice fields often presented inverted evaporated profiles because deep soil layers carried over the history of the transported evaporation signal from the previous season. Shallow soils in maize fields showed a stronger soil evaporation effect than rice fields. Compared to the original water input, greater water loss was estimated during the WS compared to the DS when referring to evaporation from the soil matrix. The observation of difference in the fractionation of δ^2H and δ^{18}O deserves further research. Even though we provided a theoretical background of how this fractionation might occur, we were not able to measure all the different components which affect the fractionation. Further research into these processes would help to better understand the evaporation process.

To conclude, water losses via soil evaporation is a major unproductive loss next to leaching losses, especially during the early growing stage. Therefore, our study helps to increase understanding of soil water transport processes and evaporation losses from soil in response to crop rotation systems. Our hypothesis of reducing the unproductive water losses by introducing dry seasonal crops is supported by isotope data. Farmers should apply mitigation methods to reduce soil water evaporation, e.g. by mulching, or growing cover crops in the fallow period and by protecting the plough pan.

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Tables and Figures

Table 1. Soil texture and average bulk densities of different depths along the soil profile

Soil depth (m)	Texture			Bulk density (g cm^{-3})	
	Clay (%)	Silt (%)	Sand (%)	Rice fields	Maize fields
0.0–0.1	58.3	33.4	8.4	0.92±0.03	1.17±0.02
0.1–0.2	59.5	30.9	9.7	1.02±0.03	1.13±0.04
0.2–0.4	58.9	29.6	11.5	n.a	n.a
0.4–0.6	50.0	26.7	23.4	n.a	n.a

Table 2. Mean±standard deviation (SD) of all water samples (rainwater weighted mean (RW), irrigation water (IW), groundwater (GW), and ponded surface water (SW)) from different crops (wet rice, dry rice, and maize) during the wet season (WS) and dry season (DS).

Season	Crop	Water type	δ^2H±SD ‰	δ^{18}O±SD ‰
WS		RW	-26.82±2.30	-4.42±0.34
		IW	-32.00±3.25	-4.34±0.65
	Wet rice	GW	-23.76±5.24	-3.03±1.21
	Wet rice	SW	-24.06±7.36	-3.22±1.69
DS		RW	8.73±0.62	0.05±0.08
		IW	-34.60±3.56	-4.89±0.56
	Wet rice	GW	-14.66±7.46	-1.75±1.27
	Wet rice	SW	-14.15±9.41	-1.80±1.41
	Dry rice	GW	-12.56±8.75	-1.37±1.52
	Maize	GW	-22.57±7.60	-3.10±1.19

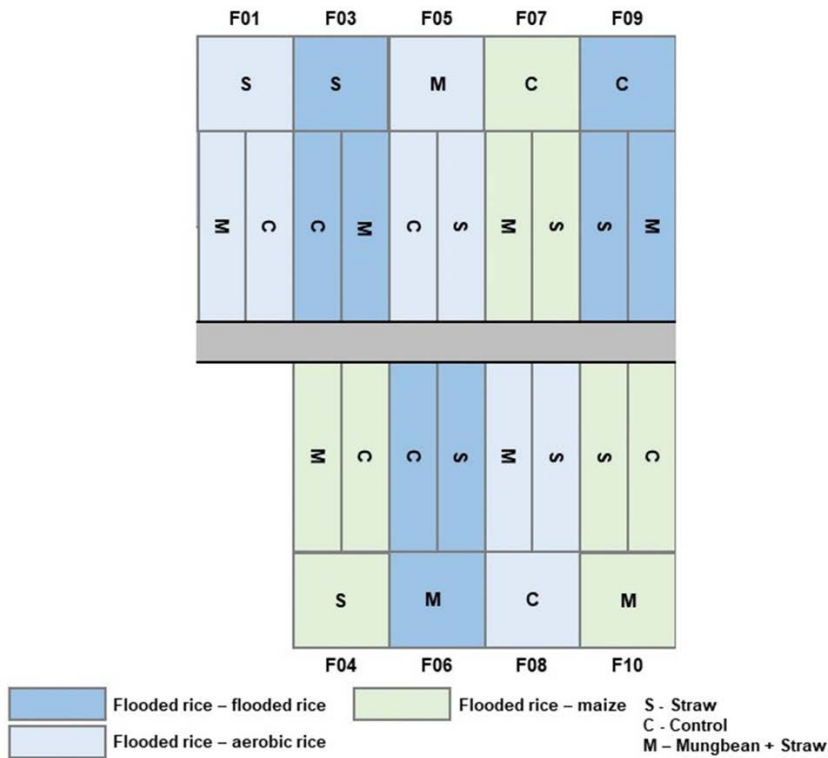



Figure 1. Experimental field design. The experiment consisted of nine fields (F) with three different crop rotations and water management practices. During the wet season, all fields were cultivated with wet rice, while during the dry season, three fields each were cultivated with wet rice, dry rice, and maize. Each field is divided into three different treatments (S=straw incorporated in the soil, C=control, M=straw plus mung bean as an inter-crop in the dry to wet transition period). Note that the mung bean plots are not part of this study but are depicted for completeness of the field trial.

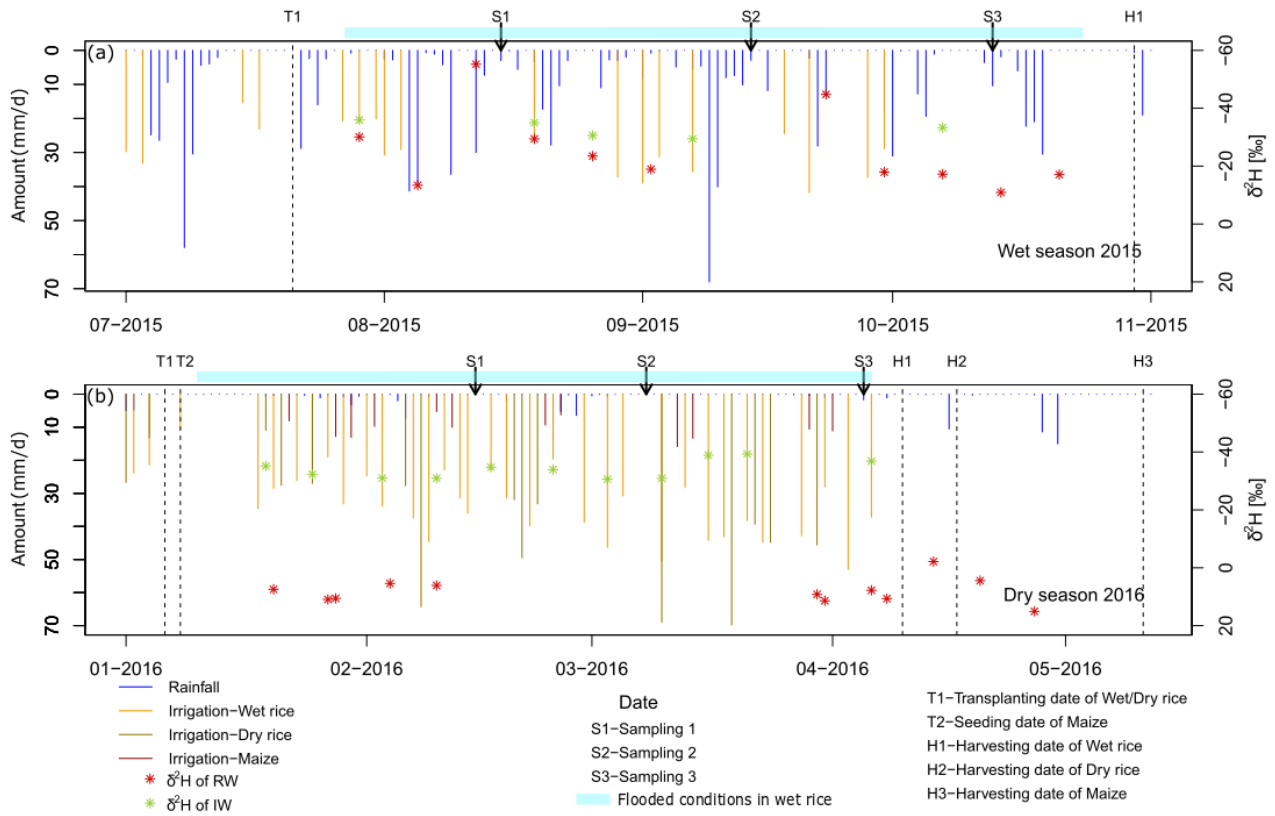



Figure 2. Temporal variation of water inputs (rainfall and irrigation water) of wet rice, dry rice and maize fields for the wet season 2015 (top) and dry season 2016 (bottom). Three main sampling dates during each season together with transplanting and harvesting dates are marked. Values of δ^2H are presented for rainwater (RW) and irrigation water (IW) during both seasons.

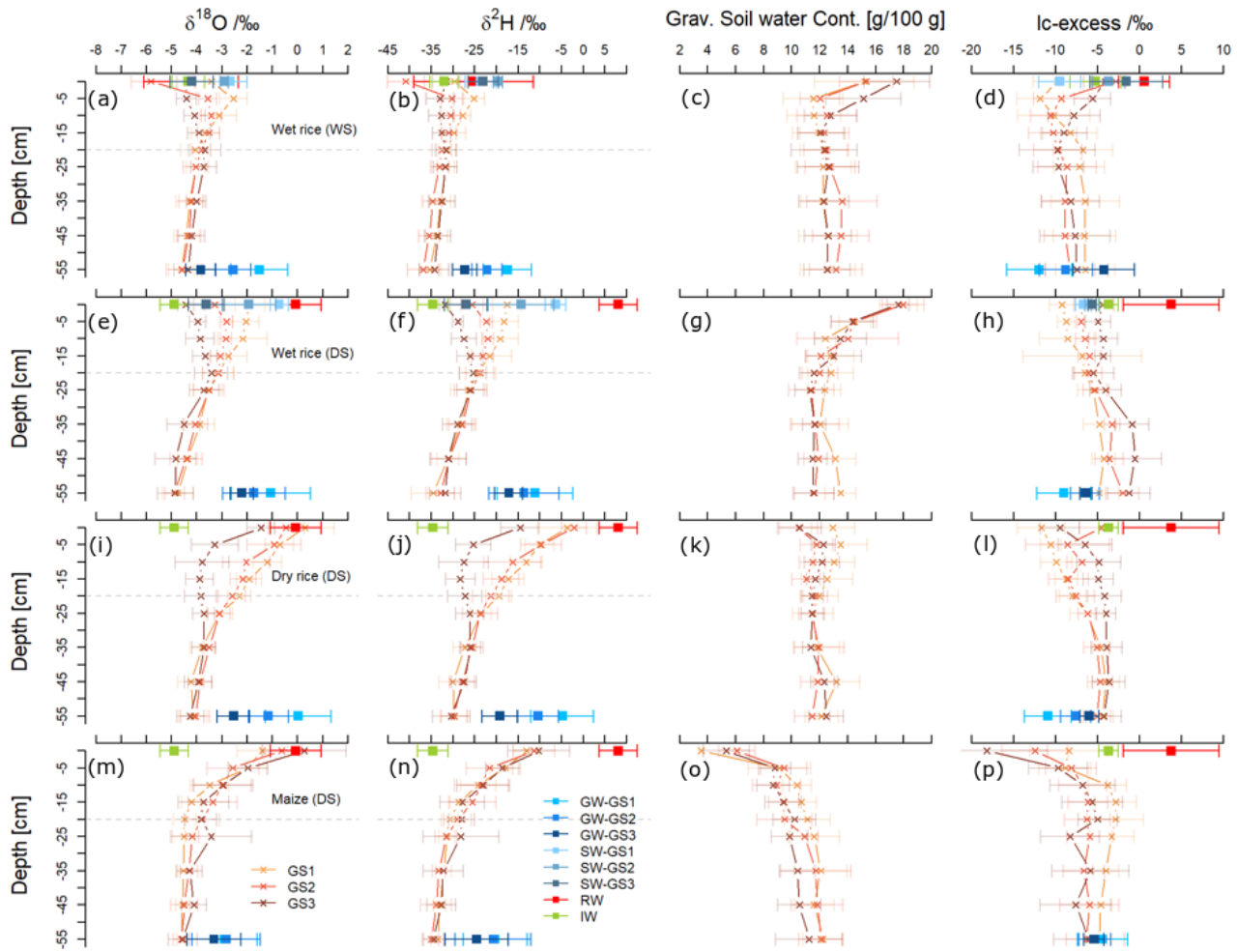



Figure 3. Depth profiles of means +/- standard deviation for $\delta^{18}O$ /‰, δ^2H /‰, gravimetric soil water content [g/100 g], and lc-excess /‰ from three main growing stages (GS1 to GS3) of wet rice (a–d) during the wet season (WS), and wet rice (e–h), dry rice (i–l), maize (m–p) during the dry season (DS). Seasonal averages +/- standard deviation of all water sources: rainwater (RW), irrigation water (IW), groundwater (GW) and ponded surface water (SW). Isotopic values are displayed at the top and bottom of the soil profiles.

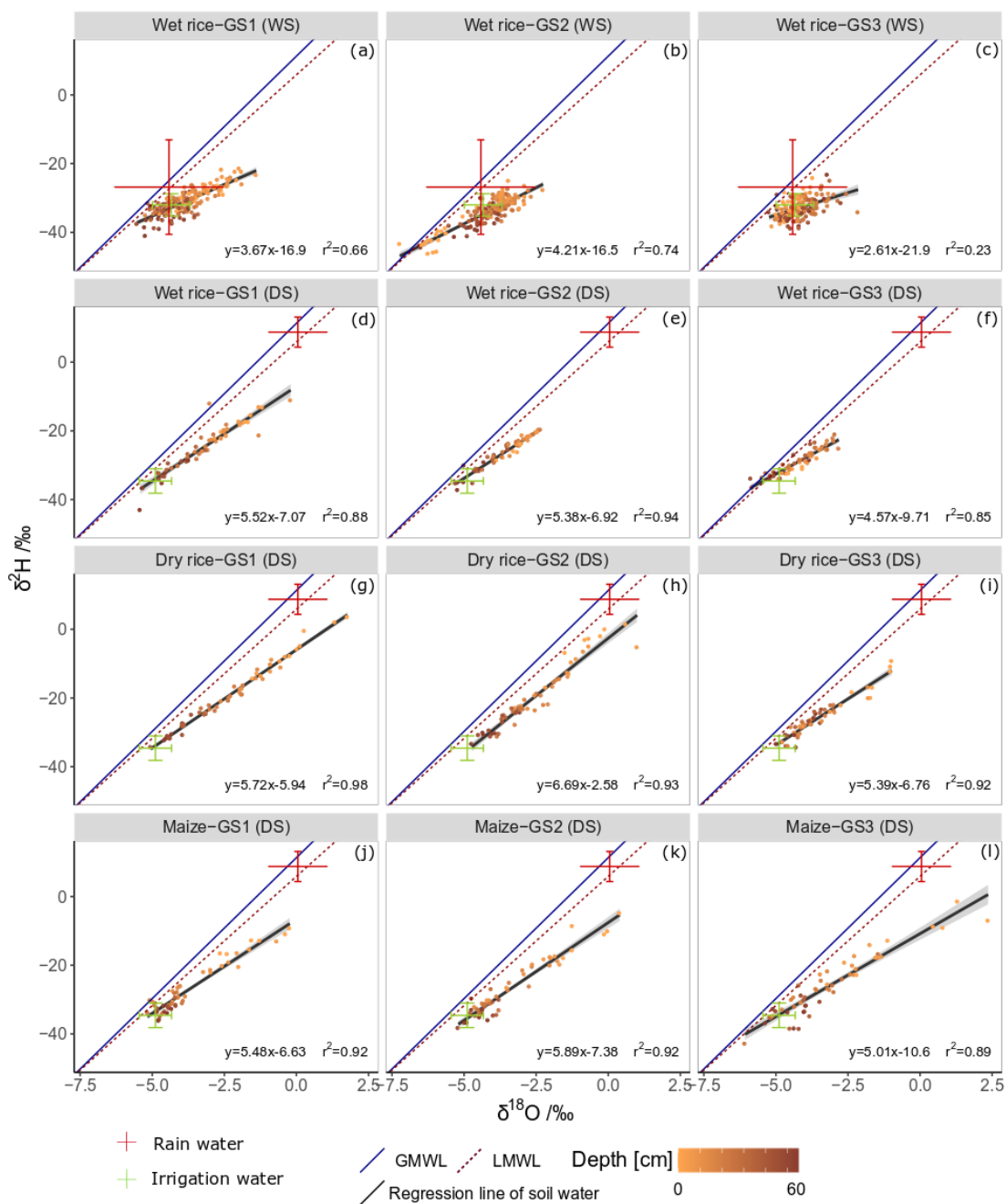
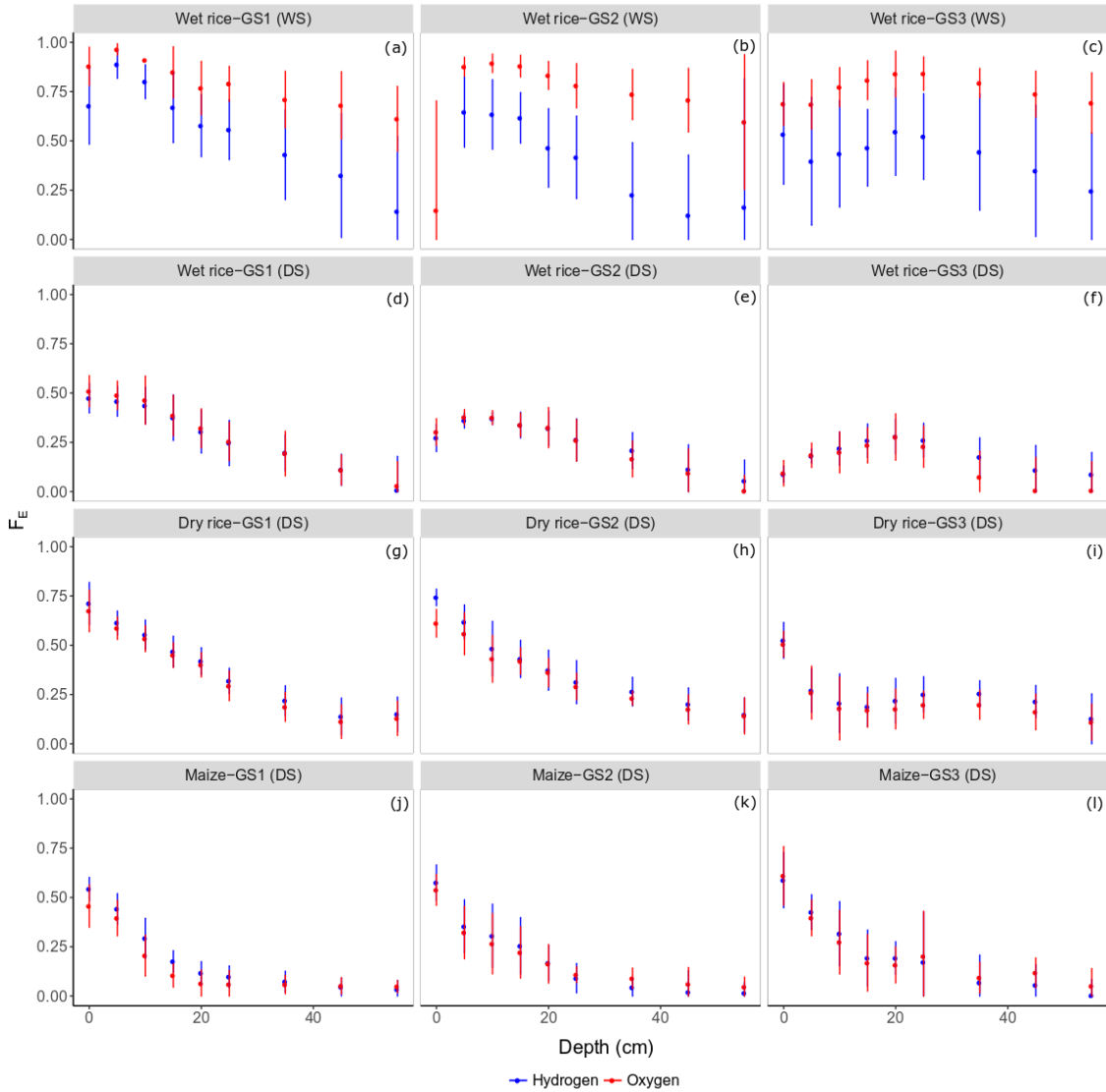



Figure 4. Dual ($\delta^{18}O$, δ^2H) isotope plots of soil water at 0–0.6 m depth, and ranges of other water sources (rainwater, irrigation water) from growing stage GS1 (a, d, g, j), GS2 (b, e, h, k), and GS3 (c, f, i, l), of wet rice (a–c) during the wet season (WS), and wet rice (d–f), dry rice (g–i) as well as maize (j–l) during the dry season (DS) in comparison to the local meteoric water line (LMWL) and the global meteoric water line (GMWL). The gray shaded areas represent the 95 % confidence interval of the black linear regression lines.

5   Figure 5. The fraction of evaporation loss (F_E) (Eq.1) estimated from $\delta^{18}O$ and δ^2H for the three main growing stages GS1 (a, d, g, j), GS2 (b, e, h, k) and GS3 (c, f, i, l) of wet rice (a–c) during the wet season (WS), and wet rice (d–f), dry rice (g–i), and maize (j–l) during the dry season (DS). Mean values at each depth (0–0.6 m) are displayed with +/- standard deviations.