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Landform transformation and long-term sediment budget for a Chernozem-dominated lowland agricultural catchment

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Abstract

The loess hilly lands of the South-West Foreland of the West Carpathians are characterized by favorable climatic conditions, smooth topography, fertile soils and a long settlement history. For the first time, we document changes to human-induced landforms in a small zero-order dry valley (0.28 km^2) with Calcic Chernozem soils and develop a long-term sediment budget. The original surface was reconstructed based on interpreting records of erosion and deposition from 185 soil cores and six valley cross-sections. The topography inherited from the Pleistocene was transformed through lowering of hillslope ridges and convexities, along with infilling of shallow depressions within the original valley bottom. On convex-convex landforms had an average net erosion of -0.32 ± 0.01 m, while average net accumulation in the valley bottom was estimated at 0.20 ± 0.01 m. The correlation between change in soil profile depth with profile curvature (-0.51, $p < 0.01$) was stronger than with slope (-0.42, $p < 0.01$). The sediment budget was calculated using the Average Per Unit (APU) approach. When slope classes were used in calculating net soil loss from catchments was $143 \cdot 10^3$ Mg·km^{-2}, whereas $154 \cdot 10^3$ Mg·km^{-2} was estimated when morphometric forms of hillslopes were used instead. Both values are < 60% of estimates for similar loess catchments in Central and Western Europe. Prior to 1949, the landscape structure was characterized by small and narrow fields and vegetative field barriers that limited water and aeolian sediment fluxes in the direction of the thalweg and main wind directions. The landscape structure favoured the low-intensity tillage, erosion and accumulation within fields, and short-distance sheet and rill erosion on hillslopes. The land-use changes in the 1950′s caused removal of barriers for water and aeolian sediment transport, and changed the direction of sediment fluxes of tillage erosion. Past landscape structures should thus be considered when establishing and interpreting sediment budgets for lowland agricultural catchments.

1. Introduction

Anthropogenic influences on soil erosion, landforms evolution, the hydrological cycle, and sediment and nutrient fluxes have been documented in agricultural areas worldwide (e.g., Bork, 2006; Quinton et al., 2010; Syvitski and Kettner, 2011), especially loess areas of temperate Europe with long histories of human occupation (Dotterweich, 2008; Notebaert and Verstraeten, 2010; Rommens et al., 2005). Reconstructing the original topography (Rejman et al., 2014; Rommens et al., 2007; Tinapp et al., 2008) allows identification of both natural and human-induced landforms changes, and calculation of spatially and temporally differentiated sediment budgets (Houben, 2008, 2012; Notebaert et al., 2011; Verstraeten et al., 2009). Such reconstructions enable descriptions of sediment production, transfer and storage zones. In addition, it can help identify the role of erosional processes and landscape structure in the evolution of landforms and historical sediment fluxes. Water erosion is considered as the dominant process during the majority of historical times (Dreibrodt et al., 2010; Van Oost et al., 2005) due to landscape structures providing pathways for concentration of overland flow, contributing to gully erosion (e.g., Stankoviansky et al., 2012; Superson et al., 2014). Tillage erosion occurs over significantly different spatial scales because it is limited to

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field boundaries (Govers et al., 1994) and affects topography and sediment fluxes in sub-mountainous agricultural landscapes (Stankoviansky et al., 2012). Houben (2008) proved landscape structures significantly influence soil erosion patterns and sediment storage in lynchets for a loess, hill-dominated landscape, while Rodzik et al. (2014) documented lowering of loess ridges in eastern Poland due to historical agriculture. In order to distinguish phases of soil erosion and soil profile truncation, different dating techniques have been applied to colluvium (Zádorová et al., 2013), alluvial sediments (Tinapp et al., 2008; Zygmunt, 2009), and catchment transects (Belyaev et al., 2013; Golosov et al., 2008, 2011, 2012; Lang and Hoenscheidt, 1999; Poręba et al., 2011; Sidorchuk and Golosov, 2003) in agricultural areas with Chernozem soils. However, the majority of past soil erosion and sediment budget studies (Delmas et al., 2012; Dotterweich et al., 2013; Janicki et al., 2002; Lang, 2003; Notebaert et al., 2009; Rommens et al., 2005; Verstraeten et al., 2009; Wolf and Faust, 2013, etc.) were conducted in Luvisol-dominated landscapes, outside of the loess areas in southwest Foreland of the Western Carpathians. Chernozem regions in Central, Southern and Eastern Europe provide favorable conditions for agriculture, with evidence dating back to the Neolithic age (Davison et al., 2006; Dolukhanov et al., 2005). Previous studies only focused on human impacts on Chernozem pedogenesis in the southwest Foreland of the Western Carpathians (e.g., Hraško, 1966; Fulajtár, 2009; Vysloužilová et al., 2015). This study is the first to research long-term catchment-scale human impacts on landform evolution for this region. The objective is to examine landforms changes and sediment fluxes subsequent to the onset of agriculture in a small catchment dominated by Chernozem soils.

2. Regional settings

The Voderady study area is situated in the northwestern part of the Danube Lowland, in the Southwest Foreland of the West Carpathians (Minár et al., 2011). It belongs to the Trnavská pahorkatina Hilly Land - one of five loess hill-dominated lands within the region. The study area is characterized by relatively flat topography partially transformed by Quaternary tectonic uplift of micro-blocks up to 15 m above sunken micro-blocks. The study area is a zero-order catchment (0.28 km^2) within a dry valley situated on an uplifted block. It is delimited downstream by a road number 061017 (Trnava – Pusté Úľany), below which, the valley connects with Ronava creek (Fig. 1). The valley ridges are asymmetric in their altitude and slope with steeper southwest-orientated hillslopes (maximum slope of 8.4%), while the majority of slopes lie between 0.75 and 3.5% in the catchment. Total catchment altitude difference is 12.8 m and valley length is 0.99 km. The direction of the thalweg changes from (i) south-north (most upstream ~0.08 km) to (ii) southeast-northwest (0.08–0.985 km) and (iii) east-west at the outlet of the study area. Mean valley-bottom width is ~56 m (maximum of 97 m) and covers ~20% of the total catchment area. The valley bottom is characteristic by concave-concave morphometric forms and slopes with inclination 0–1.7%. On the hillslopes convex-convex morphometric forms are prevailing. A road gully cuts into the southwest ridge across the catchment boundary.

The dominant soil type in the study area, Calcic Chernozem, was developed in (up to) 6-m deep loess deposits of the Würm Age. Typical soil profiles are characterized by a 0.45-m thick mollic horizon (further referred as horizon “A”) overlying a 0.3-m thick transitional horizon (further referred as “AC”), which covers the original (loess) parent material (horizon “C”, Hraško, 1966). The depth of Chernozems before the onset of the anthropogenic influence (further referred as “original”, Ao) in the wider region remains unclear, despite recent progress in a study of Chernozem paleosols (Vysloužilová et al., 2015). Estimated thicknesses of the soil horizons range from 0.55 m to 0.65 m for Ao horizon, and 0.2 m to 0.3 m for the original AC horizon (Bedrna, Čurlík, Hraško, Juráni, 2013, personal communication).

The study area has humid-continental climate with warm to hot summers (Peel et al., 2007), mean annual temperature of 9.5 °C (January - 1,8 °C, July 19.7 °C), mean annual precipitation of 583 mm and precipitation deficit of ~150 mm. June has the highest precipitation (61–64 mm) with the highest intensity (23.94 ± 14.38 mm in 24 h). Highest rainfall intensity prevailed between May and August since 1921 (nearest meteorological station Trnava, < 15 km, Fig. 2). Maximum intensity of June–September rainfalls increased during second half of 20th century, while maximum intensity of October – May rainfall decreased. January or February are the driest months (33–38 mm), and snow falls between December and March provide snow cover > 0.01 m over a maximum of 45% of days in this period (Šamaj, 1991). Wind speeds > 4–6 $m \cdot s^{-1}$ prevail for north-west and south-east directions, and parallel with the direction of the longest thalweg section (0.08–0.985 km). Wind speeds > 4–6 $m \cdot s^{-1}$ initiate aeolian erosion on dry ploughed fields with poor vegetation cover and area > 0.01 km^2 (Streďanský, 1993), primarily between February and April, and in November.

The specific paleoclimate in Voderady is unknown, but was estimated using a macrophysical climate model for Hurbanovo (Jamrichová et al., 2014, Hajnalová, 2012), which is situated in the same climate region as Voderady. General trends from the model results are discussed below and can be found in Supplementary Fig. S1. Early Holocene climate (11,000–9500 cal. BP) was characterized by increasing mean annual temperatures (from 6 to 9.5 °C), decreasing mean annual precipitation (from 500 to 470 mm), which caused substitution of cold tolerant taxa (e.g., *Pinus sp.*) in loess Chernozem regions by temperate shrubs and trees (*Quercus*, *Ulmus*, *Tilia*, *Fraxinus*,and *Cornus sanguinea*), steppe species (*Artemisia*, *Jasione montana* t., *Plantago media/major*, and *Centaurea scabiosa* t.), and mesic open-habitats species (*Pimpinella major* t. and Filipendula). Until ~5800 cal. BP, increased temperature and decreased precipitation supported landscape opening and adoption of grazing and crop production in settled areas (Jamrichová et al., 2014). Around 5000 cal. BP, climate was more continental than today, with annual precipitation deficit ~200 mm and more extreme temperatures in January (−3 °C) and July (22 °C; Hajnalová, 2012). Subsequently, the trends changed to increased precipitation and winter temperature and decreased summer temperature (to ~560 mm, ~−1.5 °C, and ~20 °C in ~150–200CE, respectively). Further warmer oscillations in 500–600 CE (~10.4 °C, ~500 mm) and 1100–1250 CE (~10.4 °C, ~520 mm) were divided by colder and drier period in ~1000CE (~9.9 °C, ~490 mm). Onset of the Little Ice Age started after 1250 CE with temperature decreases (to ~9.5 °C in ~1630 CE), and oscillations in precipitation patterns (480–510 mm). Between 1630 and 1750 CE, both temperature and precipitation increased (to ~10.4 °C and ~580 mm), followed by declines and stabilization of recent climate patterns (Hajnalová, 2012).

The first settlement proximal (< 10 km) to the study area was documented in the Paleolithic era with continued settlement supported by archaeological findings (Smetanová, 2009b; Smetanová et al., 2012). Evidence for settlement includes the large Neolithic and Early Medieval village (< 1.5 km far), Bronze Age (< 1 km) and Roman Period settlement (outer catchment border), the Migration Period (375–568 CE) and Slavs colonization (after 470 CE, König, 2007). Nearby settlements indicate possible agricultural land use in some periods (Smetanová, 2011; Smetanová et al., 2012). Cadastral borders established in the 13th century remained without significant changes until 20th century, preserving the narrow long field structure typical for Medieval villages (Podolák, 2008). Land-use and landscape structure changes are documented with historical maps and imagery obtained between 1782 and 1784 and onwards (Fig. 3). During this period, the study area was used for growing annual crops, excluding the SW ridge, which was under viticulture until the mid-19th century. The three-field system, represented by rotation of spring and winter crops, followed by grassed fallow fields, was applied between the 11th and 18th century. In the 18th century, the continuous crop rotation without fallow fields was introduced. In the first half of the 20th century, changes in